Bull Volcanol (2007) 70:139–156
DOI 10.1007/s00445-007-0125-6

RESEARCH ARTICLE

Morphologies and emplacement mechanisms of the lava flows of the Faroe Islands Basalt Group, Faroe Islands, NE Atlantic Ocean

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Received: 7 February 2006 / Accepted: 29 January 2007 / Published online: 14 March 2007

Abstract The Palaeogene Faroe Islands Basalt Group (FIBG) comprises three eruptive sequences or formations, all emplaced into a subaerial environment during the development of the extensive continental flood basalt province that stretches from East Greenland through the Faroe Islands and into the Faroe-Shetland Basin. The Beinisvørð Formation, having a tabular-classic facies architecture, is composed of a sequence of simple flows each comprising a single sheet lobe. The Beinisvørð Formation is overlain by the distinctly contrasting Malinstindur Formation that has a compound-braided facies architecture. The Enni Formation occurs at the top of the sequence and consists of a mixture of simple and compound flows with tabular-classic and compound-braided facies architectures, respectively. Surface and internal characteristics of the sheet lobes of the Beinisvørð and Enni formations indicate emplacement through inflation, which is more obvious for the tube-fed compound flows of the Malinstindur and Enni formations. The difference between the simple and compound flow sequences of the FIBG is, most likely, linked to the manner in which the lava was supplied during the eruption and the eruptive style of the volcanic system. The sheet lobes were erupted over laterally extensive areas from fissure systems which had a continuous supply of lava, which contrasts with the tube-fed compound flows which were erupted in a gradual, piecemeal manner from point-sourced, low shield volcanoes with limited areal extents.

Keywords Faroe Islands Basalt Group · Facies architectures · Continental flood basalts · Pahoehoe flows · Inflation

Introduction

The Faroe-Shetland Basin (FSB) is one of several rift basins along the NE Atlantic margin and contains a preponderance of Palaeogene extrusive and intrusive igneous rocks (Fig. 1). This paper attempts to explain the evolution of the volcanic rocks, which crop out on the Faroe Islands, in terms of their morphologies and the overall architecture of the sequence. We also explore some of the wider implications of our observations within the context of eruption mechanisms of continental flood basalts (CFBs).

The Faroe Islands Basalt Group (FIBG) comprises a fragment of an extensive CFB province that is also exposed in East Greenland (Saunders et al. 1997; Larsen et al. 1999). Prior to the commencement of ocean floor spreading at ~55 Ma, which resulted in the opening of the NE Atlantic, the Faroe Islands and East Greenland were less than 120 km apart, based upon plate reconstructions and likely geochemical correlations between the two originally contiguous sequences (Fig. 1; Larsen et al. 1999). Apart from the classic study of Rasmussen and Noe-Nygaard (1969, 1970), which was based on field observations made during the period 1940–1970, most recent studies of the onshore FIBG have concentrated on the geochemistry of the lavas (Gariépy et al. 1983; Hald and Waagstein 1983;

Editorial responsibility: A Harris

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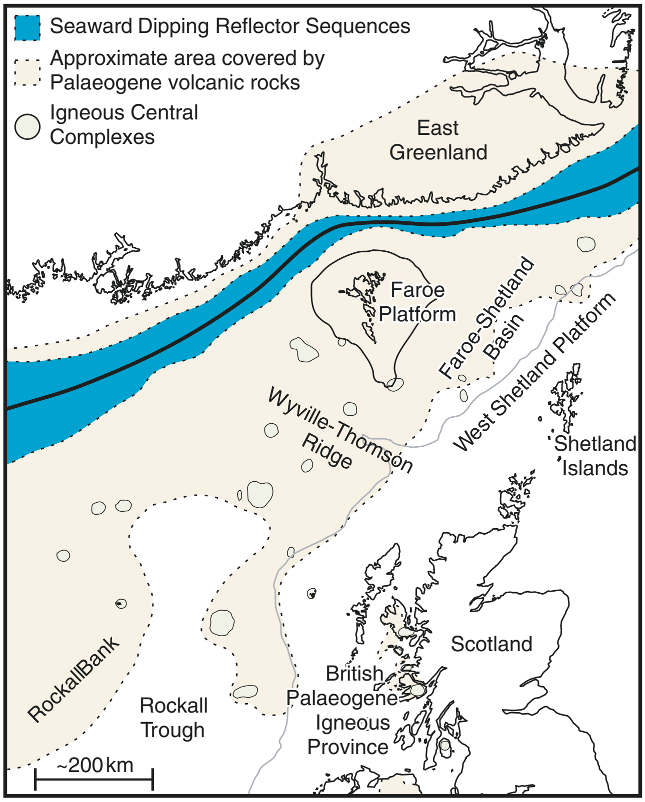


Fig. 1 Simplified map showing the location of the Faroe Islands within the North Atlantic Igneous Province (NAIP) at ~55 Ma. Based on Saunders et al. (1997); Larsen et al. (1999) and Ritchie et al. (1999)

Waagstein 1988) and their geochronology (Riisager et al. 2002; Waagstein et al. 2002). However, this somewhat neglected part of the North Atlantic Igneous Province (NAIP) preserves a spectacular record of eruption mechanisms through its various lava flow morphologies, and provides data concerning the architecture and growth of CFBs, in general.

Geological setting of the Faroe Islands Basalt Group (FIBG)

The Faroe Islands comprise 18 islands, most of which are elongated on a NW–SE trend and are separated by long, narrow fjords (Fig. 2). The archipelago is located ~280 km NW of Scotland and ~400 km SE of Iceland in the NE Atlantic Ocean. The islands cover an area of ~1400 km^2 and extend over a distance of ~115 km N–S by ~75 km E–W. The landscape of the Faroe Islands was sculpted by glacial action during the Quaternary Sub-period, producing a dissected mountainous terrain, with Slættaratindur on Eysturoy, at a height of 882 m, the highest peak.

The islands and its insular shelf (to water depths of ~150–200 m) form the Faroe Platform, which has a roughly triangular shape and is a remnant of an extensive, predominantly subaerial, CFB province of the Palaeogene NAIP (Fig. 1; Waagstein 1988; Ellis et al. 2002). The Faroe Platform is part of the NE Atlantic margin, a passive continental margin that extends from the western Barents Sea to offshore west of Ireland. The arrival of the putative proto-Iceland plume beneath Greenland led to widespread volcanic activity during the Palaeocene and early Eocene (62–54 Ma) and resulted in the eventual continental break-up between Greenland and Eurasia, culminating with the onset of seafloor spreading in magnetochron 24r (~55 Ma) (Saunders et al. 1997; Ritchie et al. 1999; Jolley and Bell 2002).

The FIBG extends from the Faroe Islands into the Faroe-Shetland Basin for at least 200 km in an easterly to south-easterly direction (Fig. 1; Naylor et al. 1999; Ritchie et al. 1999; Ellis et al. 2002). In some distal areas, the FIBG is characterised by prograding volcanic deltas, where subaerial lavas have flowed into substantial bodies of water and auto-brecciated to form hyaloclastite deposits (Naylor et al. 1999; Ellis et al. 2002; Passey 2004). To the west and north of the Faroe Islands, seaward dipping reflector sequences (SDRS) mark the transitional zone between the attenuated continental Faroe Platform and oceanic crust of the Iceland and Norway basins (Smythe 1983; Parson and the ODP Leg 104 Scientific Party 1988; Ritchie et al. 1999). It is inferred from geochemistry (Gariépy et al. 1983; Hald and Waagstein 1983; Holm et al. 2001) and geophysics (Casten

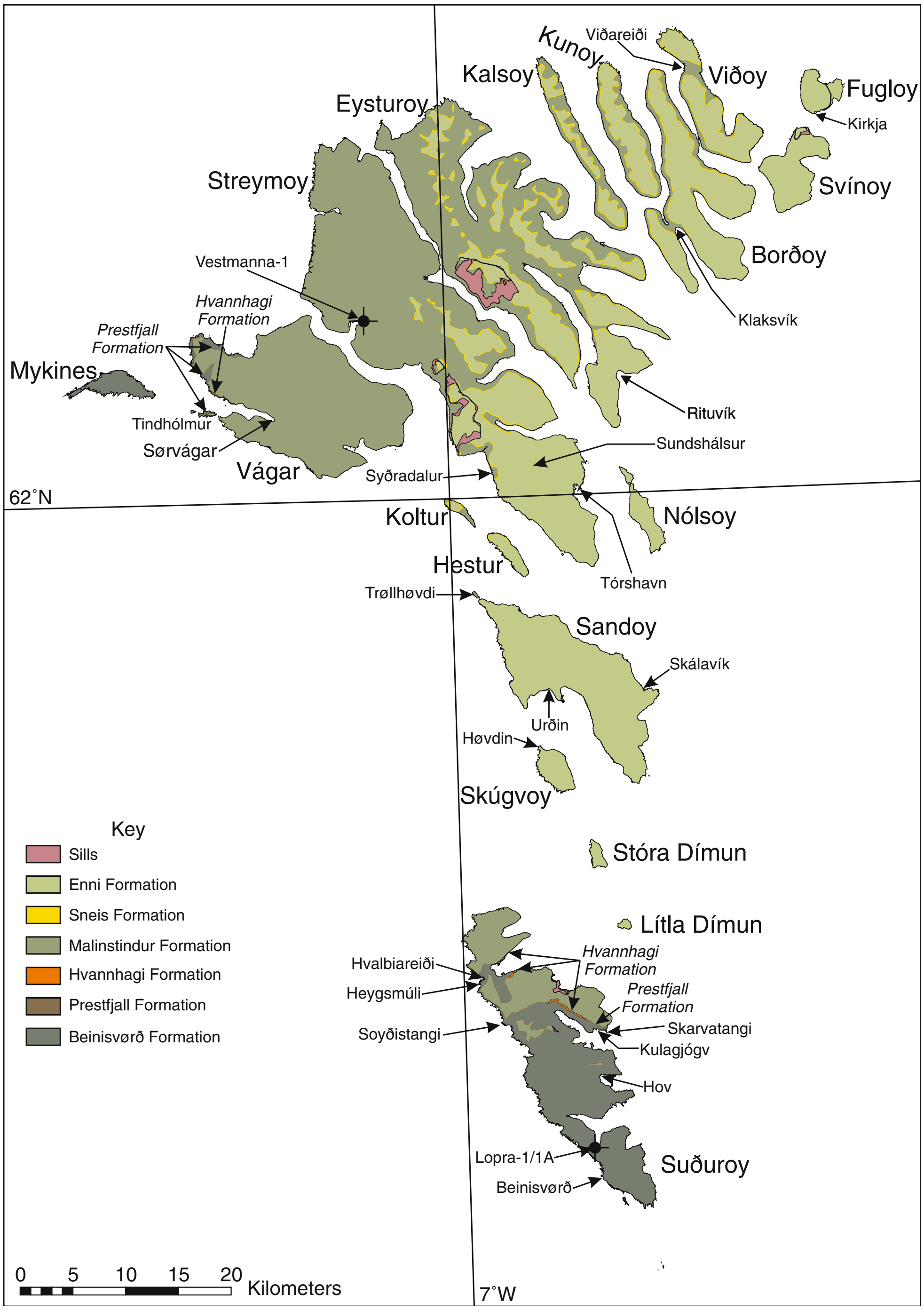


Fig. 2 Simplified geological map of the Faroe Islands, showing the main subdivisions of the onshore FIBG. Note outcrops of the Prestfjall and Hvannhagi formations are also indicated on map face

1973; Bott et al. 1974; Richardson et al. 1998, 1999) that the FIBG on the Faroe Islands overlies continental, Precambrian metamorphic basement.

Stratigraphy of the onshore FIBG

On the Faroe Islands, the FIBG has a gross stratigraphic thickness of at least 6.6 km and is dominated by tholeiitic basalt lava flows, subdivided into seven formations based primarily on lithology (Rasmussen and Noe-Nygaard 1969, 1970; Passey et al. 2006), but also on geochemistry (Fig. 3; Waagstein 1988). The Beinisvørð Formation (BF), dominated by thick, laterally extensive, aphyric sheet lobes, is ~3,300 m thick, with ~900 m exposed on the Faroe Islands and an additional ~2,400 m encountered in the onshore well Lopra-1/1A on Suðuroy (Rasmussen and Noe-Nygaard 1970; Hald and Waagstein 1984; Waagstein 1988; Ellis et al. 2002; Chalmers and Waagstein 2006). Below the BF flows in the Lopra-1/1A well are various volcaniclastic

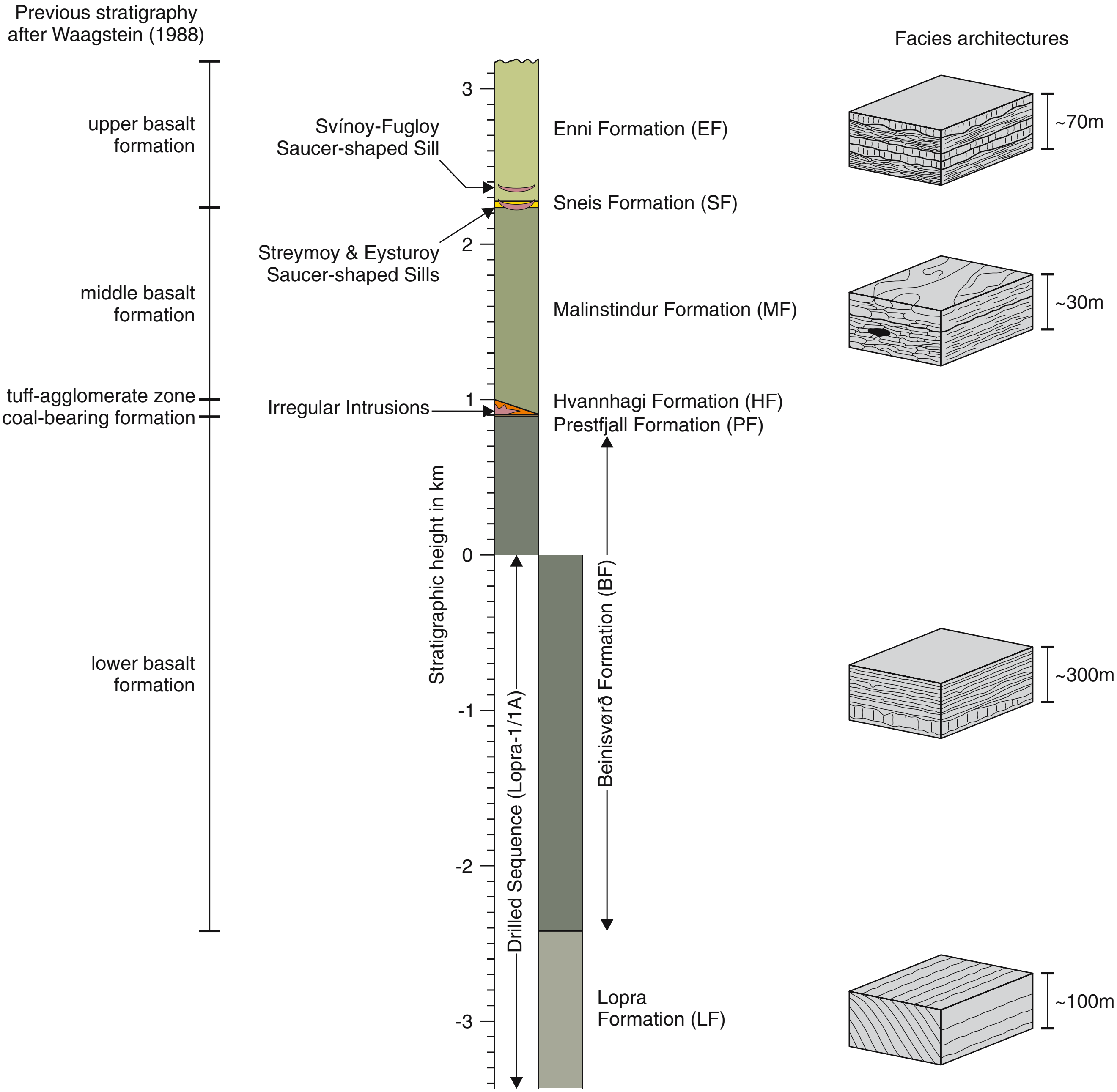


Fig. 3 Stratigraphic column for the onshore FIBG. The previous stratigraphy is given to the *left of the column* for comparison. Facies architectures are given to the *right of the column*

lithologies and sills of the ~1,000 m thick Lopra Formation (Fig. 3; Ellis et al. 2002).

The BF is overlain by the 3–15 m thick Prestfjall Formation (PF), consisting of coals, together with volcaniclastic mudstones, sandstones and conglomerates of swamp, lacustrine and fluvial association (Rasmussen and Noe-Nygaard 1970; Lund 1983, 1989), which represent a significant hiatus in the volcanic activity. Volcanism resumed with the deposition of the ~50 m thick Hvannhagi Formation (HF), composed of interbedded basaltic tuffs and volcaniclastic sedimentary lithologies (Passey 2004).

The ~1,350 m thick Malinstindur Formation (MF) marks a return to the eruption of subaerial lavas, predominantly compound flows, which evolve, up sequence, from olivine-phyric, through aphyric to plagioclase-phyric basalts (Noe-Nygaard and Rasmussen 1968; Rasmussen and Noe-Nygaard 1970; Waagstein 1988). The MF is overlain by the laterally extensive sandstone-conglomerate sequence, 15–25 m thick, of the Sneis Formation (SF). The overlying Enni Formation (EF) is composed of interbedded simple and compound flows. The EF is at least 900 m thick, and comprises plagioclase-phyric lavas, together with aphyric to olivine-phyric lava flows of MORB-like composition (Noe-Nygaard and Rasmussen 1968; Rasmussen and Noe-Nygaard 1970; Waagstein 1988). The amount of section missing due to post-volcanism erosion is poorly constrained, but considered to be at least a few hundred metres based upon the nature and abundance of amygdale mineral infills (Waagstein et al. 2002).

Larsen et al. (1999) demonstrated, through geochemistry, that the Beinisvørð Formation of the Faroe Islands and the Nansen Fjord Formation, East Greenland, form a pre-break-up succession covering an estimated area of ~70,000 km^2. In contrast to the ~3,300 m thickness of the BF, the Nansen Fjord Formation has a thickness of <1,100 m, which suggested to Larsen et al. (1999) that this pre-break-up succession was erupted into an area of evolving basins with the depocentre situated in the Faroe-Shetland region, and that East Greenland represents the northern margin of the basin. The Malinstindur and Enni formations, along with the Milne Land Formation, East Greenland, form a syn-break-up volcanic succession with a pre-rift extent of ~220,000 km^2 (Larsen et al. 1999). Volcanism continued in East Greenland with the eruption of an additional ~3–3.5 km thick succession of lavas (1–1.5 km now eroded), the remnants of which are represented by the Geikie Plateau and Rømer Fjord formations (Larsen et al. 1999).

The contraction of the putative proto-Iceland plume in early Eocene times (Saunders et al. 1997) resulted in subsidence in the Faroe-Shetland region (Ellis et al. 2002) and the general easterly dip of the FIBG (Andersen 1988). Subsequent Cenozoic compression along the Faroe-Iceland Ridge led to the present elevation of the Faroe Islands and a number of inversion structures offshore, resulting in the FIBG occurring at the seafloor, for example, at the Wyville-Thomson and Fugloy ridges (Doré and Lundin 1996; Doré et al. 1999). Elsewhere in the Faroe-Shetland Basin, post-Eocene strata have buried the basaltic lavas (Sørensen 2003).

Terminology

For simplicity we have adopted the terminology proposed by Self et al. (1997) for CFBs. A *flow lobe* is used to describe an individual package of lava surrounded by a chilled crust and a *lava flow* is the product of a single, more or less continuous outpouring of lava that can be composed of one or more flow lobes. Individual lava flows are typically separated by weathering surfaces and/or clastic lithologies. If a lava flow consists of a single flow lobe it is referred to as a *simple lava flow* (Walker 1972) and if the lobe has a sheet-like or tabular geometry it is classified as a *sheet lobe*. Conversely, a *compound lava flow* (Walker 1972) is made up of two or more flow lobes of any geometry or size. A *flow field* is the aggregate product of a single eruption or vent and is built up of one or more lava flows and is usually identified on the basis of the geochemistry of the constituent flows.

Emplacement of continental flood basalts

Examination of the morphologies of lava flow lobes in CFBs has shown that their field characteristics can aid our understanding of the mechanisms involved in their emplacement (Self et al. 1997). Jerram (2002), using the terminology of Walker (1972, 1973), discussed two end-member types of facies architecture for subaerial basalt lava flows. First, the tabular-classic facies architecture, which comprises a simple flow, typically with a sheet geometry, separated by palaeosols and/or various terrestrial clastic lithologies (Walker 1972; Jerram 2002). These simple flows have been interpreted to represent lavas that were emplaced rapidly during extremely high effusion rate eruptions (Shaw and Swanson 1970; Walker 1972, 1973). This contrasts with the second end-member, of compound-braided facies architecture, comprising a compound flow, which consists of numerous anastomosing flow lobes of pahoehoe that were emplaced at lower effusion rates (Walker 1972, 1973; Jerram 2002).

However, over the past decade, work on what were previously thought to be turbulently emplaced simple flows of the Columbia River Basalt Group (CRBG) suggests that they are actually compound in nature and are composed of inflated pahoehoe flow lobes that were emplaced passively (i.e. non-turbulently) at lower effusion rates (Self et al. 1996; Thordarson and Self 1998). Although volumes, flow velocities and effusion rates differ by several orders of magnitude, the morphologies of the CRBG inflated flow lobes are similar to those observed for the inflated pahoehoe sheet lobes of Hawai`i (Self et al. 1996). Subsequently, similar inflated pahoehoe flow lobes have been recognised in other CFB provinces, for example, the Mull Lava Field of the NAIP (Kent et al. 1998) and the Paraná-Etendeka Province (Jerram et al. 2000). Bondre et al. (2004) have also documented inflated pahoehoe flow lobes from the Deccan Volcanic Province.

Facies architecture of the lava flows

The various facies architectures of the three lava-dominated formations, Beinisvørð, Malinstindur and Enni, are illustrated in Fig. 3. The BF and sections of the EF comprise laterally extensive, sheet lobes, many with well-developed palaeosols or red-orange volcaniclastic interlava lithologies that highlight their typically planar upper surfaces. This layer-cake appearance and associated weathering-induced

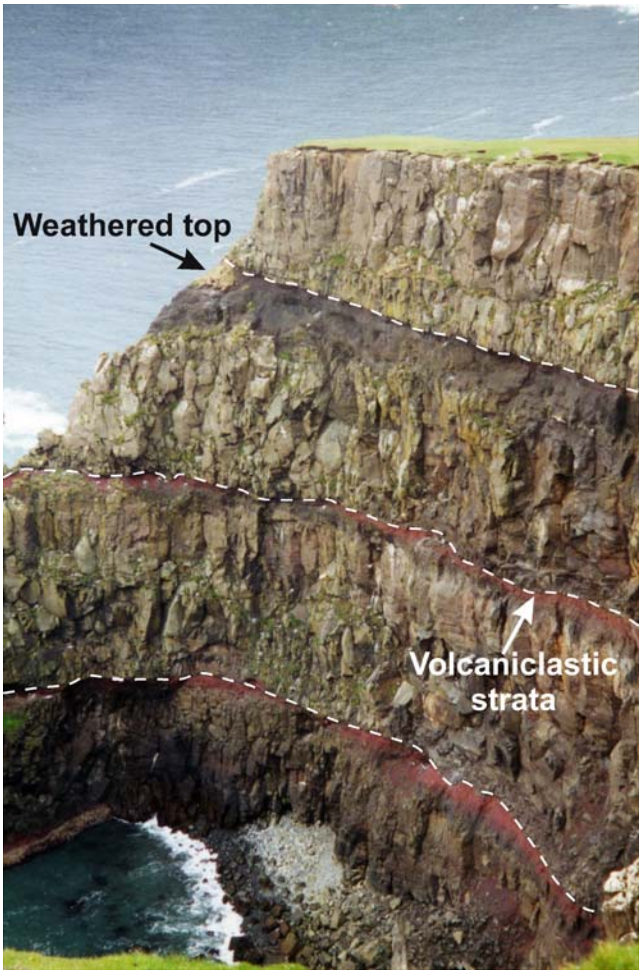


Fig. 4 Four simple flows belonging to the Beinisvørð Formation at Heygsmúli, SW of Hvalba, north Suðuroy. Each sheet lobe has poorly-developed columnar jointing, a reddened (oxidised) top or bole, and is overlain by a thin sequence of reddish-orange volcaniclastic strata, typically of silt to sand grade. Height of cliff is ~80 m

terraced terrain (step or trap topography) is characteristic of lavas with a tabular-classic facies architecture (Jerram 2002) and, therefore, the sheet lobes from the two formations shall be described collectively. The MF and sections of the EF are dominated by subaerial compound lava flows composed of numerous thin, overlapping and anastomosing flow lobes typical of the compound-braided facies architecture (Jerram 2002). These flow lobes do not appear to be laterally extensive and have more meander-like geometries, giving rise to a less organised, though vaguely terraced, topography.

Flows with tabular-classic facies architecture

Thicknesses and flow lengths

Individual flow lobes of the BF have a sheet (cf. tabular) geometry and can be traced across Suðuroy for a distance of at least 9 km (Fig. 4). The Beinisvørð cliff section, ~3 km NW of Sumba, Suðuroy, comprises at least nineteen BF sheet lobes with an aggregate thickness of 469 m, indicating an average sheet lobe thickness of ~25 m. This is comparable to an average sheet lobe thickness of ~20 m obtained for the entire exposed and drilled (Lopra 1/1A well) BF succession (Waagstein et al. 1984). However, individual sheet lobes within the Beinisvørð cliff section range in thickness from 10 to 70 m.

The average sheet lobe thicknesses for those of the EF is ~8–11 m (Rasmussen and Noe-Nygaard 1970), for example, the Kirkja Flow on Fugloy is ~8 m thick (Fig. 5), and therefore less than half the average thickness of the sheet lobes of the BF. The thickest sheet lobe within the EF is ~30 m, cropping out along the harbour wall in Tórshavn, Streymoy. Certain sections of EF sheet lobes, for example on Viðoy, indicate flow lengths of at least 3 km.

Although it is difficult to determine sheet lobe aspect ratios in detail (average thickness/lateral extent) from the dominantly two-dimensional cliff exposures, it is clear that values <0.005 are common. Except for a few valley-confined lavas, flow margins have not been observed within the Beinisvørð or Enni formations, so the aspect ratios are actually lower than reported here.

In summary, the sheet lobes are clearly separated from subjacent lobes by weathering surfaces, palaeosols or volcaniclastic lithologies over several kilometres. Even though exposures are limited, these sheet lobes are interpreted as simple flows with a tabular-classic facies architecture. Applying the terminology of Self et al. (1997), each simple flow constitutes a flow field as each flow represents a single eruption. However, it should be noted

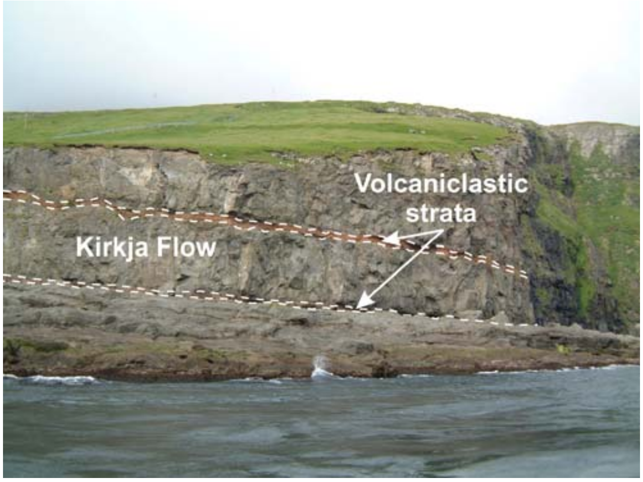


Fig. 5 The Kirkja Flow, SW Fugloy. This ~8 m thick plagioclase-phyric sheet lobe of the Enni Formation overlies a compound pahoehoe flow capped by a thin (~70 cm) sequence of reddish-brown sandstones on the coastal platform, and is, itself, overlain by a further ~1 m of reddish-brown sandstones and siltstones and an upper Enni Formation lava sheet lobe of similar character

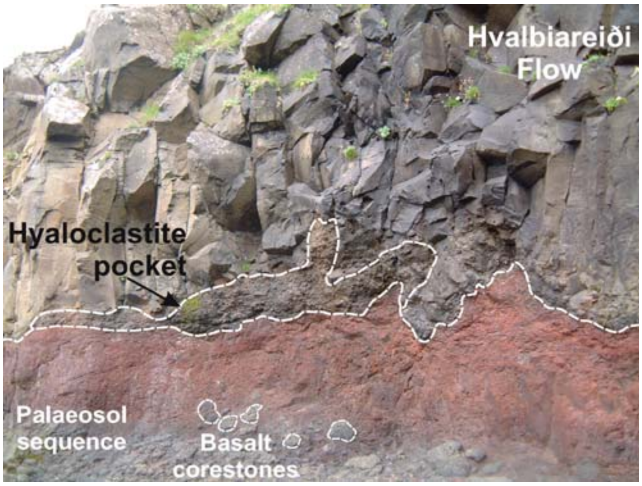


Fig. 6 A hyaloclastite pocket at the base of the Hvalbiareiði Flow, Beinisvørð Formation, Hvalbiareiði, Suðuroy. Isolated pockets of hyaloclastite with sharp basal contacts overlie a ~1.5 m thick palaeosol sequence. The blocky, almost jigsaw-fit, clasts of basalt in the hyaloclastite pocket are compositionally and texturally the same as the basalt from the overlying sheet lobe. Height of section is ~30 m

the sheet lobes of the entire ~3 km thickness of the BF have a uniform geochemistry (Waagstein 1988; Larsen et al. 1999), implying they have a common source.

Flow base and top characteristics

Sheet lobes in the BF and EF are typically massive, with sharp and planar bases and rubbly, amygdaloidal tops. In a few instances in the BF, a basal crust is present and is characterised by an amygdaloidal zone up to 50 cm thick. These lower crusts contain up to 20 vol% amygdales, which are generally spherical and have average diameters of a few millimetres. Pipe amygdales within the lower crusts have not been observed. Extensive basal rubbly zones are also generally lacking, although isolated pockets of hyaloclastite and blocky peperite occur in BF sheet lobes where they overlie interlava sedimentary units. For example, the base of the ~15 m thick Hvalbiareiði Flow contains isolated pockets of hyaloclastite with sharp basal contacts where it overlies a palaeosol sequence (Fig. 6). This contrasts with the base of the ~17 m thick Skarvatangi Flow, which overlies a sequence of fluvial volcaniclastic conglomerates, sandstones, siltstones and mudstones. Here, instead of a basal hyaloclastite facies, there are small isolated pockets of blocky peperite consisting of angular clasts of basalt derived from the sheet lobe, some with a jigsaw-fit texture, within a matrix of volcaniclastic siltstone derived from the subjacent sedimentary sequence (Fig. 7).

Hyaloclastite and peperite facies or pockets have not been observed from the bases of EF sheet lobes. However, empty, near-horizontal, cylindrical structures occur at the bases of numerous EF sheet lobes (Fig. 8). These structures are interpreted to be tree moulds. One example occurs at the base of a sheet lobe in a road-cut NW of Tórshavn, Streymoy. This EF sheet lobe has a sharp and hummocky basal contact and overlies a ~1.2 m thick sequence of fluvial volcaniclastic sandstones and siltstones. The tree mould has a diameter of ~35 cm, extends into the sheet lobe for ~75 cm and is lined with carbonaceous material.

The massive cores of BF sheet lobes contain <2 vol% amygdales, which have average diameters of a few millimetres. The majority of the amygdales are spherical, although ellipsoidal examples do occur, where they have been stretched and elongated due to shearing. Amygdale sheets and cylinders (cf. Goff 1996) have not been observed.

Rubbly flow tops are commonly several metres thick, some up to 30 m (Ólavsdóttir and Andersen 2003) for the BF sheet lobes, and 2–4 m thick for EF sheet lobes. The rubbly flow tops consist of weathered clasts that are generally amygdaloidal and petrologically similar to the underlying basalt flow lobes. Sabine (1971) and Parra et al. (1987) have shown that the BF rubbly flow tops and the overlying volcaniclastic claystones are the result of contemporaneous subaerial chemical weathering. These clay-dominated weathering profiles commonly show a gradation downward from fine-grained altered material (comminuted basalt) into fresh basalt, as well as retaining relic features of the unaltered sheet lobe and, therefore, can be classified as saprolitic boles (cf. Widdowson et al. 1997). The BF palaeosols analysed by Parra et al. (1987) still retain a high SiO_2 content, placing them in the kaolinitised basalt subfield of the bole field rather than within the extreme alteration laterite field (cf. Schellmann 1986; Widdowson et al. 1997).

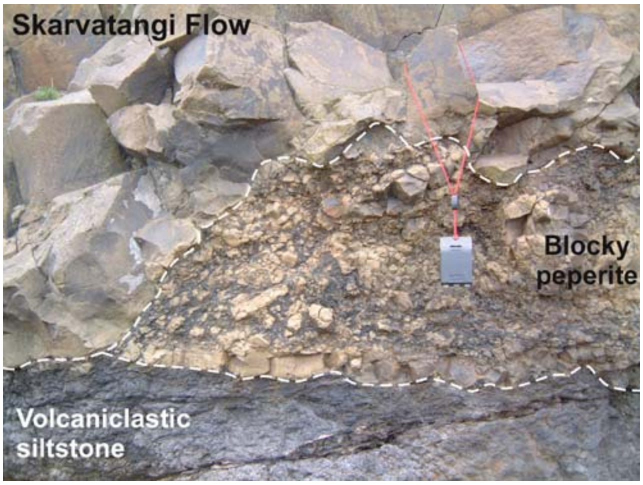


Fig. 7 A pocket of blocky peperite at the base of the Skarvatangi Flow, Beinisvørð Formation, Froðba, Suðuroy, where it overlies a 6–7 m thick sequence of volcaniclastic strata. The angular clasts of basalt were derived from the overlying sheet lobe, some with a jigsaw-fit texture, and are contained within a matrix of volcaniclastic siltstone derived from the subjacent sedimentary sequence. The compass is 10×6 cm

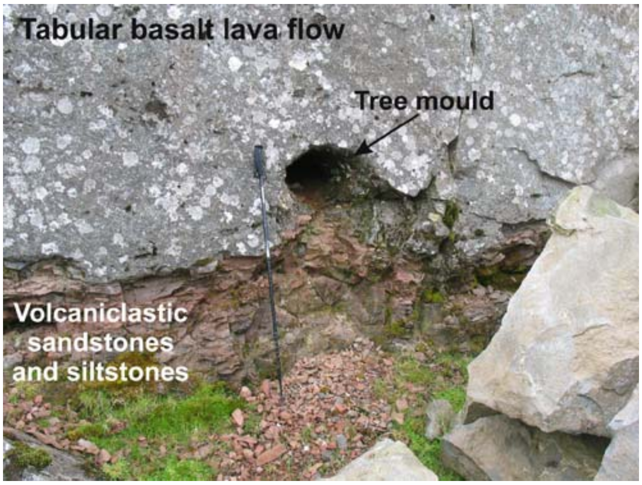


Fig. 8 Mould of a fossil tree trunk or branch at the base of an Enni Formation sheet lobe, ~900 m NW of Sundshálsur, ~5 km NW of Tórshavn, Streymoy. The mould is ~35 cm across, ~75 cm long and is lined with carbonaceous material. The underlying volcaniclastic sandstones and siltstones are at least 1 m thick. Length of pole is ~1.5 m

Hald and Waagstein (1984) demonstrated that the amygdales within individual BF flow tops, encountered in the Lopra-1 well, decrease in abundance and increase in size, downwards from the upper surface. The tops of some of the thicker BF sheet lobes in the well also display amygdale zones (Hald and Waagstein 1984) and this is supported by preliminary studies of core sections from the Hov-Øravík tunnel project on Suðuroy (Ólavsdóttir and Andersen 2003; Ólavsdóttir 2004). For example, a ~24 m thick sheet lobe from Borehole 8 (61°31′30 N, 006°47′34 W), which overlies a ~1.68 m thick reddened volcaniclastic mudstone, can be separated into three parts: a lower crust (1.05 m), a massive core (20.9 m), and an upper crust (2.18 m). The upper crust exhibits three amygdale zones (88, 62 and 68 cm thick, respectively), each of which shows a downward decrease in the abundance of amygdales (10–15% down to 2%) and an increase in maximum amygdale size (2–6 mm). The upper 1.84 m of the massive core contains <5 vol% of large (60×5 mm), deformed amygdales and the rest of the core is relatively non-amygdaloidal (<1–2 vol%). The upper ~2 m of the sheet lobe is rubbly and grades into a ~1.31 m thick reddened volcaniclastic claystone that contains rare rounded clasts similar in petrology to the underlying sheet lobe.

Columnar jointing

Poorly-developed columnar jointed sheet lobes are common throughout the BF and EF and typically occur in sheet lobes overlying weathered flow tops (palaeosols). However, within the uppermost 100–200 m of the BF, classic well-developed columnar jointed sheet lobes occur, for example, the Kulagjógv Flow, which crops out east of Froðba, Suðuroy (Fig. 9). The thickness of this sheet lobe varies along its length, but is on average 20 m. At the eastern extent of the exposure, the Kulagjógv Flow displays superb fanning arrays of columns due to the ponding of the lobe within a channel, which can be used to recognise relief of ~20 m on the underlying palaeo-land surface. The inland exposure of the lobe consists of very regular columns with a uniform width of ~2 m. The near-planar surfaces of the joint faces exhibit chisel marks, which alternate from smooth to rough portions over a distance of 8–10 cm. This alternation is considered to indicate the direction in which the column surface propagated (DeGraff and Aydin 1987) and is the product of cycles of stress build-up and release, which produced the fractures that define the columns (Ryan and Sammis 1978).

The occurrence of multi-tiered flow lobes is extremely rare within the BF and EF, but one sheet lobe from the BF in a road-cut NW of Hov, Suðuroy, can quite clearly be divided into colonnade and entablature tiers (Fig. 10). The base and top of the Hov Flow are not observed, but this sheet lobe is at least 20 m thick. The colonnade tier comprises the lower 8–10 m of the exposed section and is composed of well-developed columns that have a uniform width of ~1.8 m. There is a sharp and planar contact between the colonnade and the overlying 8–10 m thick entablature tier. The entablature consists of 20–30 cm wide, wavy or curvi-columnar (hackly) columns. It is unclear, due to lack of exposure, whether the entablature is overlain by an upper colonnade, comparable to the multi-tiered flow lobes of the CRBG (Long and Wood 1986).

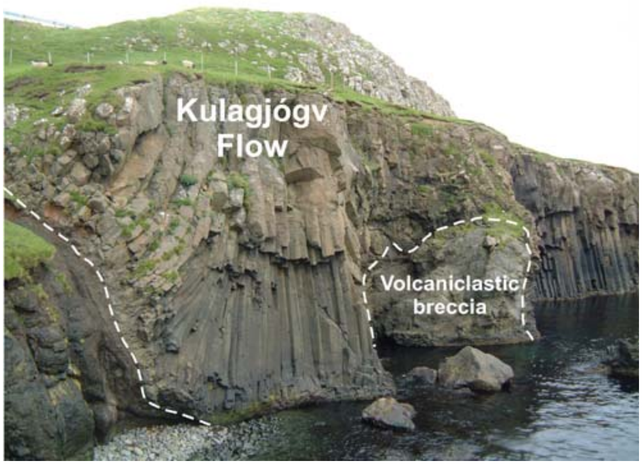


Fig. 9 Fanning arrays of columns within the Kulagjógv Flow, exposed for ~700 m along the coast east of Froðba, Suðuroy. The orientation of the columns can be used to recognise relief of ~20 m on the underlying palaeo-land surface. The underlying, apparently unbedded, volcaniclastic breccia is composed of clasts of amygdaloidal and non-amygdaloidal basalt, which has been differentially eroded and subsequently inundated by the Kulagjógv Flow. Height of cliff is ~30 m

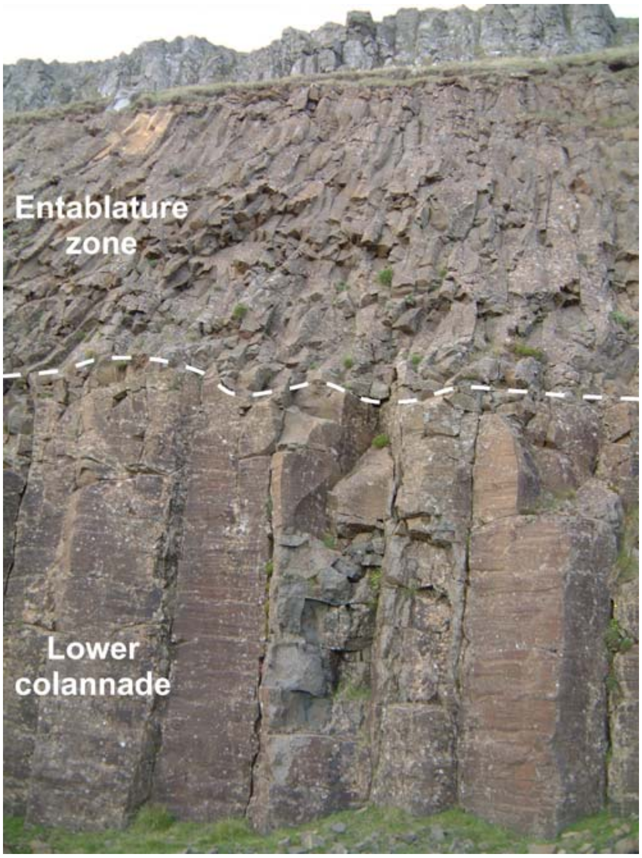


Fig. 10 The multi-tiered Hov Flow, Beinisvørð Formation, road-cut NW of Hov, east Suðuroy. The lower colonnade comprises regular vertical columns and has a sharp, relatively planar, contact with the irregular (wavy) columns of the overlying entablature. Colonnade columns are ~1.8 m across. Note obvious chisel structures on the joint surfaces, which may be attributed to stress-release during cooling and contraction

Flows with compound-braided facies architecture

Thicknesses and flow lengths

The Malinstindur Formation (MF) and sections of the Enni Formation (EF) are composed of subaerial, compound lava flows, on average ~20 m thick, each made up of thinner (typically 0.5–4 m thick) flow lobes (Fig. 11). Individual flows are easily recognised in the lower part of the formation by the presence of minor reddened (oxidised) surfaces, whereas in the upper part of the formation the compound flows are more typically separated by thin interflow sedimentary units.

The flow lobes of a MF compound lava flow at Viðareiði, Viðoy are observed infilling a small channel containing a 1.2 m thick volcaniclastic sequence (Fig. 12). At the margins of the channel the flow lobes dip at 20–30°, representing the relief of the pre-existing topography (Fig. 12b). Within the channel, the flow lobes invade and split the volcaniclastic sequence, causing bifurcation of the strata (Fig. 12c). Some of these flow lobes are <40 cm thick and are noted to terminate within the volcaniclastic strata. The interlava sedimentary strata are poorly lithified and delicate sedimentary structures have been preserved, even within the highly invaded sections.

Flow base and top characteristics

Individual flow lobes have features consistent with being of pahoehoe character (cf. Wentworth and Macdonald 1953; Macdonald 1967, 1972; Cas and Wright 1987; Wilmoth and Walker 1993; Self et al. 1998; Francis and Oppenheimer 2004). Both P-type (pipe amygdale-bearing) and S-type (spongy) pahoehoe flow lobes are recognised, although the P-type lobes are dominant. The P-type lobes can be divided into three distinct zones based on amygdale distribution patterns. For example, a ~1.6 m thick P-type lobe crops out at Viðareiði, Viðoy, comprising a lower crust, a core, and an upper crust (Fig. 12d). The 10 cm thick lower crust is characterised by pipe amygdales with a maximum length of 8 cm that start a few centimetres off the base of the flow lobe. Many of these pipe amygdales are curved, defining the localised flow direction of the lobe (Fig. 12e). The core is a compact, massive zone with irregular jointing that generally lacks amygdale cylinders (cf. Goff 1996). The upper crust is, on average, 25 cm thick and is dominated by ellipsoidal amygdales that have a maximum diameter of 2 cm. A similar pattern is observed for flow lobes encountered in the Vestmanna-1 well (Waagstein and Hald 1984), where the average flow lobe thickness is ~2.2 m, with each lobe divisible into lower,

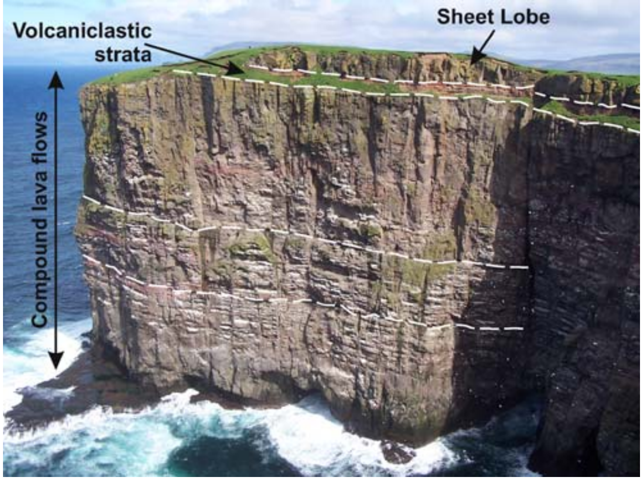
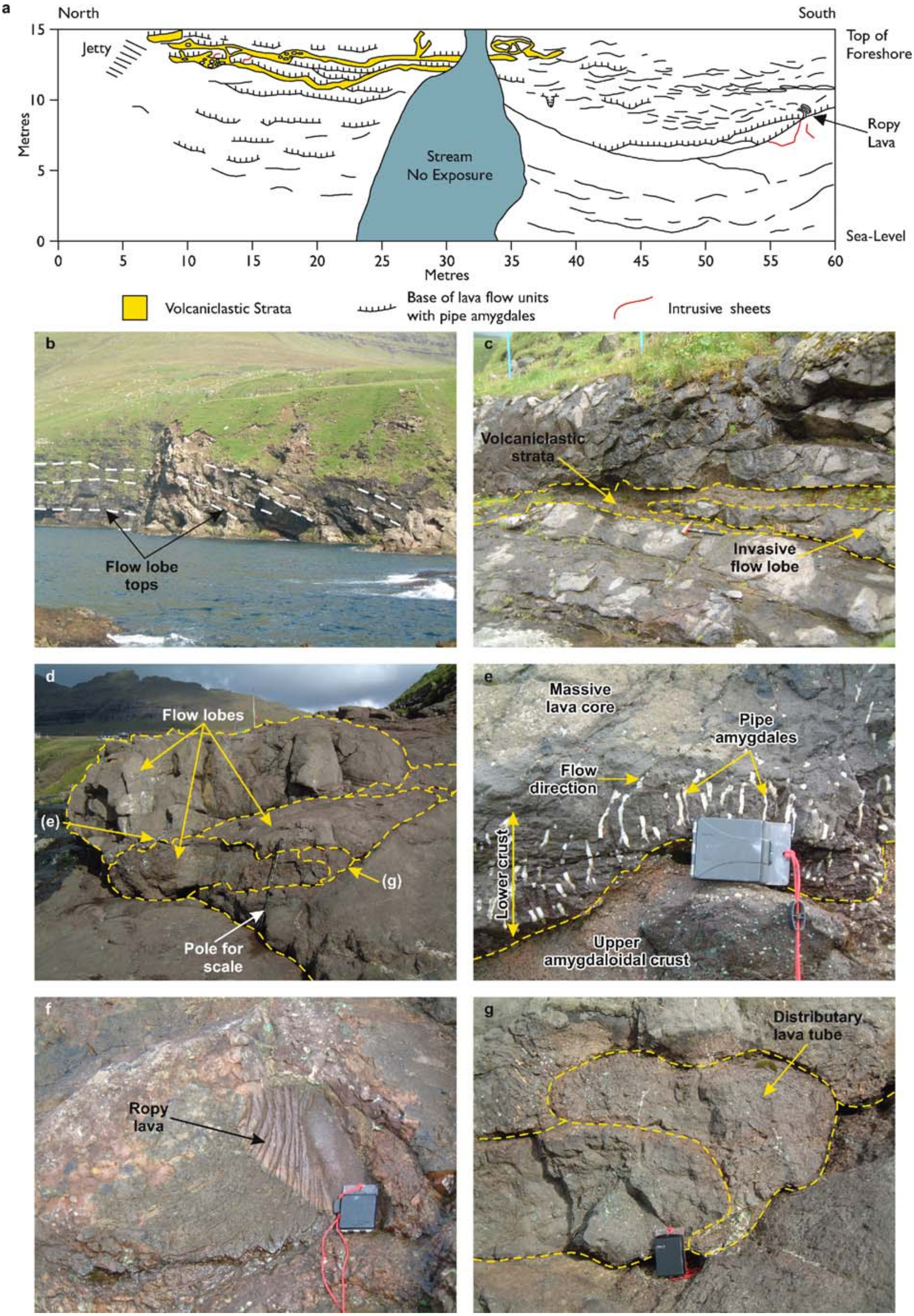


Fig. 11 Cliff-section of compound pahoehoe flows, Enni Formation, Høvdin, north Skúgvoy. The tops of two compound flows, comprising numerous flow lobes, are identified by in-weathering surfaces and the presence of volcaniclastic lithologies. The sequence is capped by the Argir Beds, a bedded sequence of reddish-brown volcaniclastic sandstones and siltstones, overlain by a sheet lobe with poorly-developed columnar jointing. Height of cliff is ~130 m


a
North
South
Jetty
Top of Foreshore
Stream No Exposure
Ropy Lava
Sea-Level
Metres
15
10
5
0
0 5 10 15 20 25 30 35 40 45 50 55 60
Metres
Volcaniclastic Strata
Base of lava flow units with pipe amygdales
Intrusive sheets
b
Flow lobe tops
c
Volcaniclastic strata
Invasive flow lobe
d
Flow lobes
(e)
(g)
Pole for scale
e
Massive lava core
Pipe amygdales
Flow direction
Lower crust
Upper amygdaloidal crust
f
Ropy lava
g
Distributary lava tube

◀ **Fig. 12** Compound lava flow of the Malinstindur Formation infilling a channel at Viðareiði, Viðoy. **a** Schematic diagram of the section with the main elements highlighted. A ~1.2 m thick volcaniclastic section has been passively invaded and inflated by thin pahoehoe flow units. **b** To the north of the Viðareiði section, the flow lobes have a near-horizontal orientation and at the margins of the channel the flow lobes mantle the topography at an angle between 20 and 30°. **c** An apophysis of lava invading and splitting the volcaniclastic sequence, causing bifurcation of the strata. **d** P-type pahoehoe flow lobes. **e** Basal crust composed of pipe amygdales with a maximum length of ~8 cm. Some of the pipe amygdales are curved in the direction of flow. **f** Ropy lava that has the appearance of twisted bread sticks. **g** A distributary lava tube with a cross-sectional area of ~0.14 m^2. In (**c**) the hammer is ~40 cm long, in (**d**) the pole is ~1.5 m long, and in (**e**), (**f**) and (**g**) the compass is 10×6 cm

core and upper tiers. Elsewhere in the formation, upper crusts of flow lobes typically exhibit amygdale zones, comprising upper amygdale-rich and lower non-amygdaloidal divisions, for example, a 4.12 m thick flow lobe exposed in the quarry, 700 m WNW of Sørvágar, Vágar, which consists of a lower crust (0.12 m), a massive core (2.2 m) and an upper crust (1.8 m) that displays a number of amygdale zones.

The lower and upper surfaces of many MF and EF flow lobes exhibit classic ropy structures characteristic of pahoehoe lavas. The ropy structures vary in scale, for example at Viðareiði, Viðoy, where large-scale convex ropes are observed covering an area of 3×2 m. Within the same section, small-scale ropy structures occur on the upper surface of a flow lobe covering an area of ~40×~40 cm (Fig. 12f). The convex nature of the ropy structures can, if statistically validated, be used as a local palaeoflow direction indicator.

Tree moulds are relatively rare within the MF and within the compound lava flow sequences of the EF. Good examples occur on the coast at Syðradalur, west Streymoy, and on the wave-cut platform on the western side of the Urðin isthmus, 700 m west of Sandur, Sandoy, where the upper surface of an ~1.5 m thick flow lobe contains the basal hemi-cylinders of seven tree moulds, all aligned in the same direction (azimuth 54–79°) and with the cavities lined with well-preserved cellular structures (Fig. 13).

Lava tubes

Infilled lava tubes are evident throughout the MF and within the compound lava flow sequences of the EF, ranging from master tubes (Fig. 14), through to smaller distributary tubes (Fig. 12g). The master tubes form prominent, stand-alone features, which typically have resisted weathering better than the flow lobes within which they formed. Cross-sectional areas for the master tubes range from as little as 3 m^2, up to ~160 m^2. The distributary lava tube within the flow field at Viðareiði, Viðoy (Fig. 12g) has a cross-sectional area of 0.14 m^2 and is akin to those described by Rowland and Walker (1990) from Hawai`i.

Emplacement mechanisms

Sheet lobes

Previous workers have interpreted the sheet lobes in the BF and EF as ′a′a (e.g., Hald and Waagstein 1984), primarily on the presence of rubbly flow tops. It has been demonstrated that ′a′a lavas form from rapid and typically confined emplacement, leading to high strain rates, which disrupts the upper crust and results in the formation of rubbly flow tops (Rowland and Walker 1990; Keszthelyi and Self 1998; Hon et al. 2003). However, we interpret the rubbly flow tops in the BF and EF to be the result of pervasive subaerial chemical weathering, as they show typical features associated with the formation of saprolitic boles. For example, the ~24 m thick sheet lobe from Borehole 8 of the Hov-Øravík tunnel project, consists of a rubbly flow top that grades upwards into the overlying volcaniclastic claystone, forming a ~3.3 m thick saprolitic bole akin to others described from the BF (cf. Parra et al. 1987). Weathering rates would, most likely, have been extremely high during the emplacement of the BF and EF sheet lobes because of the heightened humid and wet conditions imposed by the Palaeocene-Eocene Thermal Maximum (cf. Retallack 2001; Schmitz and Pujalte 2003). The rubbly flow tops, within the BF and EF, typically correspond to the upper crusts which are highly amygdaloidal, suggesting that weathering agents were able to exploit the increased surface areas of the initially unfilled vesicles (cf. Sheldon 2003). Therefore, it is inferred that the upper crusts were once smooth and undisrupted.

Fig. 13 Lava cast of charcoal on the inner surface of a tree mould, western side of the Urðin isthmus, 700 m west of Sandur, Sandoy. The charcoal shrank, forming a boxwork pattern of cracks, which were filled by the fluid lava. Scale on ruler is in centimetres

There are a number of other features that suggest that these sheet lobes were not emplaced under high strain rates. First, the amygdales in the upper crusts are predominantly not deformed, implying that the exsolved gas bubbles rose through stationary or relatively slow-moving lava (cf. Polacci et al. 1999). This feature suggests that if the brecciation is the result of high strain rate emplacement, it would have had to have occurred after bubble rise. This seems unlikely, given the large upper crust thicknesses observed in typical BF and EF sheet lobes. Second, the sharp, planar bases to the sheet lobes and the general lack of disruption to the underlying sedimentary units suggest that the lobes were emplaced passively, i.e. non-turbulently (Keszthelyi and Self 1998). Third, if the lobes of the two formations were rapidly (i.e. turbulently) emplaced ′a′a lavas, then extensive basal clinkery facies would have developed, caused by the upper clinkery surface of the flowing lava being overridden at the flow-front in a caterpillar fashion (Keszthelyi and Self 1998). Such basal facies are not seen. Where hyaloclastite and blocky peperite pockets are seen, these were caused by the quenching of the lava by small, standing bodies of water on the underlying surface, or interaction with water-saturated sediments.

There is also a number of other differences between ′a′a lavas on Hawai`i and the BF and EF sheet lobes, including a lack of the following features from the Faroese flows: (a) fragments of rubbly flow top in the core, (b) an upper core extending into the rubbly flow top, and (c) a limited lateral extent (i.e. being restricted to channels).

The data presented above indicate that the sheet lobes of the BF and the EF are not ′a′a in character and the subdivision of the sheet lobes into three tiers (basal crust, core and upper crust) based on amygdale distribution patterns suggests that they are actually P-type (pipe amygdale bearing) pahoehoe (cf. Self et al. 1998). This is significant because it has been documented that ′a′a and pahoehoe lavas form under different emplacement conditions (volumetric flow rates, flow front velocities, viscosities, etc.) (e.g., Rowland and Walker 1987, 1990; Keszthelyi and Self 1998; Cashman et al. 1999; Gregg and Fink 2000; Hon et al. 2003).

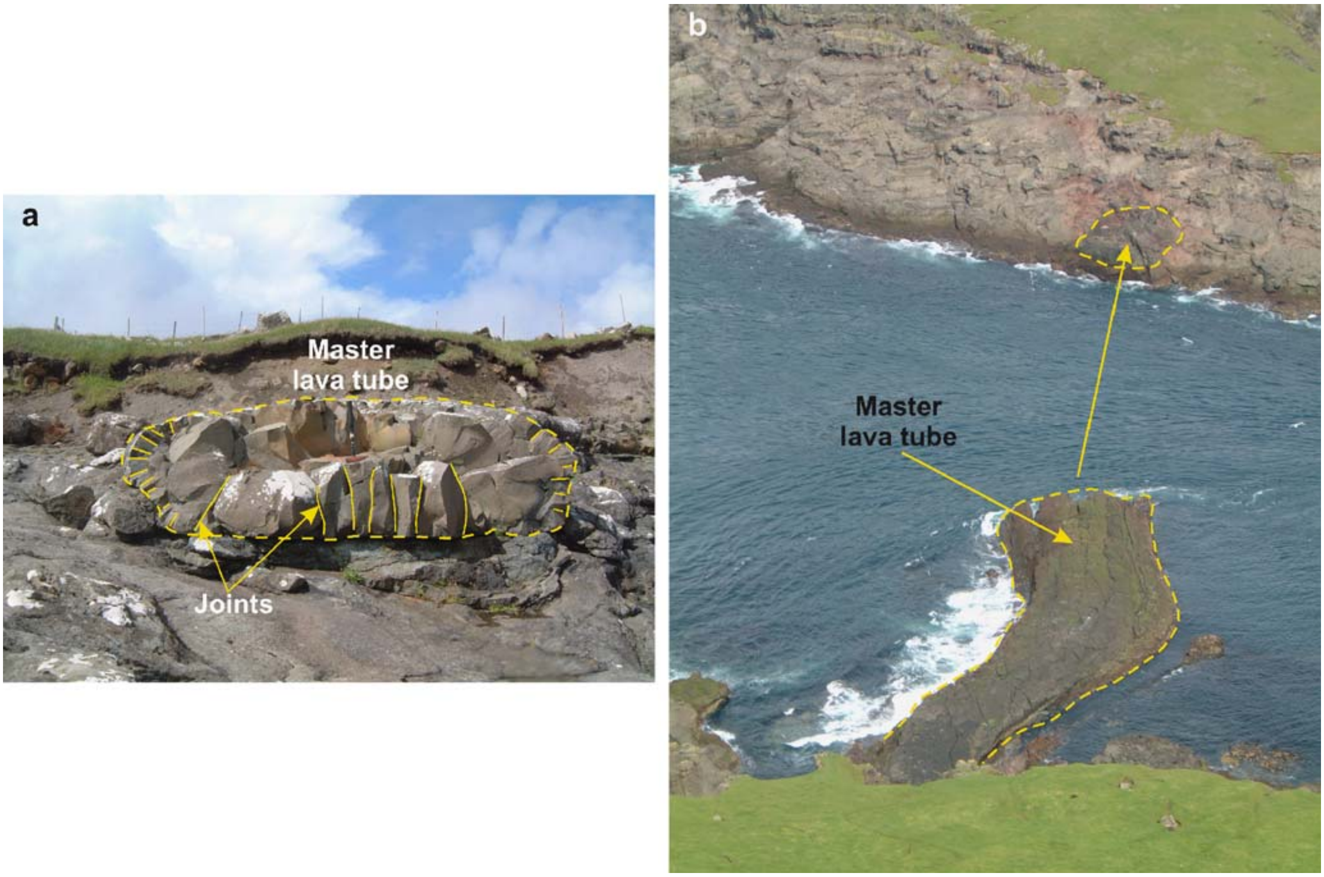


Fig. 14 Master lava tubes within compound lava flows. **a** A master lava tube in the Enni Formation, Rituvík, Eysturoy, defined by a radially-distributed joint pattern within the more resistant basalt of the tube, relative to the softer 'host' material. Hammer shaft is ~40 cm long. **b** A master lava tube from the Malinstindur Formation, Trøllhøvdi, NW Sandoy. The tube forms a spit at least 100 m in length and varies in width between 10 and 20 m, giving it a cross-sectional area of ~160 m^2. The same tube-like profile is noted on Trøllhøvdi, to the NW, thus extending the tube length by at least 140 m, without any substantial change in cross-sectional area

The occurrence of amygdaloidal zones in the upper crusts and the associated downward decrease in the abundance of amygdales and an increase in maximum amygdale size, strongly suggest that the pahoehoe sheet lobes of the BF and EF were emplaced through inflation (cf. Cashman and Kauahikaua 1997). However, the normalised upper crust thickness (upper crust thickness/total flow lobe thickness) for the BF sheet lobe from Borehole 8 of the Hov-Øravík tunnel project is 0.09, whereas inflated flow lobes of the Roza Member, CRBG, NW USA, have an average normalised upper crust thickness of 0.38, and range from 0.16 up to 0.56 (Thordarson and Self 1998). However, the upper crust of the aforementioned BF sheet lobe, from Borehole 8, is predominantly rubbly and forms part of a soil profile. Therefore, weathering and erosion of this upper crust may account for the low normalised upper crust thickness of 0.09. The upper crust thicknesses of other flow lobes may be intact, but caution is advised as some of the flow tops may also have been eroded. This is suggested by the presence of intra-formational sandstones containing lithoclasts of chilled basalt within the BF. Accordingly, a more systematic investigation of the amygdale distribution patterns of the BF and EF sheet lobes is required in order to constrain this parameter and any vertical variations in amygdale abundance and size.

As a result of weathering, erosion and lack of exposure, it is impossible to determine the areal extents of individual sheet lobes of the BF and EF and, in turn, their volumes. Therefore, it is currently difficult to quantify the effusion rates and flow front velocities for these sheet lobes. However, limited lateral extents of the sheet lobes over distances up to ~10 km, suggest that they are, most likely, more voluminous than the compound flows of the MF and EF and were erupted at higher effusion rates (cf. Self et al. 1996).

Compound lava flows

The compound lava flows of the Malinstindur and Enni formations preserve abundant evidence that they were emplaced through inflation (cf. Hon et al. 1994; Self et al. 1998). P-type flow lobes of the MF and EF display many characteristics similar to those observed for actively inflating pahoehoe flow lobes of Hawai`i (Hon et al. 1994). First, the flow lobes were emplaced in a slow and passive manner, rather than rapidly, as shown by the sequence at Viðareiði, Viðoy, which have invaded a thin clastic sedimentary sequence without disturbing or destroying sedimentary structures. The poorly lithified nature of these strata, combined with the preservation of minor sedimentary structures, point towards passive emplacement of the flow lobes, similar to the situation where aeolian sands have been preserved in South Greenland, Namibia and Brazil by compound flows (Clemmensen 1988; Jerram and Stollhofen 2002; Scherer 2002). A passive emplacement model is also supported by the presence of ropy structures on their upper surfaces, which formed when the flexible crust was deformed by the slow-moving lava of the underlying core (cf. Fink and Fletcher 1978).

The presence of tree moulds within MF and EF compound flows at Syðradalur, west Streymoy, and at Urðin, Sandoy, respectively, also implies passive emplacement, as they exhibit well preserved cellular structures that would, most likely, have been destroyed by movements of the flexible crust during flowage. The tree moulds that occur on the upper surface of the flow lobes also suggest the inflation process; initially, the trees would have been brought down by the advancing flow lobe and incorporated into the upper crust, leading to the formation of the tree moulds. Subsequent injections of fresh lava resulted in the upper crust being uplifted, along with the tree moulds, placing the moulds at an elevation of 1.5 m above the base of the flow lobe.

The division of the MF and EF P-type pahoehoe flow lobes into three tiers (basal crust, core, and upper amygdaloidal crust) is characteristic of inflated pahoehoe flow lobes, particularly where the upper crust exhibits amygdale zones (Hon et al. 1994; Cashman and Kauahikaua 1997; Self et al. 1998). Based upon observations made on actively inflating pahoehoe flow lobes on Hawai`i, it is possible to estimate the time required to form the upper crust, i.e. when the flow lobe stopped inflating (Hon et al. 1994). The time taken to form the upper crust conforms to the empirical equation:

$$t = 164.8\,C^2$$

where: t is the time in hours, 164.8 is an empirically determined constant, and C is the upper crust thickness in metres (Hon et al. 1994). Time differences (i.e. cooling rates) between the Faroes MF and EF flow lobes and Hawaiian lobes are expected due to differences in rainfall and (possibly) thermal properties (heat capacity, diffusivity, and latent heat of crystallisation). At present, the above equation is the only tool available for determining the duration of lava emplacement and, therefore, it shall be assumed that the equation applies reasonably to the compound flow lobes of the MF and EF. Using the upper crust thicknesses given above, the flow lobe at Viðareiði would have been active for 10.3 h, the average flow lobe in the Vestmanna-1 well would have been active for 9.4 days, and the flow lobe at Sørvágar, Vágar, which exhibits very good amygdale zones, would have been active for 22.2 days. These flow lobes have normalised upper crust thicknesses of 0.16, 0.53 and 0.44, respectively, which fall within the range expected for inflated flow lobes (Thordarson and Self 1998).

Rowland and Walker (1990) demonstrated that pahoehoe flow fields from shield volcanoes on Hawai`i advance at average flow front velocities between 1 and 10 m h^{-1}, although an individual flow lobe may advance as fast as 120 m h^{-1}. Using these average velocities and the calculated lava emplacement durations discussed above, it is estimated that the flow lobe from Sørvágar may have travelled up to 5,330 m from the site of eruption. However, this is a conservative estimate, as lava tubes are ubiquitous throughout the MF. The compound lava flow sections of the EF suggest that such networks, akin to those depicted by Rowland and Walker (1990), were an important and thermally efficient method of transporting magma significant distances from the source of the eruption (cf. Atkinson et al. 1976; Greeley 1982, 1987; Rowland and Walker 1990; Kauahikaua et al. 1998; Stephenson et al. 1998; Kadel and Greeley 2000). The general lack of exposure along the lengths of Faroese lava tubes makes it unclear as to how far the lava could have been transported before it emerged at the front of a lava tube. However, the longest estimated lava tube on Earth is ~100 km (Atkinson et al. 1976) and the average length of terrestrial lava tubes is considerably less (Kadel and Greeley 2000). This implies that the MF and EF compound lava flows were, most likely, erupted locally within 100 km of the Faroe Islands.

Pipe amygdales within the basal crusts of MF and EF compound flow lobes suggest that they were emplaced over slopes of <4° (Walker 1987). As no transitions between pahoehoe and ′a′a lava have been observed in the MF and EF, this indicates that the palaeotopography was relatively flat (<4°) at the time of lava emplacement, allowing for low strain rates to be maintained below the yield strength of the upper crust (cf. Hon et al. 2003; Kauahikaua et al. 2003). The shallow slopes may also account for the preponderance of filled lava tubes, hindering substantial drainage of the tube network. As the average regional dip for the MF and EF is between 2° and 4°, this implies that, prior to structural adjustments, the palaeo-surface was even lower, most likely <2°.

The dominance of inflated pahoehoe lava, lava tubes, limited areal extents (tens of kilometres) and the emplacement on low slopes (<2°) are consistent with eruption from point-sourced, low shield volcanoes (Greeley 1977, 1982; Walker 1993). This is further supported by the flow lobe (<0.5–4 m) and compound flow (~20 m) thicknesses of the MF and EF, which are comparable to those reported for the flow lobes (1–5 m) and compound flows (~35 m) of the Snake River Plains, Idaho, which were suggested by Greeley (1977, 1982) to have been erupted from low shields with slopes typically <0.5° and having areal extents of <300 km^2. This concurs with the interpretation that certain lava flows of the FIBG formed broad, flat shields, with diameters of ~15 km and slopes of <0.5° (Noe-Nygaard 1968), and which he referred to as being of the scutulum type.

Simple vs compound lava flows

The data presented here suggests that both the simple and compound lava flows of the FIBG were emplaced over low slopes (<2°) as inflating pahoehoe lava flow lobes. However, the simple flows composed of a single sheet lobe contrast with the compound flows that comprise numerous overlapping and anastomosing flow lobes. The average thicknesses of the simple and compound lava flows are comparable at ~20 m (Rasmussen and Noe-Nygaard 1970), although the simple flows appear to have greater areal extents, suggesting they are more voluminous. Also, it has been suggested that the simple flows of the BF and EF were erupted from fissure systems to the west of the Faroe Islands (Waagstein 1988; Larsen et al. 1999), in contrast to the compound flows of the MF and EF which were, most likely, erupted from low shields, local to the Faroe Islands (cf. Noe-Nygaard 1968; Greeley 1977, 1982).

The difference in lava flow morphology is still poorly understood and is controlled by many parameters (Self et al. 1998). The compound lava flows of the FIBG appear to have been emplaced from long-lived eruptions that were broken into distinct episodes by small pauses in the volcanism (little or no time for weathering, erosion or sedimentation) and/or by migrations of the vent resulting in thin and anastomosing flow lobes (cf. Self et al. 1998). Conversely, the simple lava flows of the FIBG, each consisting of a single sheet lobe, were emplaced from an eruption with a continuous supply of lava, although this does not imply that the supply rate was constant and most likely waned during the eruption. The supply rates were also low enough to maintain strain rates below the yield strength of the upper crust. This consequently limited brecciation of the upper crust and prohibited the formation of ′a′a lava (cf. Keszthelyi and Self 1998). The difference between the sheet lobes and the compound flows of the FIBG therefore appears to be intrinsically linked to the supply of eruptible magma (i.e. continuous vs punctuated) and the nature of the vent system. The sheet geometry of the simple flows suggests that they were erupted with a continuous supply of magma along sections of a fissure system and, consequently, were able to spread out laterally over a wide area. This contrasts with the compound flows erupted from point-sourced, low shield volcanoes, which had a pulsed supply of magma, although the eruption was sustained due to continued magma recharge.

Another contrasting feature is the preponderance of filled lava tubes observed throughout the compound lava flow sequences of the Malinstindur and Enni formations, whereas no lava tubes, filled or unfilled, have been observed for the sheet lobes of the Beinisvørð and Enni formations. This difference between simple and compound flows is extremely apparent on the Faroe Islands and suggests that there is a controlling factor that hinders the formation of lava tubes in sheet lobes. One explanation for this may be that the sheet lobes were emplaced during voluminous eruptions where they advanced as a single large sheet lobe rather than numerous coalescing lobes (cf. Self et al. 1998). This concept is, in part supported by the absence of lava tubes for the voluminous lava flows (100 to >1,000 km^3) of the CRBG, whereas lava tubes are clearly observed for the small-volume compound lava flows (<1 km^3) of Hawai`i (Self et al. 1996) and the Snake River Plains (Greeley 1977, 1982). The suggestion that lava tubes are difficult to recognise within sheet lobes of CFB provinces because they are typically emplaced over low slopes and, therefore, do not drain (Self et al. 1996), is discounted. Specifically, the numerous filled lava tubes that are recognised within the compound flows of the MF and EF were emplaced over low slopes, similar to the sheet lobes of the BF and EF.

The Enni Formation, which consists of a mixture of compound and simple lava flows, is similar to the plains volcanism first described by Greeley (1982) for the Snake River Plains, Idaho, NW USA. Here, small-volume compound flows erupted from low shields are intercalated with sheet lobes erupted from fissures that infill the low-lying areas between the shields, thus maintaining a regionally planar surface.

Conclusions

Our reassessment of the morphology of the basaltic lavas of the Faroe Islands indicates that the simple flows of the Beinisvørð (BF) and Enni (EF) formations, previously identified as being turbulently emplaced 'a'a (e.g., Hald and Waagstein 1984), are, in fact, large inflated pahoehoe flow lobes. These inflated flow lobes were erupted continuously as large, single, sheets of lava (i.e. sheet lobes). Such continuous, high-volume eruptions may account for the absence of lava tubes within these sheet lobes, similar to other CFB lava fields, for example, the Columbia River Basalt Group of NW USA (e.g., Self et al. 1996; Thordarson and Self 1998). Previously identified rubbly flow tops on BF and EF sheet lobes are, in fact, saprolitic boles formed by intense subaerial chemical weathering under heightened warm and wet conditions, possibly imposed by the Palaeocene–Eocene Thermal Maximum (Kennett and Stott 1991). Consequently, caution is advised when trying to apply the emplacement duration equation of Hon et al. (1994) to the upper crust thicknesses, because these are most likely minimum values due to weathering and erosion. Therefore, the idea that these lavas were emplaced turbulently is rejected, a conclusion which is supported by the dominance of sharp planar bases to the sheet lobes, which suggests that they were emplaced passively and without significantly disrupting any subjacent inter-lava sedimentary lithologies.

Conversely, the Malinstindur Formation and sections of the Enni Formation are composed of significantly thinner inflated pahoehoe flow lobes, forming compound flows. These flow lobes were, most likely, erupted from point-sourced, low shields at effusion rates (<5–10 m^3 s^{-1}) comparable to those reported from Hawaiian eruptions (Rowland and Walker 1990). Applying the emplacement duration equation of Hon et al. (1994) to MF flow lobes, values of between 10 h and 22 days are obtained. Ubiquitous to the MF and the compound flows of the EF are lava tube networks, which would have represented a thermally efficient means of transporting lava over significant distances (Keszthelyi 1995). Using terrestrial lava tube lengths and flow front velocities typical of pahoehoe lavas on Hawai`i (1–10 m h^{-1}), it appears that the compound lava flows of the MF and EF occur relatively close to their sites of eruption, most likely within tens of kilometres. Consequently, it seems likely that the compound lava flows of the MF and EF occur only local to the Faroe Islands and may not extend offshore for any significant distance. However, other compound-braided basalt lava flow fields, not necessarily time equivalent to the MF and EF, may be found in association with the central igneous complexes identified in the NE Atlantic margin.

Acknowledgements The majority of this work was carried out at the University of Glasgow as part of a PhD project and we would like to thank NERC and Statoil (UK) Ltd. for funding the CASE studentship (GT/04/99/ES/94). Thanks also go to David Brown for useful comments on an earlier version of this paper and the authors would also like to thank Scott Rowland and Steve Self for their constructive and thoughtful reviews.

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