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Volcanological evolution of Heimaey, South Iceland: From shallow-water to subaerial volcanism

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Volcanological evolution of Heimaey, South Iceland: from shallow-water to subaerial volcanism

Abstract

Heimaey is largest the island in the Vestmannaeyjar archipelago and also the centre of volcanic activity therein. Heimaey is the southernmost of nine volcanic centres in Iceland's Eastern Volcanic Zone (EVZ). The EVZ is propagating southwards in what probably is an attempt to adjoin with the submarine Reykjanes Ridge. The rift-tip is currently believed to be situated beneath Vestmannaeyjar. The evolution of Heimaey thus represents the earliest stage in the development of a central volcano in a propagating-rift environment.

The Heimaey volcanic centre (13.4 km^2) comprises ten individual eruptive units, ranging in age from early Holocene (~10 ka) to the most recent in 1973 (Eldfell). The island is composed of approximately equal amounts of tuff and lava, as most of the eruptions have started with a phreatomagmatic phase producing tuff cones and tuff rings.

The tuff cones have distal facies characterised by planar, normally graded, air-fall deposits occasionally disturbed by bomb sags. The proximal facies of the tuff cones are often exhibiting cross-bedding and other structures similar to high flow regime sedimentary structures, which are formed due to frequent emplacement of base-surges (gravitational collapses of wet eruption clouds) in the vicinity of the vent.

When the influence of the surrounding seawater diminished as the tuff rings rose above sea level, simultaneous phreatomagmatic activity and effusive lava flow emplacement were common inside the craters as shown by alternating lava flows and tuff deposits. Breccias with lava blocks, going from matrix-supported (tuff) to clast-supported with increasing stratigraphic height, are common inside craters prior to being capped by lava flows.

The final subaerial phase is characterised by Hawaiian and Strombolian activity, and lava flow emplacement. Lava flow morphologies range from clastogenic spatter-fed flows, tube-fed p-type (pipe-vesicle bearing) pahoehoe to blocky aa. Thin crater overflows present in two eruptive units indicate the presence of oscillating lava lakes inside the cones. Lava lakes, which formed as a result of lava flows ponding in topographic lows inside the craters of two tuff cones, have also been identified.

Index

1. Introduction

1.1 Volcanism in Iceland

Iceland is situated along the Mid-Atlantic Ridge (MAR) and superimposed on a mantle plume (Vink, 1984). The setting has resulted in anomalously high magma production rates, compared to the rest of the North Atlantic region. During the last 60 Ma Greenland, Eurasia and the NE Atlantic plate boundary have migrated north-westwards at a rate of 1 – 3 cm/year relative to the surface expression of the Iceland plume. The trail of the mantle plume through the history of the NE Atlantic can be seen as the Greenland-Faeroe Ridge (Fig. 1). The mantle plume is currently believed to be situated beneath the Vatnajökull glacier, approximately 200 km SE of the plate boundary (Gudmundsson, 2000). The Iceland plume has been modelled as a cylindrical zone of hotter material with low seismic (both P-wave and S-wave) velocities that extends to a depth of at least 400 km, but might in reality reach as deep as the core-mantle boundary at 2700 km (Gudmundsson, 2000). The current spreading rate for Iceland as a whole is 1.8 cm/year in the direction N105°E (DeMets et al., 1990). The oldest exposed rocks are therefore found in the areas furthest away from the active rift zones, as for example in Vestfirðir (Westfjords), North-Western Iceland. A very important feature in the geological evolution of Iceland is the periodical rift relocations (Johannesson, 1980; Helgasson, 1985; Hardarson et al., 1997). As the plate boundary drifts westwards away from the high magma production rates associated with the stationary Iceland plume, periodical relocations of main rift axis occur to keep the plate boundary centred above the plume.

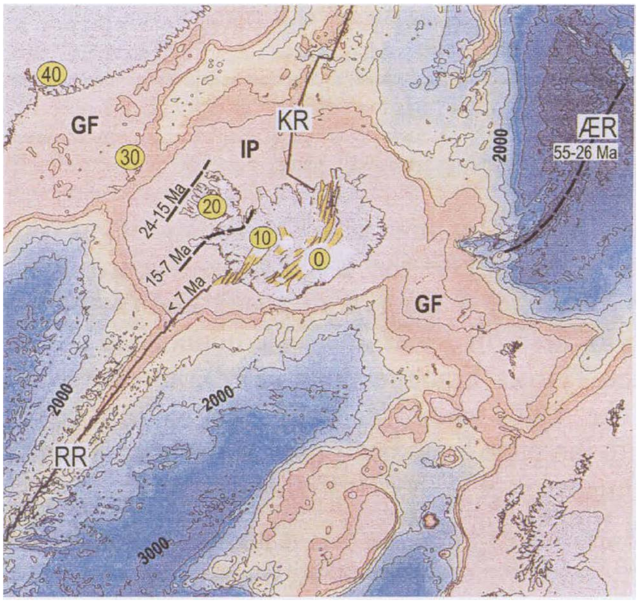

Fig. 1. Bathymetric map of the area surrounding Iceland. Depth contours for each 500 m. The position of the Iceland plume at 40, 30, 20, 10, and 0 Ma is marked by circles. GF = Greenland-Fareoe Ridge, RR = Reykjanes Ridge, KR = Kolbeinsey Ridge, IP = Iceland Plateu. Map base from Mclennan (2000).

1.2 Aim of study

The present study consists of a general introduction to the geology of Iceland combined with an overview of the characteristics of shallow water- vs. subaerial volcanism in general. Furthermore, two manuscripts are presented: (1) Geology of the Heimaey volcanic centre, South Iceland: early evolution of a central volcano in a propagating rift?, and (2) Reconstruction of the Helgafell eruption, Vestmannaeyjar, South Iceland. The first manuscript is only a draft and is still in preparation, whereas the latter manuscript has been submitted for publication in Bulletin of Volcanology and is currently under review. Both manuscripts are part of the results prepared while on a research fellowship at the Nordic Volcanological Institute, Reykjavík, Iceland.

The aim of the study is to map and characterise the volcanic geology of Heimaey as thoroughly as possible for a future comparison (i.e. both physical volcanology and petrological evolution) with other volcanic centres in the non-rifting segment of the Eastern Volcanic Zone.
Heimaey has a unique setting, as it is situated at the southernmost tip of the propagating rift, and thus represents the earliest evolution of a volcanic centre in such a setting. The thesis is mainly focused on the volcanological evolution of the Heimaey volcanic centre, but one additional manuscript is currently in preparation ("*Petrogenesis of alkali basalts in a propagating rift: evidence from the Heimaey volcanic centre, South Iceland*"), which will cover the petrological aspects in detail (major and trace elements combined with isotope analyses (U/Th, Sr, Nd, and Pb)).

2. Geology of Iceland

2.1 Rift zones and rift relocations

Iceland is at present day cut by three active spreading ridges (Fig. 2): (1) In the N by the Northern Volcanic Zone, (2) In the SW by the Western Volcanic Zone (sometimes referred to as the Reykjanes – Langjökull zone), and (3) In S by the Eastern Volcanic Zone. In addition to these rift zones, the plate boundaries in Iceland include a transform fault, known as the Tjörnes Fracture Zone (TFZ), and an overlapping spreading centre (SISZ; Fig. 2). The TFZ, which is partly exposed on land, connects the Northern Volcanic Zone with the plate boundary of the Kolbeinsey Ridge north of Iceland (Young et al., 1985; Gudmundsson et al., 1993; Rognvaldsson et al., 1998). The active rift zones are approximately 40 – 50 km wide and contain en-echelon arrays of fissure swarms. The fissure swarms can be up to 200 km in length.

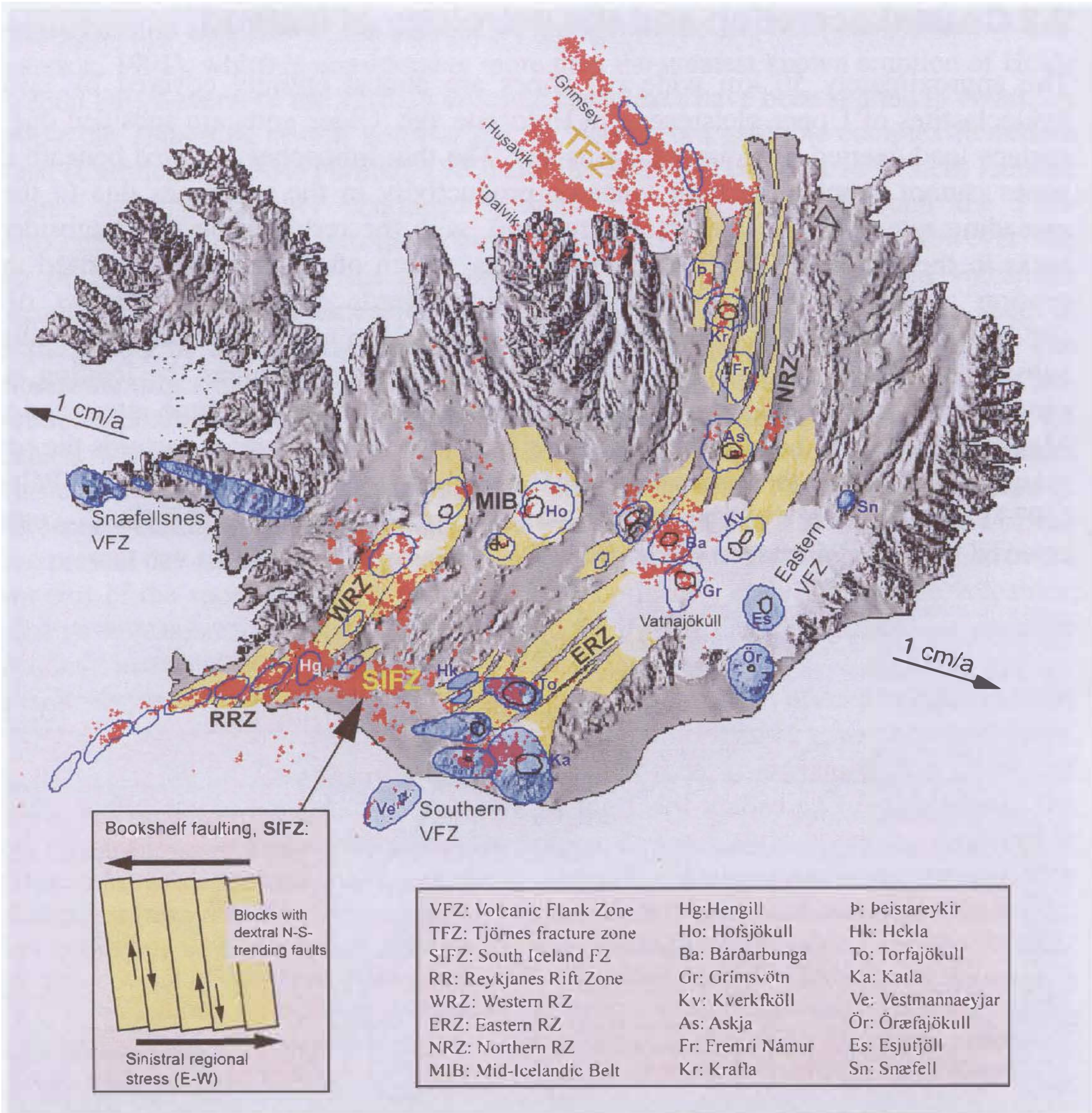

Fig. 2. Volcano-tectonic map of Iceland. Small red dots show the epicentres of the 25000 biggest earthquakes during the period 1994 – 2000. Fissure swarms are marked with yellow, volcanic centres and calderas with blue lines. From G. Guðmundsson, Icelandic Meteorological Office.

The Snæfellsnes rift zone in western Iceland (Fig. 1 & 2) became active approximately 15 Ma. At 5 – 6 Ma, the southern part of the Snæfellsnes rift became extinct and a new spreading axis was formed farther to the East (initiation of the present day WVZ). When the northern part of the Snæfellsnes rift became extinct (3 – 4 Ma), a predecessor to the northern part of the EVZ was formed. At approximately 2 Ma the volcanic activity spread southwards from the newly formed northern part of the EVZ, resulting in the present-day configuration of rift-axes in Iceland. The next step in the evolution of the rift zones might be the extinction of the WVZ and the formation of a fracture zone adjoining the EVZ with the submarine Reykjanes Ridge in the south (Johannesson, 1980).

2.2 Crustal accretion and the petrology of Iceland

The approximately 50 km wide rift zones are almost entirely covered by lavas and hyaloclastites of Upper-pleistocene to Holocene age. Older units are subsided due to the surface load exerted by younger extrusives. The thin lithosphere located beneath the rift zones cannot support the high volcanic productivity in the rift zones due to the slow spreading rate (1 cm/year in each direction), with the result being rapid subsidence of lavas in the rift zone. The Tertiary lava piles, which often have been exposed through erosion at the coast, commonly dip gently towards a current or extinct rift axis (Bödvarsson and Walker, 1964). The regional tilting is a result of continuous loading and subsidence of the crust in the vicinity of the rift zone. Whereas the loading is most pronounced under volcanic centres, which erupt frequently, the time-integrated (3 – 7 Ma) subsidence will be highest along the rift-zone axis and decrease towards the rift-zone margins. Based on these observations and other geophysical constraints, Pálmasson (1973) developed a dynamic model for the crustal accretion in Iceland, commonly referred to as the Pálmasson model (Fig. 3).

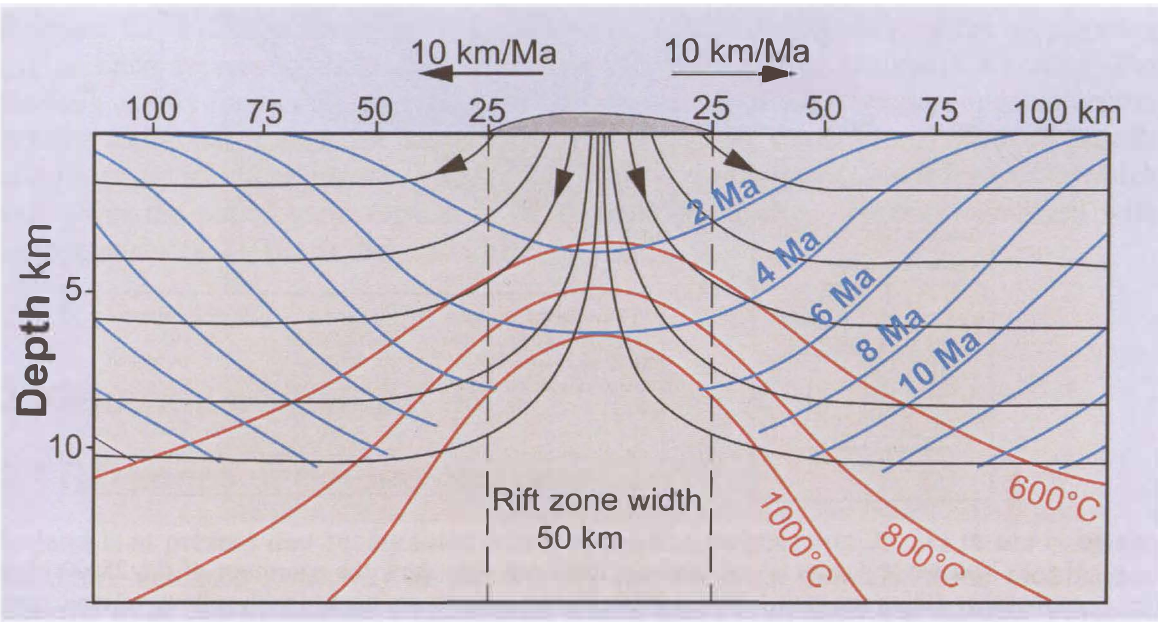

Fig. 3. Simplified model of Icelandic rift-zone dynamics (after Palmasson, 1973). The black lines are mass trajectories, blue lines are age contours (Ma), and the red lines represents temperature contours (°C).

About 80 – 85 % of the exposed volcanic pile in Iceland is composed of basalts, 10 % acidic and intermediate rocks, with the remaining 5 – 10 % being sediments/sedimentary rocks of volcanic origin (Saemundsson, 1979). Much of Iceland´s petrological diversity has been attributed to rift-propagation into older crust.

Tertiary (16 – 3.1 Ma) rocks are exposed in the eastern and western extremities of Iceland. Rocks exposed in Austfirðir (East-fjords) have been dated to 13.9 Ma, whereas in Vestfirðir (West-fjords) they date to at least 15.3 Ma (K – Ar; Hanan & Schilling, 1997). These rocks probably formed in an earlier major rift axis, but have been pushed apart as rifting progressed. Most of the Tertiary volcanics have been erupted as flood-basalts in fissure eruptions, and have therefore accumulated to substantial thickness.

There are, however, several remnants of central volcanoes that have evolved into silicic volcanism. In Fagridalur (eastern Iceland) an ignimbrite deposit is interlayered with Teritiary basaltic lava flows. The volume of the ignimbrite has been estimated to 4 km^3 (Einarsson, 1991), which is considerably more than the greatest known eruption of Hekla (H_3, 2800 BP). Several of the Tertiary volcanic complexes have been studied in detail, for example the Thingmúli central volcano and the Austurhorn gabbroic complex in eastern Iceland (Carmichael, 1964; Furman, 1989) and the Setberg I complex in western Iceland. Volcanic activity probably continued without any interruption through the Plio-Pleistocene (3.1 – 0.7 Ma) into the upper Pleistocene (Jakobsson, 1979). However, the onset of glaciation changed the environmental conditions drastically and volcanism occurred sporadically, as extensive lava flows in interglacial periods (as seen in Lyngdalsheiði) or as hyaloclastite ridges or table-mountains when ice-covered. The hyaloclastite ridges and table-mountains can be used to estimate the thickness of the ice-cover at the time of formation. The height of the hyaloclastite ridges gives a minimum ice thickness, whereas the table-mountains with their capping lava flows gives the maximum (as it needs to break through the ice-sheet in order to develop the capping lava flows). Pleistocene volcanics are today found mainly as a zone located at the margins of the active present day rift zones. They overly the Tertiary rocks, which are wide-spread in the lower part of the succession, as a capping of predominantly early Pleistocene volcanics. Explosive volcanism occurred at several locations during the Pleistocene, for example Torfajökull and Tindfjöll.

Volcanic activity was very high following the rapid deglaciation of Iceland. As much as 80 % of the Holocene volcanics were extruded over a time span of a few thousand years. The high volcanic productivity, during the early Holocene, is explained by a change of melting conditions in the mantle as a result of the rapid unloading (deglaciation). The rapid removal of an ice-sheet moves the solidus of the mantle upwards, resulting in melting of previously stable mantle. Approximately 30 volcanic systems have been active in Iceland during the Holocene with eruptions every 5 years on average (Einarsson, 1991). The volcanic activity is confined to the present day rift zones, the Snæfellsnes flank zone and the Öræfajökull volcano. The Holocene volcanic activity in Iceland is quite well known by use of tephrochronology. Tephrochronology is based on ^{14}C dating of ash layers in soil profiles and major element correlations of the composition of the glass shards (tephra) with the compositions of extrusives from known volcanoes. Since the settlement of Iceland (~ 870 AD) the source of several eruptions are also described in the annals.

The world's largest historical effusive eruption occurred at Laki in Iceland (1783-1784). The erupted lava flows cover nearly 600 km^2 and their volume has been estimated to approximately 15 km^3. About 10000 people (one fifth of Iceland´s population), and thousands of sheep, horses, and cattle died as a result of the eruption, primarily from starvation. However, the largest lava flow from the Holocene is the Þjórsá lava in southern Iceland (approx. 8500 BP). The Þjórsá lava flowed 120 km from the vent region and has a total volume of about 22 km^3.

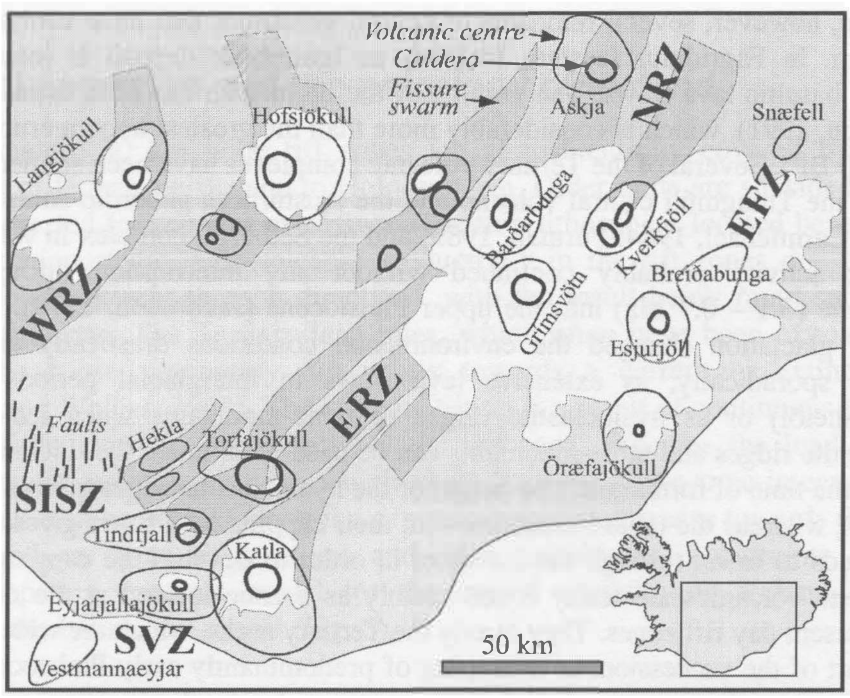

Fig. 4. Map of the Eastern Volcanic Zone (ERZ + SVZ), and the volcanic systems.

2.3 The Eastern Volcanic Zone (EVZ)

The EVZ can be divided into two separate segments (1) the central part from Bárðarbunga in the north to Torfajökull in the south (ERZ, Fig. 4), and (2) the southern flank zone covering the area south of Torfajökull to Vestmannaeyjar in the south (SVZ, Fig. 4). The central part of the EVZ contains abundant fissure swarms and active rifting takes place, whereas the southern flank zone lacks active rifting and the surface expression of fissure swarms (Fig. 4). The EVZ is linked to the WVZ via a leaky transform (South Iceland Seismic Zone), with Hekla situated at the triple-junction between the non-rifting and rifting segment of the EVZ and the SISZ. Occasional postglacial volcanism has occurred in the SISZ, for example in Hestfjall. The EVZ comprises nine volcanic systems: Bárðarbunga, Grímsvötn, Torfajökull, Hekla, Vatnafjöll, Tindfjöll, Eyjafjöll, Katla, and Vestmannaeyjar (Fig. 4). Five of the central volcanoes belonging to these systems have developed calderas. Jakobsson (1972) found a compositional trend in the lavas erupted along the EVZ, going from tholeiites in the north via FeTi-rich basalts in the central part, to alkali olivine basalts furthest south in the Vestmannaeyjar system. The spatial change in composition has been attributed to a southwards propagation of the rift-tip into older crust formed in the WVZ, in accordance with Pálmasson´s model (1973). Christe and Sinton (1981) also found a similar compositional pattern, going from tholeiites via FeTi-basalts to alkali basalts, in a propagating segment at the Galapagos spreading centre.

3. Shallow-water vs. subaerial volcanism

3.1 Shallow-water volcanism and phreatomagmatic activity

Phreatomagmatic activity involves the physical interaction between lava and water. A phreatomagmatic eruption is primarily driven by expansion of water that comes into contact with the lava and flashes to steam. The most recognized style of phreatomagmatic eruptions are explosive jets of ash (cock´s tail), often occurring simultaneously as steam plumes. This kind of activity was observed during the 1963 – 1967 Surtsey eruption in Vestmannaeyjar. The fragmentation dynamics in phreatomagmatic eruptions are controlled by a process commonly referred to as FCI (Fuel Coolant Interactions). The FCI is a heat transfer process that converts thermal energy from the lava (fuel), into kinetic energy on a small time scale. If FCI will be explosive depends on the rate of heat transfer between the fuel and the coolant (water), with the transfer rate being dependent on the surface area of the fuel (Morrisey et al., 2000). Fragmentation of the fuel increases the surface area, which causes further fragmentation. The deposits formed in phreatomagmatic eruptions are commonly tuffs consisting of ash to lapilli sized glass shards or clasts. Whole or broken pieces of co-magmatic phenocrysts are also common in the tuffs. Pillow lavas do not commonly occur in shallow water eruptions, as the confining pressure exerted by the overlying water column is not sufficient to suppress fragmentation. Base surges are produced by the gravitational collapse of a wet eruptive column and the deposits are commonly emplaced in the vicinity of the vent (Morrisey et al., 2000). Base-surge deposits are characterised by cross-bedding and dune structures, similar to high flow regime sedimentary structures.

3.2 Subaerial volcanism

As the tuff cones reach over the waterline, the effect of water on fragmentation will diminish or even stop. When FCI fragmentation seizes, the eruption will shift to a subaerial phase, which is characterized by effusive lava emplacement (pahoehoe or aa), fire-fountaining (continuous jetting of lava) or Strombolian activity (discrete explosions). The magma may still be fragmented, but due to bubble growth and rapid expansion of confined volatiles rather than to FCI (Wolff & Sumner, 2000). An eruption dominated by the outpouring of lava onto the ground is often referred to as an effusive eruption (as opposed to explosive eruptions caused by violent fragmentation of the magma). Lava flows generated by effusive eruptions vary in shape, thickness, length, and width. A common misconception is that the resulting lava morphology (aa or pahoehoe) is solely dependant upon the viscosity of the erupted lava. However, as a pahoehoe flow cools its viscosity increases through the same values as for aa-flows. The absurd consequence of this is that pahoehoe could never form (i.e. if viscosity was the only controlling factor of flow morphology, a a pahoehoe flow would always transform into an aa-flow during cooling). Peterson and Tilling (1980) found that the resulting lava morphology is dependent upon the viscosity of the lava *and* the rate of shear. However, Rowland and Walker (1990) found that the resulting lava morphology is also dependent on the volumetric flow rate.

3.3 Volcanic activity on Heimaey

Heimaey is situated on a relatively flat ocean shelf (~80 m depth dipping towards SE), approximately 10 km off the south coast of Iceland. The island is composed of roughly equal amounts of tuff/tuff breccia and lava flows. Pillow-lavas are absent, both in the field and in the Skipahellar drill-core, indicating that fragmentation of the lavas when jetted into the water could not be suppressed. Instead, tuff cones or tuff rings were formed.

The tuffs consist of brown sideromelane glass with minor amount of tachylitic glass and crystal fragments, mainly of olivine and plagioclase. All tuffs have been subject to variable degrees of post-depositional palagonitization, during which some elements were leached out of the sideromelane glass and replaced by H_2O. The tuff cones commonly exhibit cross-bedding and other structures similar to high flow regime sedimentary structures near the vent (proximal facies). These structures are believed to have been formed by base-surges. The outer slopes of the tuff cones are fine-grained, normally graded, planar, air-fall deposits containing sparse xenolithic ejecta (distal facies), except in the Sæfell tuff ring, where xenoliths and bomb sags are common features in the distal air-fall deposits. The slopes of the tuff cones/rings have variable dips ranging between 25° and 45° into the crater, and 5° to 20° out from the crater.
When the influence of the surrounding seawater diminished as the tuff rings rose above the sea-level, simultaneous phreatomagmatic activity and effusive lava flow emplacement were common inside the crater as reflected by interlayered lava flows and tuff deposits. Breccias with lava blocks, going from matrix-supported (tuff) to clast-supported with increasing stratigraphic height, are common inside craters prior to being capped by lava flows. A variety of volcanic activities characterise the subaerial stage (i.e. Hawaiian, Strombolian, and to some extent Surtseyan while sea-water is still able to flow into the vent), as reflected by the resulting deposits. Hawaiian and Strombolian activity are known from the Eldfell and Surtsey eruptions, and can be inferred at least at the Helgafell and Háin vents as these also comprises scoria cones. The 132 cm thick T2-tephra horizon of the Helgafell eruption is also consistent with fall-out deposits formed during Hawaiian or Strombolian activity.

Both spatter-fed lava flows (high eruption rate) and tube-fed flows (lower eruption rates) are present on Heimaey. Spatter-fed flows are only identified in the Helgafell unit, where they formed during an early phase of the eruption (i.e. high eruption rate) when large amounts of spatter accumulated near the vent, producing rheomorphic flows. The Helgafell spatter-fed flows are characterised by their thinness (0.5 – 2 m), large lateral extent (<200 m) and relatively high vesicularity. The spatter-fed flows have incorporated numerous volcanic bombs. The bombs are rotated in the lava flows, and range in size from 5 to 35 cm in diameter, with a 5 - 15 mm vesicular rim at the contact with the surrounding lava. All examined bombs show the same plagioclase-phyric appearance as the host lava, and are therefore believed to be co-magmatic.

As the eruption rate was lowered, lava started to flow in open channels close to the vent. While flowing in channels the lava cools and a crust begins to form and at a certain distance from the vent (depending on temperature, flow rate, etc.), the channels roof themselves over and form tubes. Transport through a system of tubes inside a lava field is a much more efficient way to deliver lava to the active flow-front than transport in open channels.

P-type (pipe-vesicle bearing) pahoehoe are typically formed in tubes. The overall characteristics of a P-type pahoehoe flow are a thin lower vesicular zone (~ 1 dm) at the base of the flow, a dense interior with few vesicles and an upper vesicular zone. In Helgafell, pipe vesicles up to 1 m in length are common in the massive interior of the flow lobes. The tube-fed pahoehoe flows are much less vesicular than the spatter-fed flows, as they are allowed longer time to degas during transport. Thin crater-overflows occur near the top of the Helgafell scoria cone and are also found in Stórhöfði. The flows are extremely thin (1 to 3 dm) and formed while oscillating lava lakes were present in the craters. In two cases (Klífið and Dalfjallshryggur), lava has ponded in topographic lows inside a pre-existing tuff cone and formed lava lakes. The lava lakes range between 20 and 40 m in thickness. Due to the slow-cooling processes in the lava lakes, segregation veins often developed.

4. Future work

I will continue my fellowship at the Nordic Volcanological Institute for the period 2002/2003, focusing mainly on the Tindfjöll central volcano, southern Iceland. Tindfjöll is supposedly one of the oldest volcanoes in the EVZ, erupting compositions ranging from basalt to rhyolite. The Tindfjöll volcano has developed a caldera and ignimbrite deposits (the Thórsmörk ignimbrite; Jörgensen 1980) that have been suggested to be the source for the Ash Zone 2 in the North Atlantic (Lacasse and Garbe-Schoenberg, 2001). Glass shards from Tindfjöll have also been identified in ice-cores from the Greenland ice-sheet.

The present thesis will be used as a part of a PhD-project on the evolution of central volcanoes in a propagating rift environment. The aim is to study the structural as well as the petrological evolution of volcanic centres as the rift propagates through the crustal plate and, if possible, determine why the most evolved lavas occur at a given distance from the rift tip. Hopefully, a detailed comparison of the two volcanic centres (Heimaey which is the youngest and Tindfjöll which is one of the oldest) will explain the evolution from alkali basalts erupted at the tip of the propagating rift, to FeTi-rich basalts and rhyolites behind the propagator. The PhD-project will continue with three major projects:

1. *Petrogenesis of alkali basalts in a propagating rift: evidence from the Heimaey volcanic centre, South Iceland.* The project is a detailed petrochemical study of the Heimaey extrusives, using major and trace element data (ICP-MS & ICP-AES) as well as mineral chemistry and isotopes (i.e. Sr-Nd-Hf-Pb isotope ratios) in order to establish their petrogenesis. In collaboration with Níels Óskarsson (NORDVULK).

2. *Geology of the Tindfjöll central volcano, South Iceland.* As the Tindfjöll volcanic centre is only sparsely described in the geological literature, a new description and re-

inetrpretation is needed. Mapping of the volcano is continued from previous research fellows work at NORDVULK. The project also include major- and selected trace element analyses (ICP-AES) of the Tindfjöll rock suite, in order to characterise the volcano. In collaboration with with Karl Grönvold, Níels Óskarsson and Rósa Ólafsdóttir (NORDVULK).

3. *Petrogenesis of the Tindfjöll eruptives*. This project is like the Heimaey equivalent a study of the petrological evolution with time of the volcano. The study will be based on the use of major and trace element data (ICP-MS & ICP-AES) as well as mineral chemistry and Sr-Nd-Hf-Pb isotope ratios. A first look at the Tindfjöll data indicates that the volcano is alkaline, with a bimodal distribution of lava compositions (alkali- to FeTi-rich basalts and rhyolites). In collaboration with with Karl Grönvold, Níels Óskarsson and Rósa Ólafsdóttir (NORDVULK).

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Geology of the Heimaey volcanic centre, south Iceland: early evolution of a central volcano in a propagating rift?

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Abstract

Heimaey is the southernmost and also the youngest of nine volcanic centres in the southward-propagating Eastern Volcanic Zone, Iceland. The island of Heimaey belongs to the Vestmannaeyjar volcanic system (850 km^2) and is situated 10 km off the south coast of Iceland. Although Heimaey probably started to form during the Upper Pleistocene all the exposed subaerial volcanics (10 monogenetic vents covering an area of 13.4 km^2) are of Holocene age. Heimaey is composed of roughly equal amounts of tuff/tuff-breccias and lavas as most eruptions involve both a phreatomagmatic and an effusive phase. The compositions of the extrusives are predominantly alkali basalts belonging to the sodic series. Repeated eruptions on Heimaey, and the occurrence of slightly more evolved rocks (i.e. hawaiite approaching mugearite), might indicate that the island is in an early stage of forming a central volcano in the Vestmannaeyjar system. This is further substantiated by the development of a magma chamber at 10–20 km depth during the most recent eruption in 1973 and by the fact that the average volume of material produced in a single eruption on Heimaey is 0.32 km^3 (dense rock equivalent), which is twice the value reported for the Vestmannaeyjar system as a whole. We find no support for the previously postulated episodic behaviour of the volcanism in the Vestmannaeyjar system. However, the oldest units exposed above sea level, i.e. the Norðurklettar ridge, probably formed over a 500-year interval during the deglaciation of southern Iceland. The absence of equilibrium phenocryst assemblages in the Heimaey lavas suggests that magma rose quickly from depth, without long-time ponding in shallow-seated crustal magma chambers. Eruptions on Heimaey have occurred along two main lineaments (N45°E and N65°E), which indicate that it is seismic events associated with the southward propagation of the Eastern Volcanic Zone that open pathways for the magma to reach the surface. Continuing southward propagation of the Eastern Volcanic Zone suggests that the frequency of volcanic eruptions in the Vestmannaeyjar system might increase with time, and that Heimaey may develop into a central volcano like the mature volcanic centres situated on the Icelandic mainland.

Keywords: Iceland; Vestmannaeyjar; propagating rift; volcanism; eruption

1. Introduction

Iceland is situated at the confluence of the Mid-Atlantic Ridge and a mantle plume (Vink, 1984).

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The plate boundary moves WNW relative to the stationary mantle plume (DeMets et al., 1990), causing periodic relocations of the main rift axis (Saemundsson, 1974, 1986; Johannesson, 1980; Hardarson et al., 1997). The most recent relocation started approximately 3 Ma ago (Saemundsson, 1979; Johannesson, 1980; Einarsson, 1991), with the initiation of the Eastern Volcanic Zone (EVZ). The EVZ propagates southward, in what probably is a future attempt to adjoin with the submarine Reykjanes Ridge. The tip of the rift has been suggested to be located beneath the Vestmannaeyjar volcanic system (Oskarsson et al., 1982).

The EVZ comprises nine volcanic systems (Jóhannesson and Saemundsson, 1998) from Bárðarbunga in the north to Vestmannaeyjar in the south (Fig. 1), in which five of the central volcanoes have developed calderas, i.e. Eyjafjöll, Grímsvötn, Katla, Tindfjöll and Torfajökull. The term central volcano is used, following Gudmundsson (1995), to denote volcanoes that: (1) erupt frequently; (2) extrude basaltic, intermediate and acidic products; (3) are commonly associated with, and fed by, shallow crustal magma chambers; and (4) are the largest volcanic structures in Iceland, often associated with collapse calderas.

A compositional change of the eruptive products along the EVZ is evident (Jakobsson, 1972; Meyer et al., 1985). The northern part of the EVZ, representing a mature rift zone, is dominated by tholeiites. The central EVZ is characterised by FeTi-rich basalts, whereas furthest south, at the rift tip in Vestmannaeyjar, alkali basalts dominate. A similar compositional pattern has been described from a propagating segment at the Galapagos spreading centre (Christie and Sinton, 1981).

Vestmannaeyjar is best described as a monogenetic volcano field (Walker, 2000), comprising tuff rings, small lava shields and scoria cones with associated lava flows. Heimaey is the largest of 17 islands in the Vestmannaeyjar archipelago and also represents the centre of volcanism in the system. Repeated eruptions and the occurrence of more evolved lavas on Heimaey compared to the rest of the system, i.e. hawaiite approaching mugearite, have been used to suggest that Heimaey might be at an early stage of developing into a central volcano (Jakobsson, 1979a). A geological map of Heimaey has previously been pub-

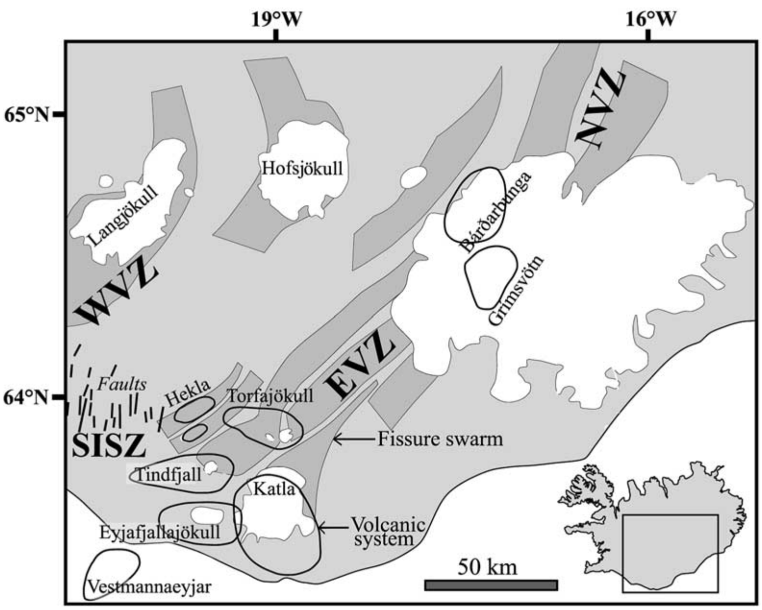

Fig. 1. Map showing the nine volcanic systems of the EVZ and the location of Vestmannaeyjar. NVZ = Northern Volcanic Zone, WVZ = Western Volcanic Zone, SISZ = South Iceland Seismic Zone.

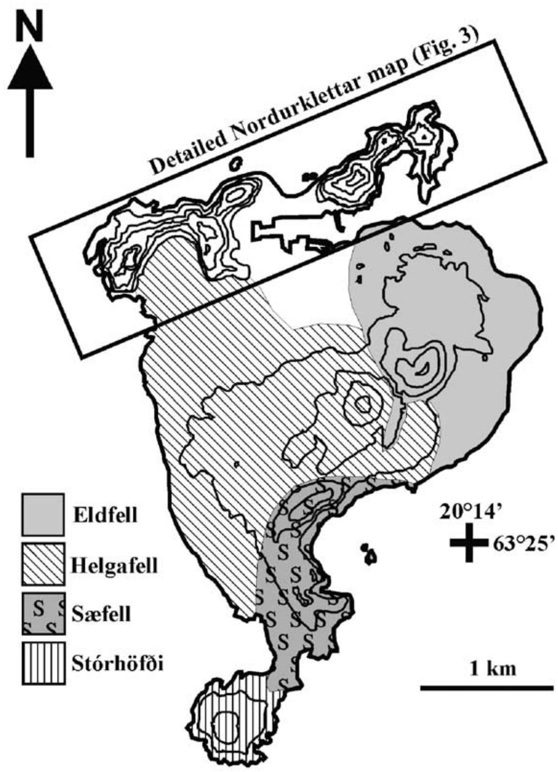

Fig. 2. Geological map of Heimaey. For a detailed map of the Norðurklettar formation see Fig. 3.

lished (Jakobsson, 1968). In the map, however, no distinction was made between the six different eruptive units belonging to the Norðurklettar formation. In this paper we present a summary of previously published papers on Heimaey as well as new detailed mapping of the island (Figs. 2 and 3) and the relative timing for some of the eruptions. New major element analyses have been made for the majority of the eruptive units, as well as a re-estimation of the erupted volumes in order to provide a general geological overview of Heimaey to facilitate a forthcoming petrogenetic study of the volcanic centre.

2. Geological setting

The Vestmannaeyjar volcanic system covers approximately 850 km^2 off the south coast of Iceland. There are 17 islands/skerries in the archipelago with 22 known Holocene subaerial eruption sites (Jakobsson, 1979a), of which 10 are located on Heimaey (Fig. 4). Eruptions in the Vestmannaeyjar system are generally of small volume (average 0.17 km^3; Jakobsson, 1979a), which originate from monogenetic vents.

Heimaey is composed of roughly equal amounts of tuff or tuff-breccia and lava, as most eruptions have started with a phreatomagmatic phase that shifted to effusive as the vents isolated themselves from the influence of the surrounding seawater. The Norðurklettar formation rises as a steep-sided ridge, reaching a maximum of 280 m above sea level (a.s.l.), on the northern side of the island. The relatively high altitude of tuff–lava transition in Norðurklettar (220–160 m a.s.l.)

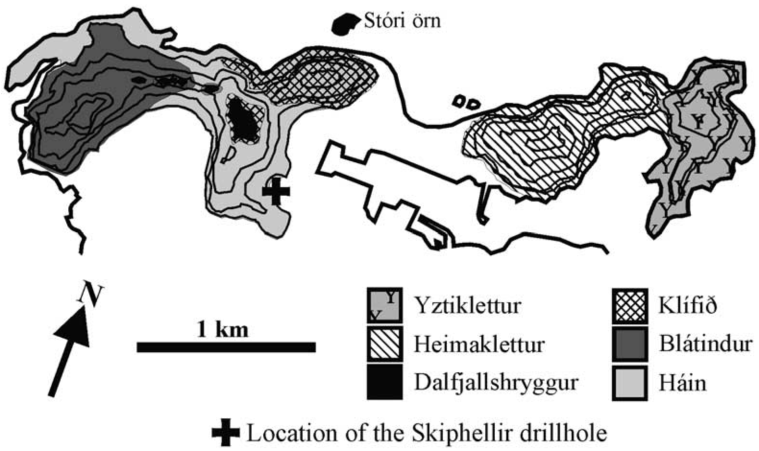

Fig. 3. Geological map showing the six eruptive units comprising the Norðurklettar formation, Heimaey.

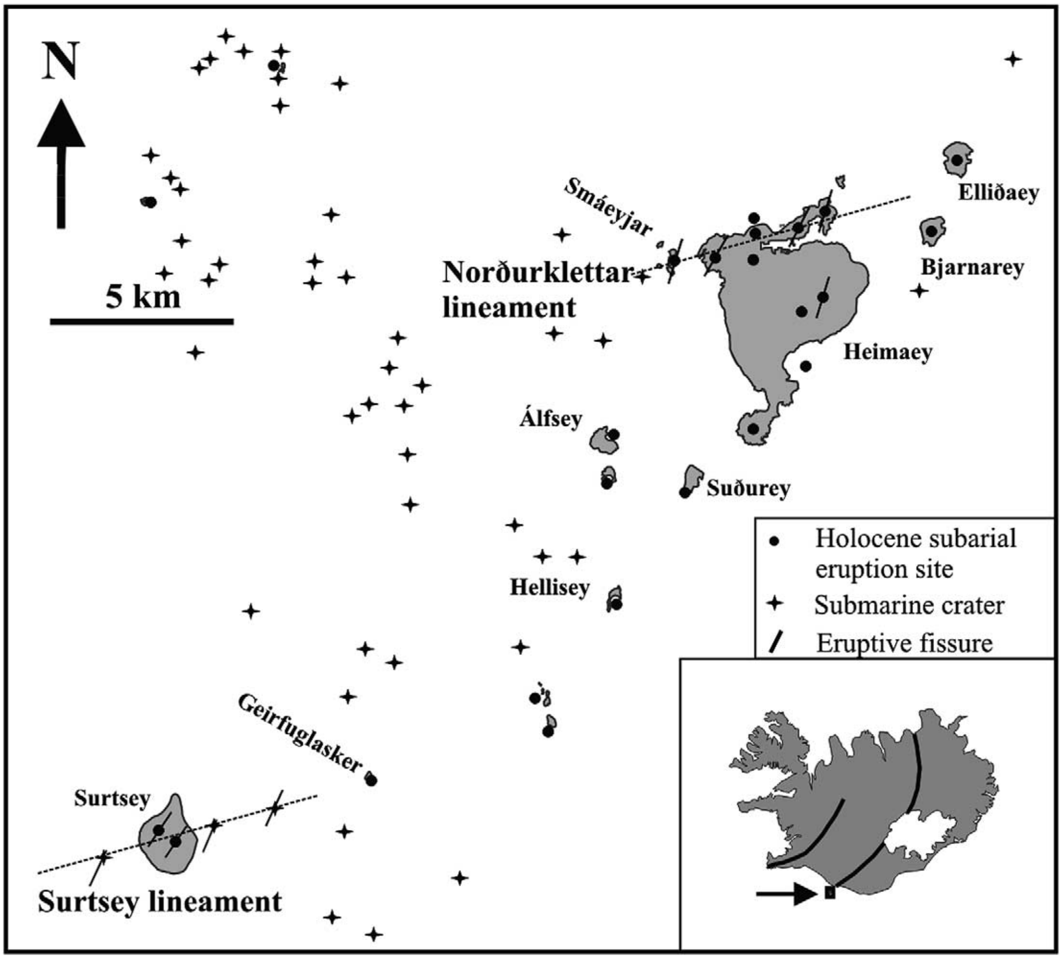

Fig. 4. Schematic map showing the main tectonic features of the Vestmannaeyjar volcanic system, the locations of subaerial Holocene eruption sites (black dots), and some of the submarine craters belonging to the system (star-shaped). Note the similar arrangement of eruptive fissures in the Surtsey and Norðurklettar lineaments.

has been used to suggest that the eruptions occurred when the sea level relative to Heimaey was higher than at present. This could occur in the final stages of glaciation when Iceland was depressed through ice loading (Jakobsson, 1968, 1979a). Two scoria cones, Helgafell and Eldfell, dominate the topography of Heimaey and their associated lava fields cover approximately 80% of the island.

Marine abrasion is constantly reshaping the island, forming coastal cliffs that range in height from a few metres to up to 200 m in the oldest part of the island. The cliffs provide excellent stratigraphic exposures and an insight into how the island formed. Previous work on Heimaey has mainly been focused on the 1973 Eldfell eruption (Jakobsson et al., 1973; O'Nions et al., 1973; Self et al., 1974; Furman et al., 1991), and to a lesser extent on the other formations (Jakobsson, 1968, 1979a; Leys, 1983; Mattsson and Höskuldsson, submitted). In 1964 a 1500 m deep hole was drilled near the harbour on Heimaey (i.e. the Skiphellir drill core; Fig. 3; Palmason et al., 1965). The results of the drilling showed that basalts and tuffs dominated down to 180 m below sea level (Tomason, 1967), which is believed to be the lowermost limit of the Heimaey extrusives probably formed during the Upper Pleistocene (Jakobsson, 1979a). From 180 to 740 m, marine tuffaceous sediments dominate, followed by altered FeTi-rich basalts below 740 m.

3. Tectonic structures

Many of the larger islands belonging to the Vestmannaeyjar volcanic system are arranged along the same trend as the EVZ (N45°E), including four of the eruptive units on Heimaey. Jakobsson (1968) noted the similar en échelon arrangement of eruptive fissures in the Norðurklettar formation on Heimaey and the eruptive fissures of Surtsey (Fig. 4). Eruptions on Heimaey occur along two main lineaments (Fig. 4): (1) Stórhöfði–Eldfell, which trends N45°E, or (2) Norðurklettar, which trends N65°E. The Surtsey eruption (1963–1967) erupted magma along a lineament with a general trend of N65°E, consisting of five eruptive fissures oriented approximately N35°E (Thorarinsson, 1965; Jakobsson, 1968). Brander and Wadge (1973) performed distance measurements across the Eldfell eruptive fissure and found that the plane of maximum compression on Heimaey, i.e. the bisecter of the angle between the left- and right-lateral components, is almost identical to the main trend of the EVZ. The authors therefore proposed a change in stress regime at the tip of the propagating rift, e.g. compressional force exceeds extensional, resulting in fissures formed due to shear. Similar conjugate sets with faults as seen on Heimaey are also found as the axes of the Hekla, Tindfjöll and Eyjafjallajökull volcanoes, situated in the southern part of the EVZ on the Icelandic mainland (Saemundsson, 1979). Saemundsson (1979) therefore suggested that the southernmost part of the EVZ may be the result of sinistral shear movement with the axis of maximum compression oriented N45°E and the axis of maximum tension subparallel to the spreading direction (NW–SE). Clay modelling by Clifton (2000) showed that a spreading ridge that is subject to an oblique extensional stress, such as the EVZ, produces an en échelon arrangement of fissures. The change in stress regime at the propagating tip suggested by Brander and Wadge (1973) is therefore not required in order to produce the surface features observed in Vestmannaeyjar.

4. Geomorphology of Heimaey

Because most eruptions on Heimaey started below sea level, the initial volcanic activity was phreatomagmatic with formation of tuff cones or tuff rings. The eruptions shifted to effusive or Strombolian as the influence of the surrounding seawater diminished. Exceptions are the three most recent eruptions, i.e. Helgafell and Eldfell, which were mostly subaerial, although their eruptive fissures interacted to a minor extent with seawater, and the Sæfell eruption which was mostly phreatomagmatic, but also included a minor lava flow originating from a parasitic vent at the flank of the tuff ring (Leys, 1983). Fig. 5 shows a cross-section through the Háin eruptive unit, which can be used to characterise most eruptions on Heimaey as it involves a phreatomagmatic tuff ring building phase followed by Strombolian/effusive activity which is characterised by scoria cone formation and lava emplacement. The following description of the Heimaey eruptive units is therefore divided into three categories: (1) tuff facies; (2) scoria facies; and (3) lava facies.

4.1. Tuff facies

The tuff facies ranges from fine-grained, normally graded, planar, air-fall deposits with sparse xenoliths (Fig. 6a) to cross-bedded deposits, formed by base surges with similar appearance as high-flow-regime sedimentary structures (Fig. 6b). The slopes of the tuff cones have variable dips ranging between 25 and 45° into the crater and between 5 and 20° out from the crater. Unconformities due to slumping mark all the transitions from outward sloping tuff, to tuff dipping towards the crater. All tuff deposits have undergone variable degrees of post-depositional palagonitisation.

4.1.1. Norðurklettar

Háin is one of the island's most prominent tuff rings, reaching over 200 m in height. The tuff ring

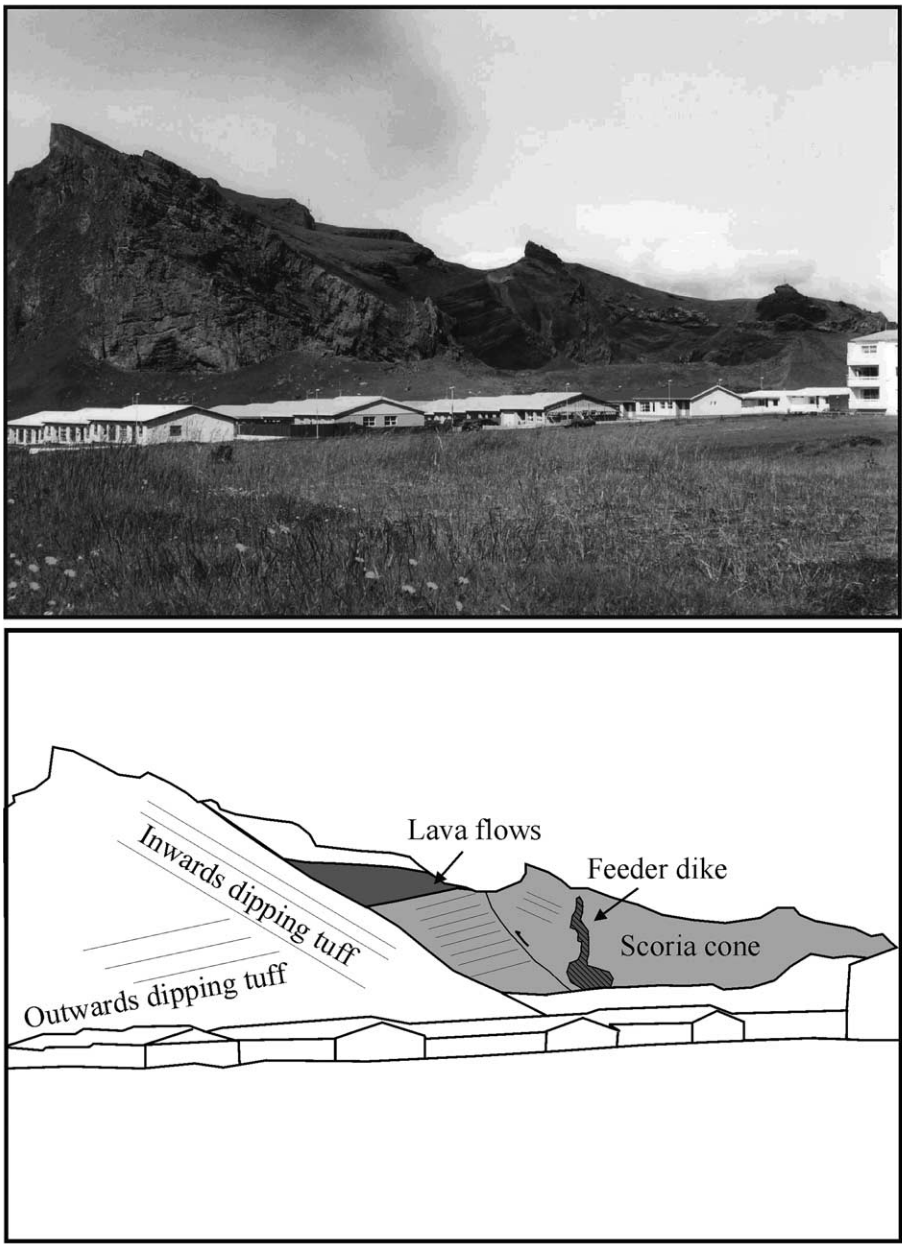

Fig. 5. Cross-section of the Háin eruptive unit. A scoria cone (light grey) built up inside the crater of the larger tuff ring. Lava flows have ponded in topographic lows inside the crater of the tuff ring.

has been subject to considerable erosion since the time of deposition. The basal diameter can be estimated to be approximately 3 km, as the distal part of the tuff ring outcrops beneath Blátindur (1.5 km from the inferred vent location). The tuff is characterised by abundant rounded lapilli-sized clasts with glassy rims in a matrix of ash (Fig. 6c). The sparse xenoliths found in the Háin tuff are commonly confined to the immediate vicinity of the crater, and are well-rounded igneous or angular sedimentary blocks originating from the seafloor (Fig. 6b). The tuff dipping into the crater is

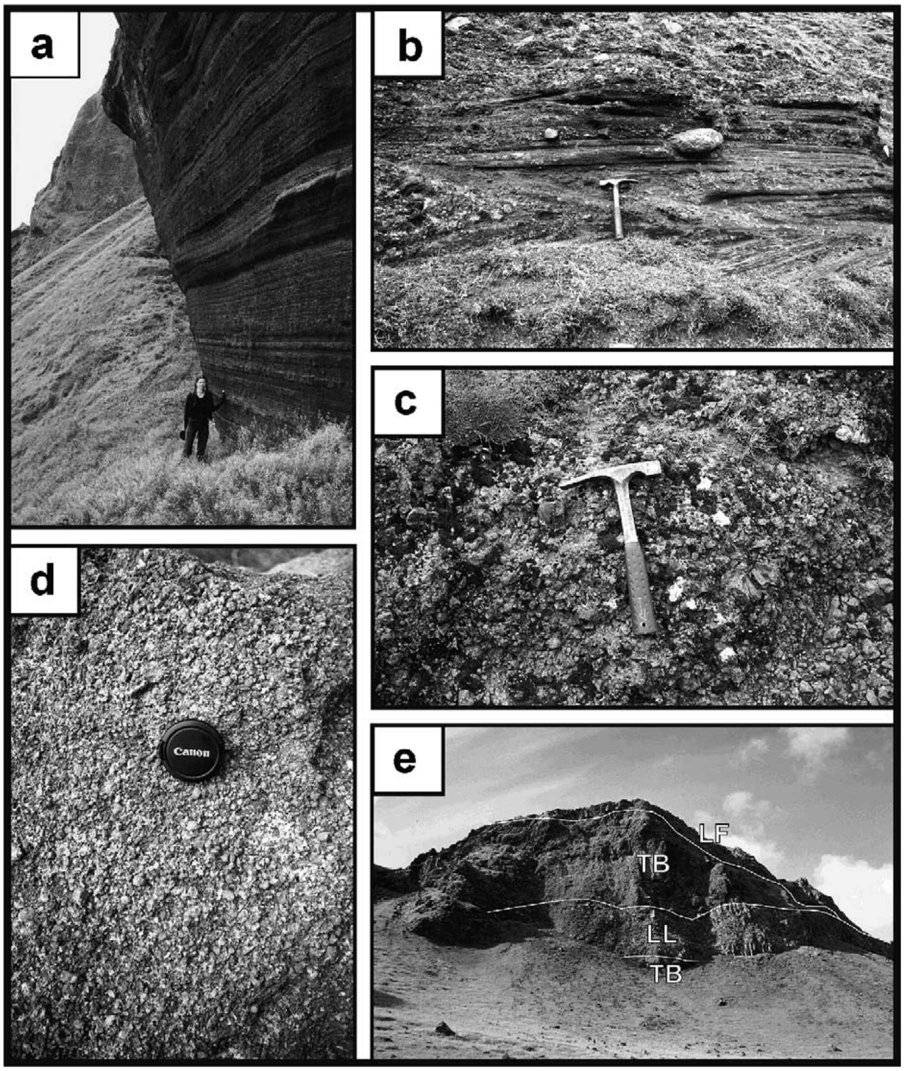

Fig. 6. Characteristic features observed on Heimaey. (a) Planar air-fall tuff belonging to the Blátindur unit. (b) Cross-bedded tuff in Háin containing a well-rounded xenolithic block originating from the seafloor. (c) Lapilli-sized tuff from Háin. (d) Tuff-breccia in Blátindur. (e) The Klífið unit. TB = tuff-breccia (similar to panel d), LL = lava lake, LF = subaerial lava flows.

in many locations cross-bedded (Fig. 6b), indicating that base surges were common inside the crater. Faulting and slumping at a later stage reworked the tuff.

4.1.2. *Sæfell*

The Sæfell tuff ring (188 m a.s.l.) had a crater diameter that probably exceeded 1 km, of which only half remains today due to extensive marine abrasion. Because half of the crater rim remains, and the outer slope of the tuff cone is exposed in the coastal sections beneath the Helgafell lava, the basal diameter of the tuff ring at the time of formation can be estimated to be at least 3 km. The Sæfell eruption was mostly phreatomagmatic, as the eastern side of the tuff ring probably remained breached throughout the eruption and allowed seawater access to the vent area (Leys, 1983; Hand, 2001). Sæfell exhibits many features consistent with simultaneously formed base-surge and air-fall deposits, i.e. large antidunes, ripples, and cross-bedding are common features in the proximal deposits whereas fine-grained normally bedded air-fall deposits dominate the distal parts. Indications of post-depositional reworking by processes such as slumping, debris flows and wind action are found at several locations in the tuff ring (cf. Leys, 1983). The Sæfell tuff ring contains a large amount of xenolithic ejecta. Most of the xenoliths are accidental lithic fragments, derived from the underlying basement or seafloor together with gabbroic nodules of deeper origin. The abundant xenoliths of basement rocks in the tuff ring, e.g. FeTi-rich basalts, suggest that explosions frequently occurred at depths exceeding 740 m as indicated by the Skiphellir drill-core stratigraphy (Palmason et al., 1965; Tomason, 1967).

4.2. Scoria facies

Hawaiian and Strombolian eruptions are the least violent form of volcanism, characterised by sustained jetting of magma into the air (Hawaiian) or discrete explosions (Strombolian). Hawaiian and Strombolian activity are generally characteristic for magmas of basaltic or basalt-andesitic compositions. The coarseness of the ejecta and the low eruption column associated with Hawaiian and Strombolian eruptions prohibits a widespread dispersal of the ejecta, and they commonly form spatter or scoria cones.

At least three scoria cones have been built up on Heimaey, although there have probably been several others in the Norðurklettar formation which have since been eroded away. The Helgafell and Eldfell scoria cones (Fig. 2) are situated in the vicinity of the town of Vestmannaeyjar and dominate the topography of the island (excluding the Norðurklettar ridge).

4.2.1. Háin

The final phase of the Háin eruption was effusive and produced a small scoria cone with associated minor lava flows inside the crater of the tuff ring. Only a minor part (approx. 15%) of the Háin scoria cone remains today, due to extensive erosion (Fig. 5).

4.2.2. Helgafell

The Helgafell cone (228 m a.s.l.), which formed about 5900 BP, is a classical example of a Strombolian scoria cone. The cone is 120 m high with a basal diameter of approximately 500 m, and thus has a slightly higher aspect ratio (height/width) (0.24) than the Eldfell cone. The Helgafell scoria cone, however, has a lower slope (22°) than the Eldfell cone, which is consistent with degradation of an older cone (Hooper and Sheridan, 1998). Three lava channels run towards north and west on the lower slopes of the cone.

4.2.3. Eldfell

The Eldfell eruption started on January 23 1973, issuing lava from a 1.6-km-long NE-trending fissure characterised by Hawaiian fire-fountaining activity. However, the activity quickly became localised to three vents, and by February 6 all activity was concentrated to a single vent at which the Eldfell scoria cone was built. Approximately one month after the onset of the eruption, the northeastern wall of the cone collapsed and was carried away as fragments (up to 1.8×10^6 m^3) by the lava flows as far as 1.5 km N of the vent. The cone (223 m a.s.l.) rises 130 m above the surroundings and has a basal radius of 300 m. The aspect ratio of the Eldfell cone is 0.22, with an average slope of 31°.

4.3. Lava facies

Lava flows on Heimaey, like most flows in the Vestmannaeyjar system, have the surface morphology of rough pahoehoe. Other lava morphologies visible on Heimaey include spatter-fed clastogenic flows, thin crater overflows, tube-fed P-type (pipe vesicle-bearing) pahoehoe lava as well as a′a lava. The thickness of individual lava flows varies from 0.1 to 10 m. The majority of the lava flows are either: (1) emplaced and occasionally ponded in topographic lows inside the crater of the tuff cone, as observed in the Klífið and Háin units (Fig. 5), or (2) crater overflows as observed in the Stórhöfði unit. Lava flows cap all of the tuff cones, with the exception of Sæfell. Lava lakes are present in the Dalfjallshryggur and Klífið formations.

4.3.1. Klífið

Interlayered with the Klífið tuff-breccia is a lava lake (Fig. 6e). The lake is 20–30 m thick and has solidified slowly (as it is the only unit on Heimaey that has developed equilibrium phenocryst assemblages containing plagioclase, olivine, clinopyroxene and titanomagnetite). Coarse-grained segregation veins, up to 20 cm in width, are also common in the lava lake.

4.3.2. Dalfjallshryggur

The Dalfjallshryggur lava is homogeneous, 30–40 m thick and exhibits poorly developed columnar jointing. There are also some minor ($\sim$20 000 m^3) outcrops of the Dalfjallshryggur lava along the Dalfjallshryggur ridge (Fig. 3). The extreme thickness of the unit, compared to others on Heimaey, and the location in a former topographic low suggest that the unit is the remnant of a ponded lava lake. The location of the eruptive vent is uncertain, however, one possible candidate is Stóri örn (a columnar jointed feeder dike), sticking up from the ocean 700 m north of Dalfjallshryggur (Fig. 3).

4.3.3. Stórhöfði

The Stórhöfði lava flows are characterised by numerous thin flows (10–30 cm each), emplaced in various directions from the crater and gently capping the tuff cone. There are, however, several thicker major flows that have been emplaced towards the north. These flows are presently under the sea surface north of the vent (Fig. 2). The many thin flows probably originated as crater overflows and owe their extreme thinness to the primitive composition/low viscosity of the lava.

4.3.4. Helgafell

The lava flows of Helgafell are predominantly tube-fed pahoehoe characterised by abundant inflation features such as tumuli and pressure ridges, but also include spatter-fed clastogenic flows, emplaced in the beginning of the eruption when the eruption rate was high (Mattsson et al., 2002; Mattsson and Höskuldsson, submitted). Two small crater overflows are present near the top, on the northern and eastern side of the cone.

4.3.5. Eldfell

The Eldfell lava flows differ substantially from other flows on the island as they are predominantly of blocky a′a type. The Eldfell lava has the most evolved composition (hawaiitic that is approaching mugearitic; Jakobsson et al., 1973) of the products erupted in the Vestmannaeyjar system, which may be the determining factor for flow morphology. The Eldfell lava initially flowed to the east of the crater. The collapse of the northeastern wall of the Eldfell cone during the eruption caused lava to flow in the direction of the town, burying houses on the eastern side of the town of Vestmannaeyjar. Today, after three decades of marine abrasion several lava tubes have been exposed in the coastal cliffs of the Eldfell lava. None of the tubes has drained to form tunnels, but solidified in situ, forming concentric vesicle-rich bands.

5. Stratigraphy

The Skiphellir drill core stratigraphy can be divided into three major units: (1) the Vestmannaeyjar formation; (2) marine tuffaceous sediments; and (3) tholeiitic and FeTi-rich basement rocks, often containing abundant pyroxene phenocrysts. The two lowermost units belonging to the Vestmannaeyjar formation are separated by a tillite layer at approximately 120 m beneath sea level (Palmason et al., 1965; Tomason, 1967), indicating that Heimaey started to form during the Upper Pleistocene. The stratigraphic relations of the different units exposed above sea level are rather straightforward, with the exception of the oldest part (Norðurklettar formation). The locations of the three stratigraphic profiles presented in this study are shown in Fig. 7a. The profiles roughly follow the orientation of the two lineaments along which eruptions occur on Heimaey. A section through the Blátindur–Klífið section of Norðurklettar is shown in Fig. 7b. Háin is undoubtedly the oldest unit (cf. Jakobsson, 1968) as it makes up a large basal part of present-day Norðurklettar and outcrops at Stafsnes (WNW of Blátindur, Fig. 3) under the Blátindur unit. Háin is followed by Blátindur,

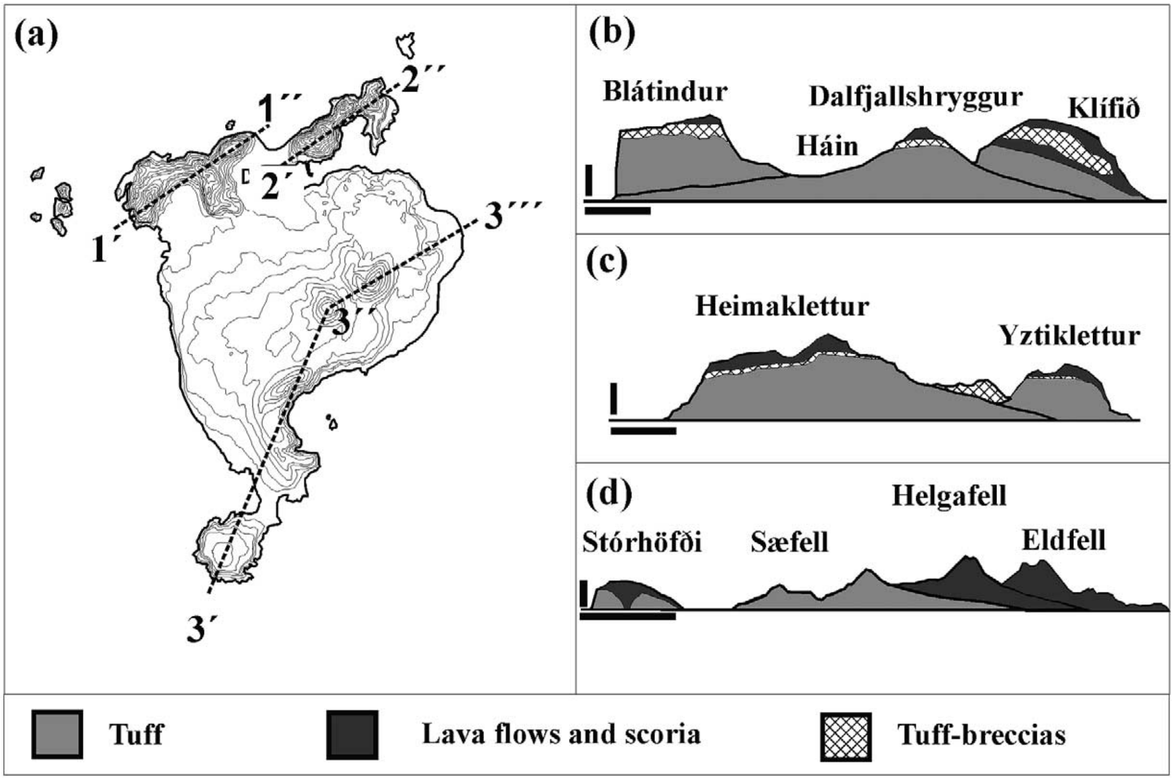

Fig. 7. Schematic sketch of the stratigraphy of Heimaey. (a) Locations of the profiles. (b) Profile 1′-1″. (c) Profile 2′-2″. (d) Profile 3′-3″-3‴. Horizontal scale bars (black) in b and c represent 500 m, whereas in d it represents 1 km. Vertical scale bars (black) in b, c, and d all represent 100 m.

which is deposited upon Háin. Klífið is deposited on the northern flank of the Háin tuff ring, but overlies Blátindur in its western extremities. The youngest unit in Norðurklettar is Dalfjallshryggur, which is probably the remnant of a 30–40 m thick lava lake. Outcrops of the Dalfjallshryggur lava lake are only preserved along the Dalfjallshryggur ridge (overlying Blátindur) and between Háin and Klífið (overlying both). The only unit that exhibits considerable erosion prior to the following eruptions is Háin, indicating a longer time span between the Háin and Blátindur eruptions than for the rest of the formation. The second profile (Fig. 7c) shows the remaining two units belonging to the Norðurklettar formation (i.e. Heimaklettur and Yztiklettur). The third profile (Fig. 7d) comprises Stórhöfði, Sæfell, Helgafell and Eldfell. Mapping these units is rather straightforward and simple as exposures are common and easily accessible. The Stórhöfði and Sæfell units are capped by thin soil covers.

6. Petrography and geochemistry

6.1. Petrography of the Heimaey lavas

The Heimaey lavas are mainly plagioclase-porphyritic (0–12%) alkali basalts with minor amounts of olivine macrophenocrysts (1–3%). The most primitive units have olivine as the only phenocryst phase (6–10%). Clinopyroxene phenocrysts have only been found in the coarse-grained Klífið lava lake. The groundmass of the lavas is fine-grained, exhibiting intergranular textures, and consists of plagioclase, olivine, clinopyroxene and titanomagnetite. Chromite inclusions are present in olivine phenocrysts in the Háin and

Stórhöfði units. Jakobsson (1968, 1979a) reported apatite as a minor groundmass mineral. Segregation veins and pockets have been reported from the Klífið lava lake and the Helgafell lava (Jakobsson, 1968). The segregation veins consist of clinopyroxene (aegirine-augite), titanomagnetite, ilmenite, nepheline, apatite, amphiboles and alkali feldspar.

6.2. Petrography of the Heimaey tuffs

The tuffs consist of brown sideromelane glass with a minor amount of tachylitic glass and crystal fragments (olivine+plagioclase). All tuffs have been subject to variable degrees of palagonitisation, which can be a very rapid process as shown by Jakobsson et al. (2000). During palagonitisation the main components to be leached out of the sideromelane glass are (in order of volume basis): Na_2O, CaO, Al_2O_3, K_2O, SiO_2 and MgO (Jakobsson, 1979b). Instead, H_2O enters the glass, and ferrous iron is oxidised to ferric iron, which gives the tuffs the characteristic rust brown colour (Jakobsson, 1979b). In the Skiphellir drill core, new minerals appear with depth in the tuffs as the extent of the hydrothermal alteration increases, e.g. chlorite, chalcedony, calcite and various zeolites. Commonly the tuffs in the drill core contain whole or broken phenocrysts of mainly olivine and plagioclase (Palmason et al., 1965).

6.3. Petrology of the extrusives

New major element analyses of the majority of the Heimaey extrusives were carried out by ICP-AES at the University of Iceland (Table 1). The volcanic products of Heimaey form a mildly alkaline suite (Fig. 8), ranging from alkali olivine basalt to hawaiite that is approaching a mugearitic composition (nomenclature after Le Bas et al., 1986). Fig. 9 illustrates the major element variation with the relative age of the units, for both lineaments on Heimaey (as deduced from the stratigraphic mapping). A similar evolution with time in the two lineaments on Heimaey is evident. The first eruptives in each lineament are the most primitive (11.1 wt% and 9.7 wt% MgO respectively) and the last eruptives the most evolved (3.3 wt% and 4.0 wt% MgO respectively). Attempts to explain the petrogenesis of the Vest-

Table 1
Major element compositions of the Heimaey rock suite

Sample	HEM01	HEM02	HEM03	HEM04	HEM05	HEM06	VE76[a]	HEM07	VE111[b]
Type	lava	lava	lava	lava	lava	lava	tuff	lava	lava
Classif.	AB	AB	AB	AB	hawaiite	AB	AB	AB	hawaiite
CIPW	hy	ne	ne	ne	ne	hy	ne	ne	ne
SiO_2	48.0	49.0	47.3	47.2	49.2	48.0	47.1	48.5	49.5
TiO_2	1.96	2.10	2.76	2.67	3.46	1.80	2.67	2.67	2.38
Al_2O_3	14.7	16.0	15.4	16.1	15.9	14.0	15.8	16.9	16.3
FeO[c]	10.9	10.5	13.0	12.4	12.7	10.8	12.8	11.4	12.5
MnO	0.18	0.17	0.21	0.21	0.21	0.18	0.19	0.18	0.26
CaO	11.1	11.1	10.9	10.8	8.9	11.6	10.7	10.4	6.95
MgO	9.71	6.94	6.35	6.12	3.99	11.06	5.54	5.47	3.26
Na_2O	2.71	3.39	3.34	3.74	4.51	2.14	3.87	3.62	5.61
K_2O	0.46	0.56	0.47	0.54	0.83	0.29	0.67	0.69	1.49
P_2O_5	0.02	0.07	0.09	0.09	0.11	0.03	0.33	0.09	0.73
Total	99.8	99.9	99.9	99.9	99.9	99.8	99.8	99.9	99.0
Mg#	47	40	33	33	24	51	34	32	21

HEM01 = Háin, HEM02 = Blátindur, HEM03 = Klífið, HEM04 = Heimaklettur, HEM05 = Dalfjallshryggur, HEM06 = Stórhöfði, VE76 = Sæfell, HEM07 = Helgafell, VE111 = Eldfell.
AB = alkali basalt.
[a] Analysis by Thy (1983).
[b] Analysis by Jakobsson et al. (1973).
[c] All Fe as FeO.

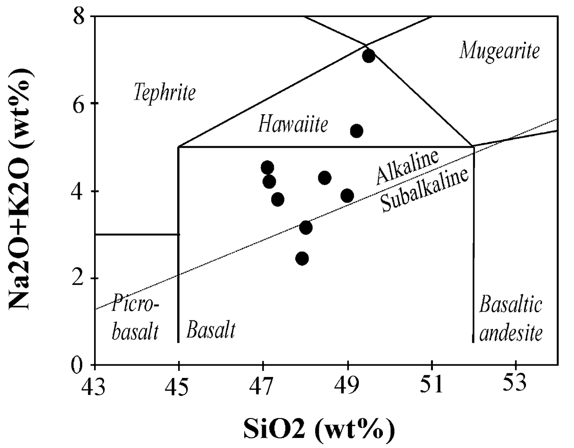

Fig. 8. TAS plot of the Heimaey rock suite. Alkali–subalkali division line by MacDonald (1968).

mannaeyjar basalts have previously invoked models involving contamination of a parental olivine tholeiitic magma by a ne-normative crustally derived melt produced by the breakdown of amphibole (Oskarsson et al., 1982, 1985; Steinthorsson et al., 1985). However, these models failed to account for various trace element and isotope data (Furman et al., 1991). The petrogenesis of the Heimaey rock suite has in more recent publications been attributed to various degrees of partial melting of a lherzolitic mantle source (Thy, 1991; Furman et al., 1991).

6.4. Existence of a magma chamber beneath Heimaey?

The absence of low-pressure equilibrium phenocryst assemblages (i.e. co-existing olivine, plagioclase and clinopyroxene) in the Vestmannaeyjar extrusives has been used to suggest a rapid ascent of magma prior to the eruptions, without longtime ponding in shallow crustal magma chambers (Thy, 1983, 1991; Sigmarsson, 1996). However, the presence of a crustal magma chamber at depth beneath Heimaey has been inferred from several

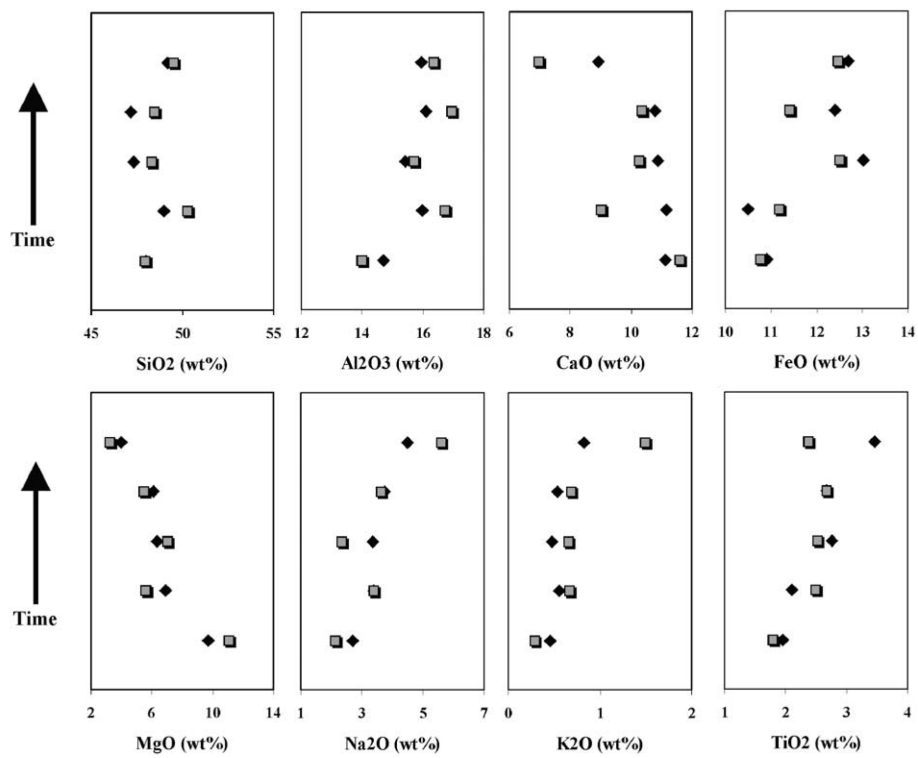

Fig. 9. Variation in major element compositions with relative age of the Heimaey units. Grey squares: the Stórhöfði–Eldfell lineament, black diamonds: the Norðurklettar lineament.

lines of evidence: (1) earthquakes accompanying the 1973 Eldfell eruption outlined a spherical body at 15–25 km depth in the crust (Einarsson and Björnsson, 1979); (2) a zone interpreted as being of high melt concentration was depicted at 10–15 km depth beneath Heimaey using seismic refraction techniques (Gebrande et al., 1980); and (3) in order to generate the compositional variation observed in the Eldfell eruptive products, fractional crystallisation at a pressure of approximately 6 kbar is required, i.e. at approximately 20 km depth (Furman et al., 1991). Sigmarsson (1996) suggested, based on U-series disequilibria, that the Surtsey and Eldfell magmas were derived from the same magma batch and that the Eldfell magma resided and fractionated in the crust for an additional period of 10 years relative to the Surtsey magma.

7. Volcanic history of Heimaey

7.1. Norðurklettar

The oldest part of Heimaey (i.e. the Norðurklettar formation) has been suggested to have formed, at least partly, at the end of the last glaciation (Jakobsson, 1968, 1979a). Jakobsson based his suggestion on the high altitude of the tuff–lava transition in the units (varying between 220 and 160 m a.s.l.). However, the oldest unit in Norðurklettar, i.e. Háin, consists of tuff up to approximately 60 m (a.s.l.), followed by minor lava flows and a scoria cone nested inside the tuff ring. There are no indications that suggest that Háin was formed, or eroded, in the presence of a glacier. However, the tuff ring was subject to considerable erosion before the next eruptive unit was deposited (Blátindur). Measurements of the tuff–lava transition altitude combined with the new stratigraphic mapping presented in this study show that the five remaining units comprising the Norðurklettar formation exhibit a successive decrease in transition altitude (Table 2), from 220 m to 160 m a.s.l.

In order to understand the variable tuff–lava transition altitudes it is important to know the deglaciation history of the area. The deglaciation of southern Iceland is marked by two periods of cooling and glacial advance (Geirsdóttir et al., 2000): (1) the Younger Dryas cooling event (YD, 10–11 kyr ^{14}C), which has generally been pictured as a period of considerable cooling and extensive glaciation (Ingólfsson et al., 1997), and (2) the Preboreal Oscillation (PBO, 9.7–9.9 kyr ^{14}C) which was of shorter duration and lesser extent (Björck et al., 1996, 1997; Hald and Hagen, 1998).

It has been suggested that during the YD a vast glacier extended onto the shelf of southern Iceland (Hjartarson and Ingólfsson, 1998), which also explains why there are no marginal moraine deposits found on land today. However, Geirsdóttir et al. (2000) argued that the glacier advanced to the same location (Búði morainal complex approximately 20 km north of Heimaey; Haraldsson, 1981) during both the YD event and the PBO. The findings of Geirsdóttir et al. contradict the idea of a vast glacier present on the southern shelf during this period, which is consistent with the absence of exposed subglacial volcanics on Heimaey. A previous study by Sigmundsson (1991) has shown that post-glacial rebound was rapid in central Iceland (up to 0.5 m/year). If applied to Heimaey, with a distance of 160 km from the inferred glacial centre, the same model by Sigmundsson yields a rebound rate of 0.13 m/year. Assuming a glacial rebound rate of 0.13 m/year for Heimaey, the 60 m successive lowering in tuff–lava transition altitude over time (excluding Háin) corresponds to a period of 460 years. Hence, the entire Norðurklettar formation might have been deposited during a 500-year interval.

A possible scenario for the timing and forma-

Table 2
Altitude of tuff–lava transitions in Norðurklettar

Unit	Approximate altitude of tuff–lava transition (m a.s.l.)
Háin	60
Blátindur	220
Klífið	190
Heimaklettur	180
Yztiklettur	160

Note the successive lowering with time (i.e. Blátindur to Yztiklettur).

tion of Norðurklettar may then be put forward as follows: (1) formation of the Háin tuff ring immediately after the YD event (approximately 10 ka), when the glacier had retreated to some extent and the sea level was quite low; (2) glacial advance in southern Iceland at 9.9 ka (marking the onset of the PBO), resulting in renewed compression of the crust (higher sea level relative to Heimaey); (3) Blátindur and the subsequent units in Norðurklettar were erupted during the glacial retreat following the second cooling period (∼9.5 ka), as reflected by the successive lowering of the transition altitude (Table 2). The time estimate, however, is somewhat uncertain because the altitude of the tuff–lava transition is variable and does not occur at any given height above the waterline.

7.2. Stórhöfði, Sæfell, Helgafell and Eldfell

Jakobsson (1968, 1979a) suggested, based on ^{14}C datings of the Sæfell eruption (5470 ± 160 years, average of six samples; Kjartansson, 1967) combined with the observed stratigraphy (Fig. 7), that Stórhöfði was only slightly older than Sæfell and that Helgafell was slightly younger than, or possibly simultaneous with, the Sæfell eruption. Jakobsson, therefore, proposed a second volcanic episode occurring in the Vestmannaeyjar system at 5–6 ka, as some of the other islands appear to be of a similar age. Sigmarsson (1996) estimated the age of the Helgafell eruption at 5.9 ± 0.3 ka based on U-series disequilibria.

The 1973 Eldfell eruption was closely monitored and for more detailed information about the eruption the reader is referred to Jakobsson et al. (1973), Self et al. (1974), Williams (1997) and Morgan (2000). Jakobsson (1979a) further suggested a third, currently ongoing, volcanic episode on Heimaey, including the Surtsey and Eldfell eruptions.

7.3. Episodic volcanism?

Volcanism in the Vestmannaeyjar system has been postulated to be episodic, with three major phases occurring at approximately 10 ka (Norðurklettar, etc.), 6–5 ka (Stórhöfði, Sæfell, Helgafell), and a current episode including the Eldfell (1973) and Surtsey (1963–1967) eruptions (Jakobsson, 1968, 1979a). Jakobsson based his suggestion on the observed stratigraphy, the ^{14}C datings by Kjartansson (1967) of one eruption, combined with tephra layers found in soil profiles from some of the islands in the Vestmannaeyjar system.

Heimaey is surrounded by some 60–75 submarine peaks (Fig. 4), which have been interpreted as craters of mainly Holocene age (corresponding to approximately 75% of the vents in the Vestmannaeyjar system). Without any possibility of dating the submarine craters, we find it difficult to argue any periodicity in the Vestmannaeyjar volcanics. Furthermore, alkali basalts are generally considered to form by a small degree of melting at depth, which is consistent with the relatively small volumes erupted in a single eruption in the Vestmannaeyjar system. The exception is the Norðurklettar formation, where the variable altitude of the tuff–lava transition may serve as an indicator of higher magma production rates, immediately following the deglaciation.

8. Volume of erupted material

The calculated volumes for the different eruptions on Heimaey are shown in Table 3. In order to calculate dense rock equivalents (DRE), a porosity of 10% was assumed for the lavas (as estimated by vesicle measurements in thin sections

Table 3
Volume of erupted material on Heimaey

Unit	Tuff (km^3)	Lava and scoria (km^3)	Total (km^3)	DRE (km^3)
Háin	0.74	0.27	0.91	0.75
Blátindur	0.05	0.07	0.12	0.1
Klífið	0.05	0.01	0.06	0.05
Dalfjallshryggur	0.00	0.02	0.02	0.01
Heimaklettur	0.24	0.10	0.34	0.28
Yztiklettur	0.07	0.03	0.10	0.08
Stórhöfði	0.05	0.03	0.08	0.06
Sæfell	1.30	0.00	1.30	1.04
Helgafell	0.00	0.65	0.65	0.59
Eldfell	0.00	0.23	0.23	0.21
Total	2.49	1.41	3.80	3.18

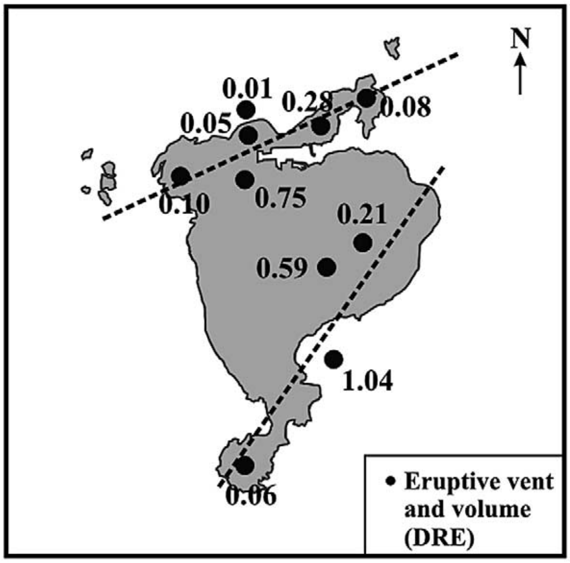

Fig. 10. Erupted volumes (DRE km^3) for the 10 vents located on Heimaey. Note the increase in volume towards the central part of each lineament.

from the Helgafell lava flows; Mattsson and Höskuldsson, submitted) and an effective porosity of 20% for the tuffs. It is interesting to note that the most voluminous eruptions occurred in the central parts of each lineament on Heimaey (Fig. 10), which was also the case during the Surtsey eruption. Eruptions on Heimaey have an average volume of 0.38 km^3 (0.32 km^3 DRE), which is twice the value reported as an average for the Vestmannaeyjar system (Jakobsson, 1979a).

9. Summary and conclusions

The Heimaey volcanic centre is the youngest volcanic centre in the southward-propagating EVZ, and thus represents the initial activity at the tip of the rift. The volcanic centre is composed of roughly equal amounts of tuff/tuff-breccias and lavas as most eruptions involve both a phreatomagmatic and an effusive phase. The eruption frequency for Heimaey during the Holocene is on average 10^{-4} eruptions/year, with the possible exception of Norðurklettar, which formed during a 500-year interval immediately following deglaciation (10^{-2} eruptions/year). The average volume of material produced in a single eruption on Heimaey is 0.32 km^3 (DRE), which is twice the value reported for the Vestmannaeyjar system as a whole. The chemistry of the extrusives is predominantly alkali basaltic belonging to the sodic series. The compositional evolution of the lavas erupted along the Norðurklettar lineament, from primitive (9.7 wt% MgO) to evolved (4.0 wt% MgO), may indicate their derivation from a common source, especially considering the relatively short time span during which they were erupted. The absence of equilibrium phenocryst assemblages in the Heimaey lavas suggests that magma rose quickly from depth, without long-time ponding in shallow crustal magma chambers. Further, the regularity of the lineaments along which eruptions occur (N45°E and N65°E) suggests that it is seismic events associated with the southward propagation of the EVZ that opened the pathways that allowed magmas to reach the surface. An increase in the frequency of volcanic eruptions in the Vestmannaeyjar system might be expected with the continuation of southward propagation of the EVZ, and it is then likely that Heimaey will develop into a central volcano like the mature volcanic centres situated on the Icelandic mainland.

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Manuscript 2

Reconstruction of the Helgafell eruption, Vestmannaeyjar, South Iceland

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Abstract

The 5900 BP Helgafell eruption started with Hawaiian activity, depositing tephra and emplacing clastogenic spatter-fed lava flows near the vent. There is a clear geochemical trend in the Helgafell lava flows, dominated by fractionation of olivine. However, the fractionation trend is partially overprinted by processes such as incorporation of plagioclase xenocrysts and late stage segregation during solidification. When the effusion rate decreased, the activity at the vent shifted from Hawaiian fire fountains to Strombolian. Already nucleated crystals of olivine and plagioclase grew rapidly, significantly affecting the viscosity of the lava. The lava morphology also shifted with the change in effusion rate, beginning with widespread spatter-fed flows and changing to P-type pahoehoe transported to the active flow front through a system of tubes inside the lava-field. The decrease in effusion rate combined with cooling and degassing of the magma increased viscosity of the lava, which resulted in frequent clogging of lava tubes and subsequent formation of inflation features such as tumuli. By measuring lava inflation clefts in tumuli in the Helgafell lava field a minimum eruption duration of nine months can be estimated. The calculated volume-time ratio for Helgafell (0.06) is somehow higher than that of 1973 Eldfell eruption (0.04) and 1963-1967 Surtsey eruption (0.02). Taking into account that this is a minimum estimate of eruption duration and any increase in time will lower the volume-time ratio. Approximately 0.6 km^3 lava (DRE) was emplaced in the eruption, which abridged pre-existing islands to form Heimaey.

Introduction

The 228 m high Helgafell scoria cone is located on the island of Heimaey, Southern Iceland (Fig. 1a). Heimaey is the largest (13.4 km^2) of the seventeen islands comprising the Vestmannaeyjar volcanic system. The island is believed to be the volcanic centre in the Vestmannaeyjar system. Previous studies have almost exclusively focused on the 1963-1967 Surtsey eruption and 1973 Eldfell eruption (Jakobsson et al. 1973; Kokelaar and Durant 1982; Furman et al. 1991; Jakobsson 1992; Sigmarsson 1996; Thordarson 2000). Jakobsson (1968, 1979) estimated the age of the Helgafell eruption to be approximately 5 ky based on stratigraphy and a C^{14}-dating of the volcanic deposits from the Sæfell tuff ring forming eruption, carried out by Kjartansson (1967). Jakobsson based his suggestion on an assumption that the Helgafell eruption was almost simultaneous with that of the neighbouring Sæfell tuff ring. He further suggested that the lava flows from the Helgafell eruption abridged pre-existing islands in the north and south into one. A new age estimate (5900 ± 300 years) of the Helgafell eruption by Sigmarsson (1996) based on U-series

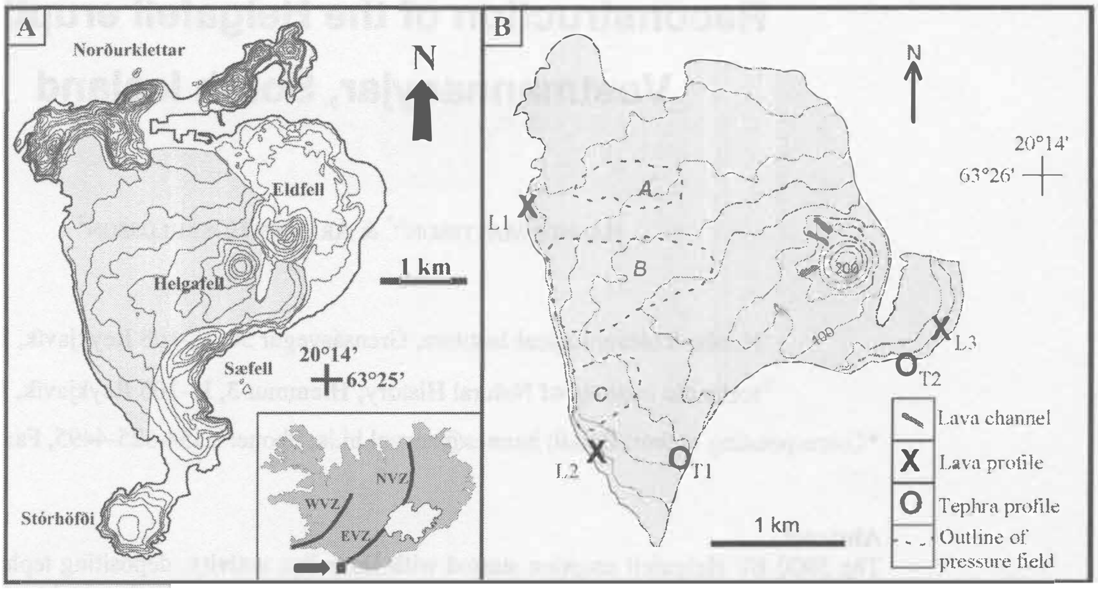

Fig. 1. (A) Map showing the volcanic zones of Iceland and the location of Heimaey. The Helgafell lava field is outlined in light grey. Abbreviations are as follows: WVZ = Western Volcanic Zone, EVZ = Eastern Volcanic Zone, NVZ = Northern Volcanic Zone. (B) Detailed map of the Helgafell lava field showing sample locations and the outlines of the tumuli areas.

disequilibria supports the approximate age concluded by Jakobsson. Cameron et al. (1992) reported a Bouguer gravity anomaly of Heimaey. Based on their findings, they suggested that the Helgafell vent was emptied in the final stage of the eruption and would now be filled with low-density scoria and ash. In this paper, we present a detailed mapping of the Helgafell lava field combined with extensive measurements of lava inflation clefts in tumuli and pressure ridges. By using the methods described by Hon et al. (1994) and Rossi and Gudmundsson (1996) we can approach time of inflation in the lava flow field. We will show that an estimate of the total eruption duration can be made in fossil lava fields by combining stratigraphic mapping and cleft measurements. Geochemical data indicate that fractionation processes dominated the compositional evolution of the Helgafell lava, but also include modification by incorporation of xenocrysts and late stage segregation.

Geological setting

The Vestmannaeyjar system is the southernmost of eight volcanic systems that form the Eastern Volcanic Zone (EVZ) of Iceland. There is a spatial change in chemistry of the lavas erupted in the EVZ, from tholeiites in the north via transitional FeTi-basalts to alkali basalts in the south. The variation has been attributed to the southward propagation of the EVZ with the rift-tip currently situated beneath Vestmannaeyjar (Saemundsson 1978; Oskarsson et al. 1982; Meyer et al. 1985; Oskarsson et al. 1985). A similar spatial change in composition has been reported from a propagating segment of the Galapagos spreading centre (Christie and Sinton 1981). Heimaey is the volcanic centre in the Vestmannaeyjar system with at least ten individual eruptions during the Holocene. The most recent eruption being Eldfell in 1973. The presence of a magma chamber at 10 – 15 km depth beneath Heimaey has been inferred from seismic data (Gebrande et al. 1980). Earthquakes accompanying the 1973 Eldfell eruption outlined a spherical body in the crust at approximately the same depth (Einarsson and Björnsson 1980). Additional seismic evidence indicates a zone of partial melting beneath Heimaey at 20 – 30 km depth (Palmasson 1971; Gebrande et al. 1980), which is believed to represent the zone where the parental Vestmannaeyjar magma is separated from the source. Alkali olivine basalt is the dominating rock type erupted on Heimaey, although two occurrences of hawaiites approaching mugearitic composition exist (Jakobsson 1979). Eruptions on the island follow the main trend of the EVZ (N45°E; Saemundsson 1979) or occur in an en-echelon arrangement (N65°E) with eruptive fissures sub-parallel to the main trend (Jakobsson 1968). Most of the eruptions on Heimaey are believed to have started with a phreatomagmatic phase, shifting to an effusive phase as the influence of seawater diminished, similar to the evolution observed in the 1963 - 1967 Surtsey eruption (Thorarinsson 1966, 1969). Exceptions are the Dalfjallshryggur, Helgafell and Eldfell eruptions, which were sub-aerial. The Sæfell eruption only produced a small lava flow from a parasitic vent in the tuff cone but was otherwise strictly phreatomagmatic (Leys 1983).

Methods

The lava flows exposed in the coastal sections were photographed from the sea, using a zodiac. The photographs were used to map individual flows and to establish their relative chronology. Twenty-three lava samples from various stratigraphic levels and three tephra samples were collected in order to depict eventual chemical variation of the eruptive products with time. The samples were crushed and ground in an agate mill prior to analyses. Major elements and selected trace elements were analysed in whole-rock samples by ICP-AES at the University of Iceland, using a modified version of the procedure described by Govindaraju and Meville (1987). The precision of the ICP-AES analyses was determined by duplicate analyses of alkali olivine basalts. The relative errors of analyses were better than 5 % for the major elements, with exceptions for Na_2O (7 %), K_2O (10 %) and P_2O_5 (10 %). Ni, Cr, Sr and V have a relative error of less than 7 %. Rare earth elements (REE) were analysed in six lava samples and one tephra sample by inductively coupled plasma mass spectrometry (ICP-MS) at CNRS-Nancy. Polished thin sections were prepared of the same six samples and analysed for mineral chemistry, using an ARL-SEMQ electron microprobe at the Nordic Volcanological Institute. Analytical conditions were 15 kV accelerating voltage, 15 nA beam current, focused beam with counting times ranging from 20 to 40 seconds, using natural standards. To constrain the duration of the eruption, depth of inflation clefts were measured in over 300 tumuli on the Helgafell lava field. A minimum time for lava supply corresponding to a minimum duration of the eruption was calculated with the equation presented by Hon et al. (1994, p. 361, eq. 2).

Results

Stratigraphy

The Helgafell lava flows have the surface texture of rough pahoehoe and cover an area of 6.6 km^2, or close to 50 % of the island. Two tephra sequences were found at locations T1 and T2 (Fig. 1b). The T1-tephra sequence is 10 – 15 cm thick and deposited directly upon a discordant eroded surface of the Sæfell tuff cone and overlain by the Helgafell lava. The T2-tephra is deposited on top of a 32 cm thick soil layer, which in turn rests on an eroded surface of the Sæfell tuff cone.

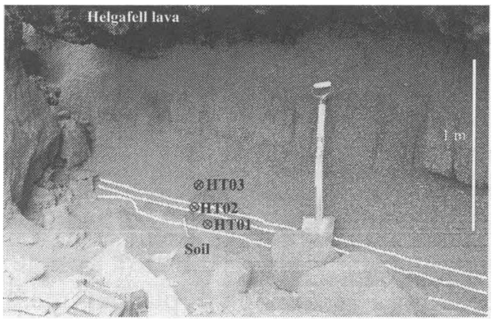

Fig. 2. Tephra profile at location T2. White lines marks contact between the HT01, HT02 and HT03 horizons.

The T2 sequence is located closer to the vent than the T1-sequence, and the total thickness of the tephra is 132 cm. The tephra horizon at location T2 can be divided into three horizons (HT01, HT02, and HT03) based on colour and grain-size (Fig. 2). The HT01 and HT03 horizons are 8 and 120 cm thick, respectively, and have identical physical appearance, but are separated by the finer grained, heavily oxidized HT02-layer (4 cm thick). The top of the T2-tephra sequence is almost completely undisturbed and gently capped by the lowermost Helgafell lava flow. Tephra grains at both the T1 and T2 locations exhibit ragged, angular shapes and high vesicularity, indicating Strombolian activity (Heiken 1974, Fisher and Schminke 1984, Vergniolle and Mangan 2000).
The lava flows on the eastern side of the Helgafell lava field (Fig. 3a) are relatively thin (~ 0.5 - 2 m) with considerable lateral extension (< 200 m) and have high vesicularity. The flows contain numerous volcanic bombs, which are rotated in the lava flows (Fig. 3b). The bombs range in size from 5 – 35 cm in diameter and commonly have a 5 - 15 mm vesicular rim at the contact with the surrounding lava. All examined bombs show the same plagioclase-phyric appearance as the host lava. Individual lava flows are separated by 0.1 - 1.5 meters of scoria, with the thicker scoria horizons in the lower stratigraphic sections (early eruptives).
Lava flows on the western side of the lava field are significantly different in physical appearance from those on the eastern side (Fig. 3c). Flow-lobes dominate and individual flows seldom exceed 20 m in width. The lava flows on the western side also show much lower vesicularity. Pipe vesicles up to 1 m in length are common in the massive interior of the flow lobes (Fig. 3d). Scoriaceous horizons separating individual flows are thinner than on the eastern side (< 15 cm) and poorly developed pahoehoe surface textures are occasionally observed in between the flows. Close to the vent, on the slopes of the scoria cone, three lava channels run towards the N and W of the crater (Fig. 1b). The channels can be traced to a distance of 200 m from the vent. Two minor crater overflows occur near the top on the northern and eastern side of the Helgafell scoria cone. No signs of lava/water interactions, such as littoral cones, are found at any location in the lava field.

Tumulus formation and use as time indicator

Tumuli are common inflation features in pahoehoe flow fields (Nichols 1936; Chitwood 1987; Walker 1991; Rossi and Gudmundsson 1996; Duraiswami et

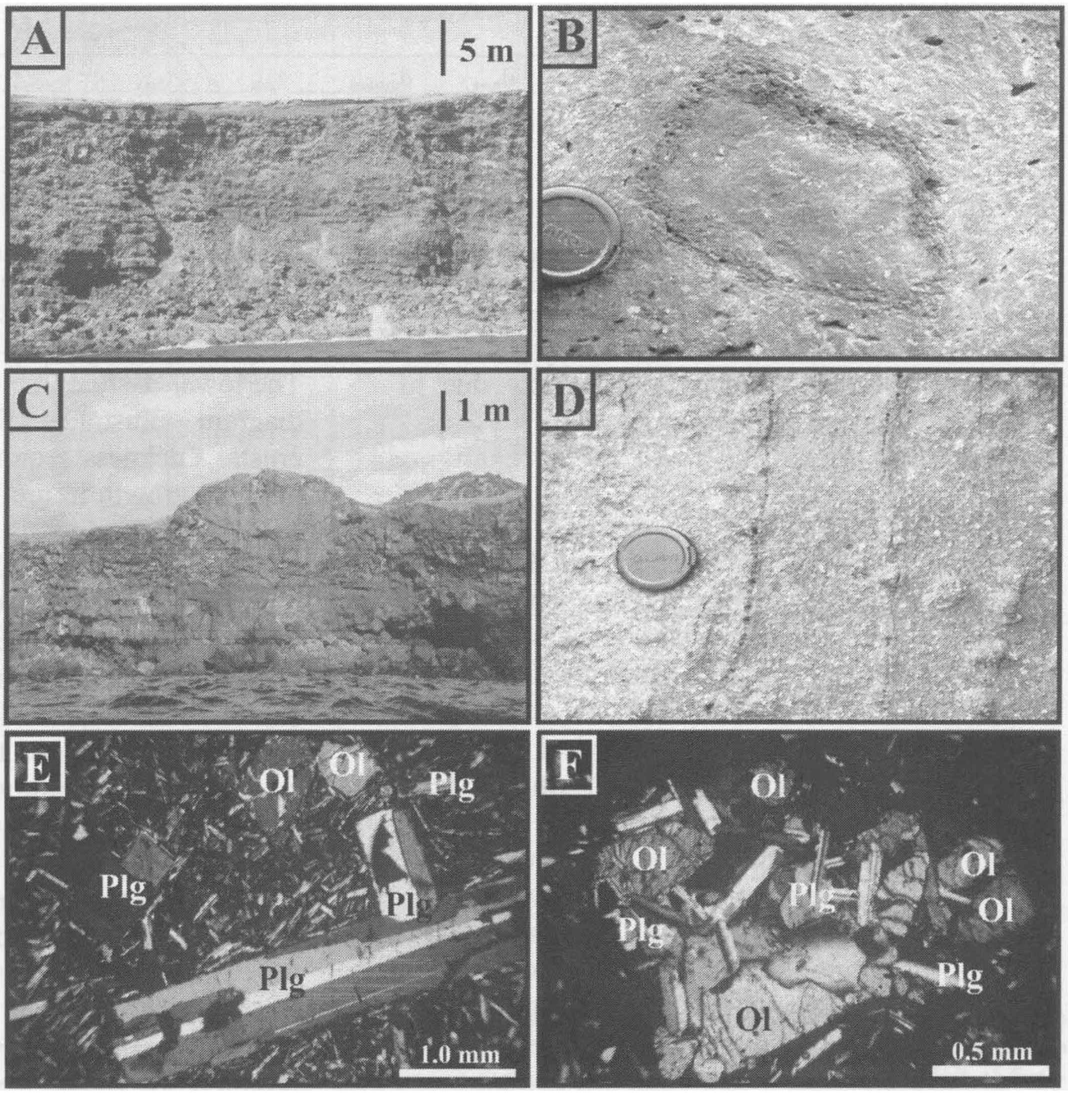

Fig. 3. Typical features observed in the Helgafell lava field. (A) Wide-spread flows characterising the eastern side of the island, (B) Bomb rotated in lava flow at location L3 (Fig. 1b), (C) Lava flows and flow-lobe tumuli in cross-section on the western side of the island, (D) Pipe vesicles in lava from location L1 (Fig. 1b), (E) Thin section of the Helgafell lava showing seriate texture with a large plagioclase phenocryst (cross-polarized light), and (F) Thin section showing olivine and plagioclase laths growing in a glomerophyric cluster (cross-polarized light).

al. 2001), and have also been reported in submarine lava flows (Fornari et al. 1978; Appelgate and Embley 1992; Umino et al. 2000). Tumuli are domed uplifts (Fig. 4a), which are cross cut by axial and radial fractures ('lava-inflation clefts' of Walker, 1991). The axial inflation cleft dominates (Rossi and Gudmundsson, 1996), and can be used as an indicator of flow direction. Tumuli form when lava is supplied to the interior of a flow and its fluid pressure rises, causing previously formed crust to inflate and buckle upwards forming the characteristic fractured whaleback shape (Swanson 1973; Appelgate and Embley 1992; Rossi and Gudmundsson 1996). The inflation may result from two processes: (1) clogging of individual tubes in a flow field, or (2) when a cooling flow front ceases to advance but is still supplied with lava through a system of tubes generating more regional inflation. By measuring depths of inflation clefts, the time of inflation can be estimated by using the formula presented by Hon et al. (1994). The formula describes the

Area	n#	d_{min}	d_{max}	d_{median}	t_{min}	t_{max}	t_{median}
A	20	0.97	3.45	1.62	6.5	82	39
B	318	0.31	2.77	1.03	0.7	53	24
C	20	0.77	2.52	1.42	4.1	44	34

Table 1. Result of inflation cleft measurements and calculations of time during which active inflation occurred. n# denotes the number of clefts measured, d the depth in meters, and t equals the calculated time in days of active inflation. Location of areas A, B, and C, is shown in Fig. 1b.

cooling rate of a lava flow due to conductive cooling:

$$t = 6.87 \times d_{max}^{\ 2} \qquad (1)$$

where t equals the time in days of active inflation and d_{max} is the maximum crustal thickness in meters (i.e., maximum depth measured in an inflation crack). Carslaw and Jaeger (1959) gave the theoretical expression behind the formula. The values used in the original calculation by Hon et al. are based on measurements from Hawaiian basalts, but can be used as an approximation for Icelandic equivalents as shown by Rossi and Gudmundsson (1996).

Measurements of inflation clefts

Areas with tumuli occur at three locations on the Helgafell lava field (Fig. 1b), which are located relatively close to the coastline. Depth and direction of inflation clefts were measured in over three hundred tumuli and the results of measurements and calculations of active inflation are shown in Table 1. Inflation cleft depths range from 0.31 to 3.45 m, highest value being measured in pressure field A. The distribution of the number of clefts versus their depth is shown in Fig. 4b. The most common value for the cleft depths ranges from 0.7 to 0.9 m. Figure 4c shows a plot of measured depth vs. direction of the axial cleft. The lower dashed line (0.3 m) in the diagram illustrates the minimum crustal thickness required to initiate tumulus growth in the Helgafell lava field. This value is identical to crustal thickness (0.2 – 0.4 m) reported by Rossi and Gudmundsson (1996). The area enclosed by the dashed lines is uniform and no trend can be distinguished. This field probably corresponds to inflation of capillary tubes inside the lava field. The area above the upper dashed line, characterised by deeper inflation clefts and showing a trend of deeper clefts towards north, most likely reflects main feeder tubes in the field. The trend in Figure 4c becomes more apparent if the time of inflation is plotted on the y-axis instead of depth due to the exponential relationship between the measured cleft depth and time of active inflation (equation 1). Figure 4d shows that the deepest inflation clefts are located approximately 1.0 – 1.4 km from the vent. The lack of data at 1.2 km from the vent is due to buildings and infrastructure in the town of Vestmannaeyjar, whereas the gaps at 0.7 – 0.9 km and 1.5 km indicate the actual distribution of tumuli. Plotting cleft depth versus altitude gives similar results, showing that the largest clefts are found at the lowest altitude. This is coherent with that the lowest altitude is furthest away from the source and the fact that higher pressure lifts the tumuli roof higher (Fig. 4e). A common feature in the inflation clefts of tumuli in the Helgafell lava

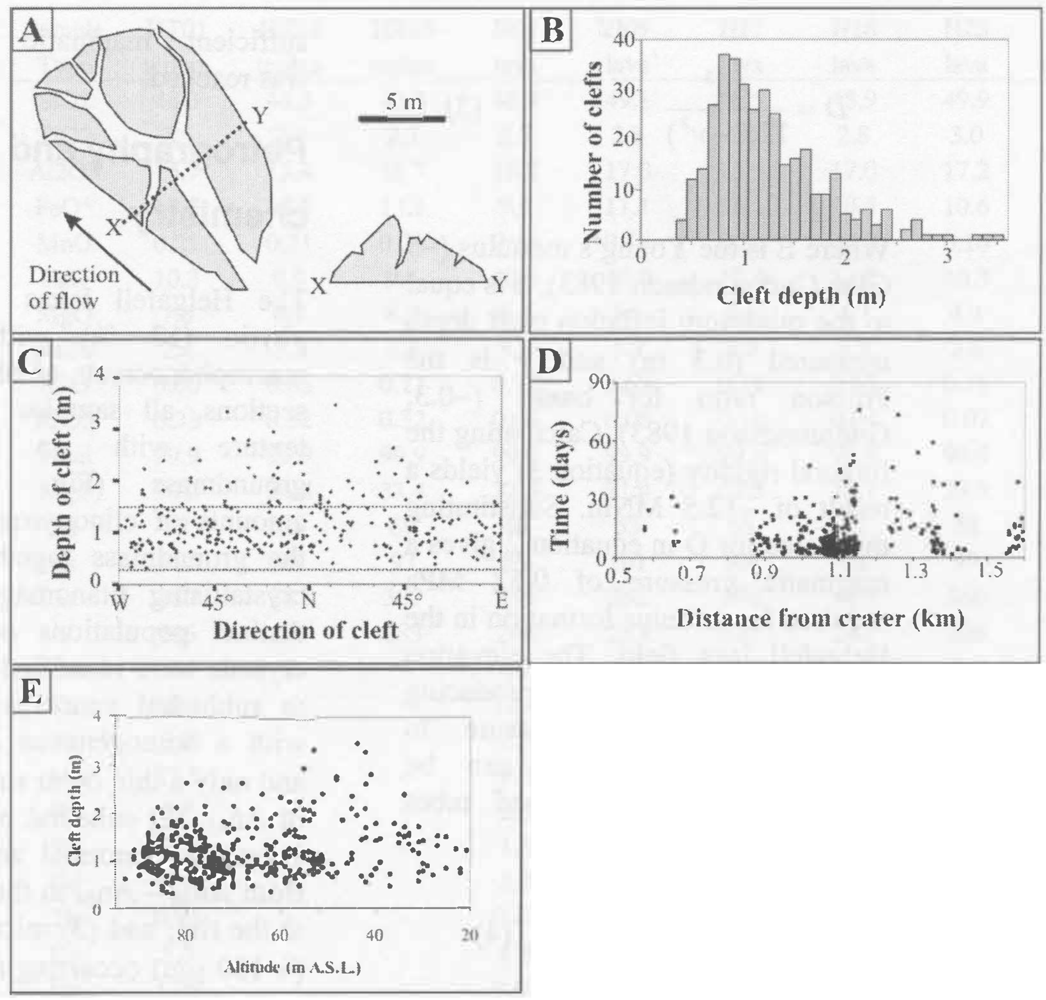

Fig. 4. Results of tumuli measurements in the Helgafell lava field. (A) Typical morphology of a flow-lobe tumulus. Modified after Walker (1991). (B) Distribution of the measured depth in inflation inflation clefts. (C) Plot showing depth versus direction of main inflation cleft of flow-lobe tumuli. Note the lower dashed line (0.3 m) under which no inflation clefts form. (D) Plot of active time of inflation vs. distance from the crater. (E) Cleft depth versus altitude.

field is the lava grooves, developed perpendicular to the direction of the axial cleft. Such features are produced by inflation occurring while the sidewalls of the cleft are hot and can deform easily (Nichols 1939). Squeeze-ups of lava in tumuli clefts occur when the magmatic pressure exceeds the strength of the confining crust and lava escapes (Rossi and Gudmundsson 1996). Squeeze-up features in tumuli are only occasionally observed in the Helgafell lava field. The magmatic overpressure required to form tumuli can be calculated according to the model by Rossi and Gudmundsson (1996). The similar size range of the observed Helgafell tumuli and the tumuli studied by Rossi and Gudmundsson indicates that the same values for tumuli radius (a = 5.5 m) and maximum deflection (ω_{max} = 2.5 m) can be used as an approximation. The pressure required for tumulus growth is:

$$P_m = \frac{64D\omega_{\max}}{a^4}\frac{1+\nu}{5+\nu} \qquad (2)$$

Where P_m is the magmatic pressure (Pa), ν is the Poisson ratio, ω_{max} is maximum deflection in the centre of the tumulus (m), a is the radius at the base of the tumulus (m) and D is the flexural rigidity (Nm) of the crust defined as:

$$D = \frac{Eh^3}{12(1 - \nu^2)} \qquad (3)$$

Where E is the Young's modulus (~5 GPa; Gudmundsson 1983), h is equal to the minimum inflation cleft depth measured (0.3 m) and ν is the Poisson ratio for basalt (~0.3; Gudmundsson 1983). Calculating the flexural rigidity (equation 3) yields a result of ~12.5 MNm. Substituting this value for D in equation 2 gives a magmatic pressure of 0.52 MPa required for tumulus formation in the Helgafell lava field. The elevation difference required for producing sufficient magmatic pressure to favour tumulus growth can be calculated (assuming closed tubes and no leakage) by:

$$\Delta z = \frac{P_m}{g \times \rho_m} \qquad (4)$$

Where Δz equals the elevation difference between the tumulus and the source (m), P_m is the maximum magmatic pressure in a tumulus (Pa), g is the constant of gravity (N/kg) and ρ_m is the density of the magma. The density of the magma was calculated to 2685 kg/m^3 using the MAGMA software for the average composition of the lava based on 23 samples and assuming an average vesicularity of 10 %. The elevation difference required for formation of tumuli in the Helgafell lava field was calculated to approximately 20 meters. The highest located tumulus in the lava field is situated at 90 m A.S.L. and the point where the open lava channels roof themselves over and disappear into the lava field (e.g. uppermost point of lava tubes), at approximately 110 meters A.S.L (Fig. 1). This suggests that tumuli started to form immediately after sufficient magmatic overpressure was reached.

Petrography and mineral chemistry

The Helgafell lavas are feldspar-phyric (12 %) with occasional macrophenocrysts of olivine. In thin sections, all samples show seriate texture with an intergranular groundmass (Fig. 3e). Minor amounts of clinopyroxene occur in the groundmass together with late crystallising titanomagnetite. Three distinct populations of plagioclase crystals were identified: (1) euhedral to subhedral xenocrysts (< 7 cm), with a homogeneous core of An_{71} and only a thin outer rim (5 - 10 μm) of An_{55}; (2) euhedral phenocrysts (~ 1 cm) with normal zoning ranging from An_{68} – An_{64} in the core to An_{55} at the rim; and (3) microphenocrysts (< 100 μm) occurring as laths in the groundmass with a homogeneous composition of An_{53}. Olivine occurs as both phenocrysts and in the groundmass. The phenocrysts show slight normal zoning Fo_{76} – Fo_{66}, and often form glomerophyric clusters with plagioclase laths (Fig. 3f). The olivines of the groundmass are homogeneous Fo_{62}. Jakobsson (1979) reported apatite as a minor groundmass mineral in the Vestmannaeyjar lavas but it has not been detected in the present study. Jakobsson (1968, 1979) also found small segregation pockets (5 x 5 mm), containing nepheline, apatite, aegerine-augite and glass in the Helgafell lava. No such pockets were analysed in this study, although their presence was noted in some flow-lobes on the western side of the lava field. Two thin sections of the tephra found at location T1 and T2 (Fig. 1b) showed occasional phenocrysts of

Sample	HT01	HT02	HT03	H02	H09	H17	H18	H23	H31
Type	tephra	tephra	tephra	lava	lava	lava	lava	lava	lava
SiO2	48.3	48.3	47.8	48.9	49.1	46.5	48.9	49.9	49.6
TiO2	2.5	2.7	2.7	2.7	2.8	3.1	2.8	3.0	2.6
Al2O3	15.7	15.4	15.7	18.2	17.0	15.8	17.0	17.2	17.2
FeO*	12.5	12.5	13.1	9.9	11.1	13.1	10.5	10.6	10.0
MnO	0.21	0.21	0.21	0.16	0.18	0.20	0.17	0.16	0.16
CaO	10.3	9.9	9.6	10.6	10.0	10.4	11.2	10.3	10.9
MgO	7.0	6.7	6.5	4.8	5.1	6.6	4.7	4.4	4.8
Na2O	2.4	3.2	3.4	3.9	3.7	3.3	3.9	3.6	3.7
K2O	0.66	0.65	0.71	0.72	0.68	0.71	0.67	0.71	0.76
P2O5	0.31	0.32	0.32	0.06	0.08	0.10	0.04	0.07	0.09
Total	99.9	99.9	99.9	99.9	99.9	99.8	99.9	99.9	99.9
Mg#	36.0	34.8	33.4	32.5	31.4	33.3	30.1	29.3	32.4
Ni	72	65	69	46	53	63	42	38	47
Cr	111	98	97	89	101	104	96	99	89
Sr	380	382	387	386	382	378	416	360	358
V	286	271	283	294	283	337	282	326	257
La	15.7			4.7	7.6	7.6	4.8	5.7	4.1
Ce	38.4			12.4	19.3	19.2	13.6	14.9	10.4
Pr	5.2			1.9	2.7	2.6	2.0	2.2	1.5
Nd	22.8			9.0	12.1	12.4	9.6	10.9	7.3
Sm	5.8			2.6	3.6	3.5	2.6	2.9	2.2
Eu	2.1			1.5	1.7	1.7	1.5	1.6	1.5
Gd	6.1			3.2	4.0	3.8	3.3	3.4	2.3
Tb	0.91			0.46	0.65	0.58	0.48	0.56	0.41
Dy	5.5			2.9	3.8	3.7	3.0	3.5	2.6
Ho	1.08			0.60	0.71	0.71	0.62	0.71	0.55
Er	2.8			1.7	2.2	2.0	1.8	2.1	1.5
Tm	0.43			0.27	0.35	0.33	0.30	0.35	0.27
Yb	2.8			1.9	2.2	2.2	2.0	2.4	1.9
Lu	0.42			0.33	0.37	0.37	0.36	0.36	0.34

Table 2. Major element and selected trace element analyses of the three tephra layers at location T2 and six lava samples collected from various stratigraphic levels in the Helgafell lava field. Major elements are given in wt % and trace elements in ppm. *All Fe as FeO.

olivine and plagioclase in a cryptocrystalline groundmass.

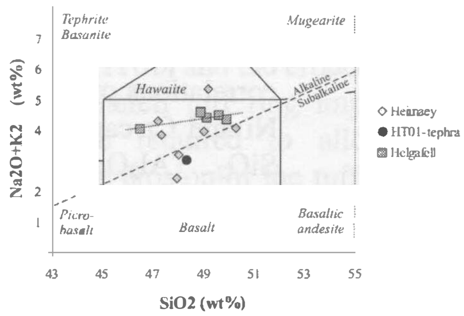

Fig. 5. Total alkali silica diagram (TAS), showing the composition of the Helgafell lavas (squares) compared to the other formations on Heimaey (diamonds). The HT01-tephra sample is included for comparison (black circle). Nomenclature after Le Bas et al. (1986). Dotted line illustrates the trend of the Helgafell lava flows. Alkaline – subalkaline division line from MacDonald (1968).

Geochemistry

Table 2 shows the result of major element and selected trace element analyses of the three layers of tephra found at location T2, and six lava samples collected from various stratigraphic levels in the lava field. The Helgafell lava flows are typical mildly alkaline olivine basalt of the sodic series (Na_2O – 2.0 ≥ K_2O), which is apparent when plotted in a total alkali – silica diagram (Fig. 5).

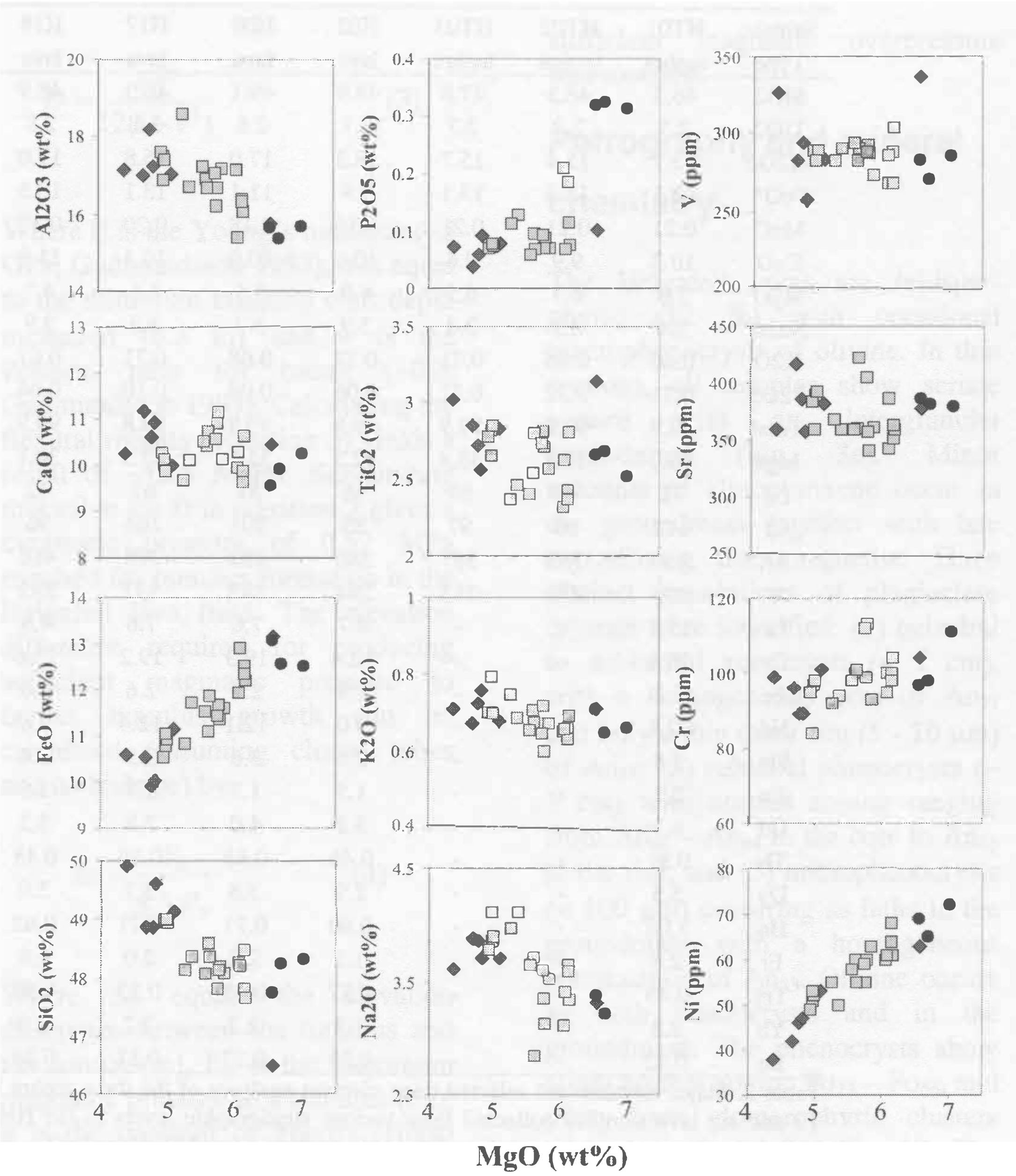

Fig. 6. Major element- and selected trace element variation diagrams for the Helgafell lava flows. Diamonds indicates the samples presented in Table 2, squares additional samples from Helgafell. The black circles represent the three tephra samples.

Major element variation diagrams are presented in Figure 6, in which Ni, Cr, Sr and V are included for comparison. The three tephra samples (HT01, HT02, HT03) are the most primitive in composition (MgO = 7.0 – 6.5 %, variation within error limits), and are most likely to reflect the actual composition of the Helgafell magma. The MgO content of the lava flows vary from 6.6 % in the most primitive flow to 4.4 % in the most evolved. A relative large variation in SiO_2-content is also present (46.5% to 49.9 %). Positive correlations of MgO with FeO, P_2O_5, Ni and Cr can be observed, whereas SiO_2, Al_2O_3, Na_2O, and K_2O correlate negatively. However, no compositional trend with time during the eruption can be detected in the sequential lava flows. The trend for the Helgafell lavas and are shown in a TAS-diagram (Fig. 5). The internal change in REE within time during the eruption is shown in Fig. 7. There is a tendency towards lower REE concentrations and stronger

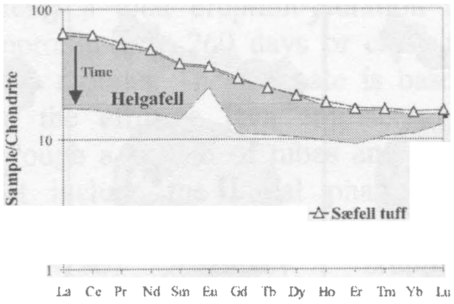

Fig. 7. REE patterns for the Helgafell tephra and lavas compared with the neighbouring Sæfell tuff. Sæfell data from Mattsson and Oskarsson, in prep. All values are normalised to C1 chondrite values of Sun and McDonough (1989).

Europium-anomaly (Eu/Eu* = 1.08 – 2.00) with time during the eruption. The complete data set used for Fig. 5 and 6 can be provided by the corresponding author upon request.

Discussion

Tephra deposition

The tephra found at locations T1 and T2 has a REE-composition that closely resembles the Sæfell tuff (Fig. 7). However, we find by the stratigraphic mapping that the tephra cannot be attributed to any phase in the Sæfell eruption but must have been deposited during the Helgafell eruption. This conclusion is based on several lines of evidence. First, the thickness of the soil profile (32 cm) located in between the lowermost tephra (HT01) and the eroded surface of the Sæfell tuff ring implies that the time required to allow such extensive erosion of the tuff and soil production exceeds the expected duration of the Sæfell eruption. Secondly, the morphology of individual tephra grains is characteristic for Strombolian and not phreatomagmatic activity. Finally, there is a smooth transition from virtually undisturbed tephra sequences to the capping lava flows, indicating a short time span from tephra deposition to lava emplacement.

Lava emplacement

The thin but widespread, vesicular lava flows occurring on the eastern side of the lava field are interpreted as fountain-fed flows. They formed in an early phase of the eruption when the effusion rate was high and large amounts of spatter accumulated near the vent, resulting in rheomorphic flows (Fig. 8b). The abundant volcanic bombs found as rotated inclusions in the same flows further strengthen this hypothesis. The fountain-fed flows are only found on the eastern side of the lava field closest to the vent. Similar flows were probably emplaced in other directions as well, with a radius of approximately 1 km, but have subsequently been buried by lava flows as the eruption progressed (Fig. 8c,d). When the effusion rate decreased, activity at the crater shifted from Hawaiian to Strombolian. Degassing of the lava in the Helgafell crater and vent, combined with cooling resulted in rapid growth of the already nucleated olivine and plagioclase. Lava viscosity increased significantly in response to phenocryst growth. Lava started to flow in open channels, initially west of the crater and later towards north, extending to a distance of approximately 200 meters from the vent. Cooling of lava transported in channels favours overgrowth and development of tubes (Greeley 1971; Swanson 1973; Hon et al. 1994), which in the Helgafell case occurred approximately 200 m from the vent. Lava transported in tubes is allowed longer time to degas during transport and commonly forms P-type pahoehoe (Wilmoth and Walker

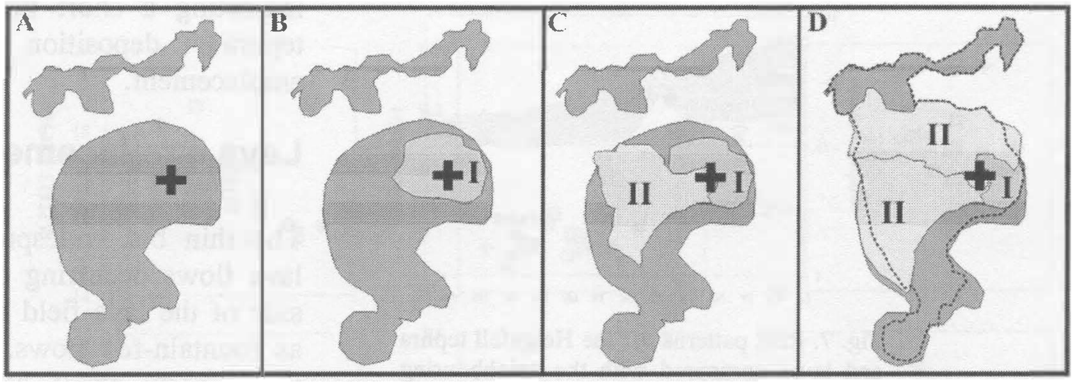

Fig. 8. Schematical sketch of lava emplacement. + marks the location of the vent. (A) eruption occurs on the northern flank of the pre-existing Sæfell tuff cone, (B) fountain-fed flows (I) are emplaced near the crater, (C) P-type pahoehoe (II) flows to the west of the crater, and (D) flows are deflected towards north adjoining pre-existing islands. Dashed lines represent the approximate outline of Heimaey prior to the 1973 Eldfell eruption.

1993). The overall characteristics of P-type pahoehoe with a thin lower vesicular zone (~ 1 dm), a dense interior with few vesicles and an upper vesicular zone are found in the western side of the Helgafell lava field. Due to the proximity of the coastline to the tumuli areas, it is likely that rapid cooling of the flow front, when entering the sea caused the inflation features observed in the Helgafell lava field as the lava supply rate exceeded flow front displacement. The lava emplaced in the eruption has been calculated to have an approximate volume of 0.65 km^3 (0.6 km^3 DRE). The two small crater overflows observed on the Helgafell cone suggests that the crater, at a late stage of the eruption, was filled with lava and oscillating. The suggestion by Cameron et al. (1992) that the vent either drained out or evacuated the lava in explosive activity is contradicted by the field observations. We find that the gravity low measured over Helgafell is most likely a result of the stratigraphic relations, e.g. a scoria cone deposited upon a tuff ring. Whereas, further to the sides, thick lava flows cover the tuff ring, and would result in the higher gravities measured.

Eruption duration

An estimate of the total eruption duration can be made by calculating the time of inflation of flow-lobe tumuli combined with stratigraphic mapping. Inflation of flow-lobes does not occur when lava supply stops, hence the maximum time calculated for inflation must correspond to the minimum eruption duration. Tumuli are surface features and only reflect inflation of the uppermost flows emplaced at a certain location. The stratigraphic mapping clearly shows that the lava flows forming pressure field C (Fig. 1b), were emplaced prior to those in pressure field A. By adding calculated values for active uplift in fields A (82 days) and C (44 days), a total eruption duration of 126 days can be estimated. However, at least two more generations of tumuli were found at lower stratigraphic levels in the exposed coastal sections. Since no measurements of inflation cleft depths in these tumuli were possible, their sizes (width/height) served as an indicator of the time of inflation. Adding the calculated values of inflation to the estimated values suggests that the total duration of the Helgafell eruption was at least twice the above calculated value, which yields a total eruption duration of approximately 260 days or close to nine months. The estimate is based on the effusive lava emplacement through a system of tubes and does not include the initial phase that produced the spatter-fed flows. The 1963-1967 Surtsey eruption produced 0.8 km^3 of lava equivalent in 42 months (Jakobsson et al. 1998) and the Eldfell eruption in 1973 produced 0.25 km^3 lava in five months (Jakobsson 1979). Calculating the volume-time ratio for these two eruptions yields a result of 0.02 and 0.04 repectively. The Helgafell eruption (nine months producing 0.6 km^3 lava) has a volume-time ratio of 0.06. The calculated value for Helgafell seems realistic when compared to the observed values for Eldfell and Surtsey, considering that it is an estimate of the minimum duration and any increase in time will lower the volume-time ratio.

Geochemistry

The observed scatter in the major element variation diagrams (Fig. 6) are not consistent with simple fractionation of a parental magma, but is partially overprinted by processes such as incorporation of plagioclase xenocrysts and late stage segregation resulting in considerable scatter in the variation diagrams.

Fractional crystallisation

The abundant normally zoned plagioclase phenocrysts with continuous core compositions ranging from An_{68} to An_{64} present in the Helgafell lava suggests that the Helgafell magma was residing and fractionating for some time in a magma chamber prior to the eruption. The fractionation was mainly controlled by olivine, as indicated by the strong covariance of MgO and Ni (Fig. 6). Fractionation of olivine also explains the positive correlation of MgO with FeO and negative correlation with SiO_2.

Incorporation of plagioclase xenocrysts

Large xenocrysts of plagioclase were incorporated into the Helgafell magma in various amounts during the eruption, creating dilution effects in Ti and Fe, which are manifested as scatter in the variation diagrams for these elements. The incorporation probably occurred during ascent of the magma leaving the xenocrysts little time to equilibrate with the magma, as suggested by the narrow reaction rim (5 – 10 μm) present on the xenocrysts and the absence of resorption textures. In the variation diagrams the presence of the more calcic plagioclase xenocrysts (An_{71}) can be strongly inferred from low Na_2O and high CaO and Sr in some of the samples, creating considerable scatter from the main trend. Incorporation of plagioclase xenocrysts may also be inferred from the increasing size of Eu/Eu* anomaly with time in the REE plot (Fig. 7).

Late stage segregation

Segregation occurring at a late stage during the solidification of the flows is clearly displayed in the groundmass-rich samples, where titanomagnetite has accumulated as indicated by anomalously high Ti and V levels. Fractionation of alkali basalts due to filter pressing of a late crystallising liquid generates more evolved chemical compositions, closely resembling the FeTi-rich basalts of the EVZ found north of Vestmannaeyjar (Sigmarsson et al.

2002). Such segregation features have previously been reported from Heimaey occurring in the Helgafell lava and from the Klífið intrusion (Jakobsson 1968, 1979). The volume of segregated material is minor compared to the volume of the lobes in which they occur (2 - 3 % in Surtsey; Sigmarsson et al. 2002) resulting in a significant increase in Ti and V, without noticeably affecting the FeO content of the flows.

Conclusions

The geochemistry of the Helgafell extrusives indicates that fractionation of olivine has dominated the compositional evolution, although the fractionation trend is partially overprinted by processes such as incorporation of plagioclase xenocrysts and late stage segregation occurring in the flows. The Helgafell eruption started with Hawaiian fire-fountain activity, depositing tephra and spatter close to the vent. The initially emplaced fountain-fed clastogenic lava flows are confined to a radius of approximately one kilometre from the vent. Forming rather thin, vesicular flows with large lateral extent often containing numerous volcanic bombs. When the effusion rate decreased activity at the crater shifted from Hawaiian to Strombolian and the lava started to flow in channels close to the vent. Lava was in the beginning flowing west of the crater. At approximately 200 m from the vent, cooling of the lava stream favoured overgrowth, and the lava was from this point supplied to the active flow front through a system of tubes inside the lava field. The change in activity at the crater, from Hawaiian activity with a continuous up-rush of lava to Strombolian characterized by discrete explosions, allowed the lava more time to degas and cool in the Helgafell crater causing rapid growth of already nucleated crystals of olivine and plagioclase. The viscosity of the lava increased drastically in response these factors, resulting in clogging of lava tubes, and formation of characteristic inflation features such as tumuli. Later in the eruption lava flows were deflected from west towards north, probably due to the topographic barrier created by previously emplaced lava and tumuli areas. The depth of inflation cracks in tumuli and the observed lava stratigraphy constrain the duration of the Helgafell eruption to a minimum of nine months. This result is in agreement with the duration-volume relations of the Surtsey and Eldfell eruptions. Towards the end of the eruption two small crater overflows were emplaced on the slopes of the scoria cone, indicating that the crater was filled with lava and oscillating. No more volcanic activity occurred on Heimaey for almost 6000 years.

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