Paleogene volcanic rocks in the northern Faroe-Shetland Basin and Møre Marginal High: Understanding lava field stratigraphy

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Onshore exposures of the North Atlantic Igneous Province have been studied in detail for over 200 years, whereas the more extensive offshore volcanic stratigraphy is significantly less well-constrained with the exception of a small number of boreholes. Within this study we integrate seismic and well data across the northern Faroe-Shetland Basin and Møre Marginal High to improve understanding of the volcanic stratigraphy and its relationship to rifting in the NE Atlantic. Volcanic seismic facies, including compound and tabular lavas and hyaloclastites (representing subaerial and subaqueous emplacement) are interpreted across the study area, calibrated by the Lagavulin borehole. The volcanic sequence was erupted between ~56.1-55.2 Ma, when subaerial conditions dominated in the region, but extensive lava deltas developed in a seaway east of the main lava field. Geochemical and thickness variations within the volcanic pile have important implications for the regional rifting history. MORB-like lavas at the base of Lagavulin, which thicken substantially northward, support an early onset of rifting near the Møre Marginal High prior to major thinning associated with continental breakup to the south and north. Following a period of erosion, smaller-degree melting caused eruption of higher-TiO_2/Zr lavas, marking the final "pre-breakup" volcanism before emplacement of seaward-dipping reflectors.

Introduction

Magma-rich rifted margins form through extension of continental crust accompanied by voluminous intrusive and extrusive magmatism, typically emplaced over short timescales (1-5 Myrs) preceding and during continental breakup (e.g. Fitton et al., 2000; Geoffroy, 2005; Bryan and Ernst, 2008; Franke, 2013). Such magmatic accumulations, often characterised as Large Igneous Provinces (LIPs), are found globally (e.g. Mahoney and Coffin, 1997) and typically contain extremely large volumes of magmatic rocks. For example the volume of the North Atlantic Igneous Province (NAIP) has been estimated at 6-10 x 10^6 km^3, covering an area of 1.3 x 10^6 km^2, approximately five times the size of the UK landmass (Eldholm & Grue 1994; Hansen *et al.* 2009). Interest in volcanic rifted margins has increased in recent years, as the volcanics are often emplaced into pre-existing sedimentary basins on volcanic margins with proven petroleum systems (e.g. Rohrman, 2007; Varming et al., 2012). There are significant challenges associated with exploration in such basins, including seismic imaging beneath the volcanic pile, drilling through volcanic sequences, and poor understanding of how the intrusive and extrusive systems influence the petroleum system (e.g. enhanced maturation of source rocks due to intrusive complexes, compartmentalisation and distribution of reservoir units; e.g. Schutter 2003; Varming et al., 2012; Rateau et al., 2013; Schofield and Jolley, 2013; Jerram 2015; Millett et al., 2016; Angkasa et al., 2017; Spacapan et al. 2017; Hardman et al., 2018a; Mark et al., 2018). These challenges have resulted in volcanic margins being typically underexplored, with a high potential for new hydrocarbon discoveries.

The NAIP is one of the best studied LIPs in the world (e.g. Roberts et al., 1984b; Eldholm and Grue, 1994; Saunders et al., 1997; Horni et al., 2017). The NAIP has been extensively investigated, particularly where volcanic rocks outcrop on the Faroe Islands, Iceland, Greenland, Ireland and Western Scotland (e.g. Mussett et al., 1988; Emeleus and

Bell, 2005; Passey and Jolley, 2009; Brooks, 2011; Millett et al., 2017; Figure 1). As a result, knowledge of the onshore stratigraphy of extrusive sequences is detailed, but the offshore succession, which represents the bulk of the province (~80% of its areal extent - Emeleus and Bell, 2005) is less well constrained. On a localized basin scale, detailed understanding is available where wells have penetrated the volcanic succession, but these are low in number (less than 50 exploration wells across the whole NE Atlantic Margin), so that the margin-wide volcanic stratigraphy is only broadly understood (e.g. Ritchie and Hitchen, 1996; Jerram et al., 2009; Ritchie et al., 2011; Abdelmalak et al., 2016a; Horni et al., 2017). More detailed knowledge of the volcanic succession allows better constraint of the evolution of basins along the margin and the progression from continental rifting to breakup and seafloor spreading. In addition, understanding the distribution of volcanic units within the succession allows more accurate pre-drill predictions of the thickness of the volcanic pile and its constituent facies, which will help to mitigate issues when drilling through the succession.

In order to form a more detailed understanding of the volcanic stratigraphy, in particular the phases of volcanism and relationship to the rift-to-drift transition of the NE Atlantic Margin, this study integrates seismic and well data to extend detailed stratigraphic interpretations within the northern Faroe-Shetland Basin and Møre Marginal High (Figure 1).

Geological Setting and Tectonic History

The Faroe-Shetland Basin is located on the NW European Shelf between the Faroe and Shetland Islands. Its northern limit is defined by the Møre Marginal High, a NE-SW trending platform ~200 km due north of Shetland (Figure 1).

The Faroe-Shetland Basin and its constituent sub-basins have a dominant NE-SW orientation, inherited from the pre-existing Caledonian structural trend (Ritchie *et al.* 2011). The sub-basins are bounded by structural highs comprising Precambrian crystalline basement and Palaeozoic-Mesozoic sedimentary rocks, and are filled by a variety of

Mesozoic-Recent sedimentary and igneous strata (Lamers & Carmichael 1999). The Faroe-Shetland Basin formed through multiple periods of extension, from the Permo-Triassic breakup of Pangaea to the Paleocene and North Atlantic rifting (Ziegler 1988; Doré *et al.* 1999). The Møre Marginal High is untested by drilling so its internal structure, and therefore evolution, are poorly constrained. It is interpreted to have formed during Late Cretaceous-Paleocene extension of the Atlantic Margin, and to comprise a thick sequence of Paleocene-Eocene volcanics resting on Mesozoic and older rocks (e.g. Blystad et al., 1995; Berndt et al., 2001; Ritchie et al., 2011; Theissen-Krah et al., 2017)

A number of NW-SE trending structural lineaments or "transfer zones" have been interpreted from potential field data, orientated perpendicular to basin trends along the entire margin (Rumph et al., 1993; Doré et al., 1997; Ellis et al., 2009; Figure 1). These are thought to have formed during the Caledonian Orogeny, with some authors considering that they may have influenced sedimentation routes and magma emplacement (Rumph *et al.* 1993; Archer *et al.* 2005; Jolley *et al.* 2005; Ellis *et al.* 2009). Others, however, suggest that the inferred lineaments had no regional, structural or geomorphological expression during the Cenozoic, and that their effects can be attributed to igneous intrusions and the underlying Late Cretaceous rift system (Moy & Imber 2009).

Seafloor spreading in the North Atlantic region is thought to have initiated ~66 Ma in the Labrador Sea, west of Greenland, but did not begin in the North Atlantic until ~55-54 Ma (Ritchie *et al.* 2011; Ellis & Stoker 2014). Rifting was associated with intense and voluminous magmatism of the NAIP and was preceded by regional uplift in the Danian (66-61.6 Ma; Jolley and Bell, 2002), commonly attributed to impingement of the proto-Iceland Plume at the base of the lithosphere (e.g. White, 1989; White and McKenzie, 1989; Saunders et al., 1997; Hansen et al., 2009; Hardman et al., 2018b). Spreading east of Greenland first occurred along the Reykjanes Ridge (forming the Iceland and Irmiger Basins) and the Aegir Ridge (forming the Norway Basin). The Aegir Ridge and Labrador Sea spreading centres became inactive around ~34 Ma; this led to significant plate reorganisation in the North Atlantic, and eventual spreading along the Kolbeinsey Ridge (Roest and Rivastava, 1989; Gaina et al., 2009; Gernigon et al., 2012; Ellis and Stoker, 2014; Figure 1). The remainder of the Cenozoic involved regional thermal subsidence, interrupted by episodic compressional pulses, resulting in inversion, uplift and erosion (Roberts, 1989; Boldreel and Andersen, 1993; Doré et al., 1999; Kimbell et al., 2017)

In order to remain consistent with previous work on the Paleocene volcanics of the UK, this study uses the lithostratigraphy for the Faroe-Shetland Basin presented in Ritchie et al. (2011) and Stoker and Varming (2011) and the Paleocene sequence stratigraphic framework (T-sequences) of Ebdon et al. (1995), which is based largely on biostratigraphy (Figure 2).

Lower Paleogene Sedimentary Succession in the Faroe-Shetland Basin

The lower Paleogene sedimentary succession in the Faroe-Shetland Basin comprises five main formations (Figure 2): the Sullom Fm. (oldest), Vaila Fm., Lamba Fm., Flett Fm. and Balder Fm. (youngest). The Sullom Fm. is Danian in age (sequence T10-T22; ~66-62.9 Ma) and is dominated by basinal marine mudstones with submarine fan sandstones around the margins of the Faroe-Shetland Basin (Knox *et al.* 1997a; Ritchie *et al.* 2011; Mudge 2015). The Vaila Fm. is Danian-Selandian in age (sequence T22-T35; ~62.9-58.7 Ma) and comprises mainly deep-marine mudstones punctuated by thick basinal sandstones representing submarine fan progradation (Lamers & Carmichael 1999; Ritchie *et al.* 2011). The Lamba Fm. was deposited in the Thanetian (sequence T36-T38; ~58.7-56.1 Ma) and consists of a predominantly NW-prograding package of mudstones and subordinate sandstones (Smallwood & Gill 2002; Shaw Champion *et al.* 2008; Ritchie *et al.* 2011; Hardman *et al.*

2018a). The Flett Fm. is Thanetian-Ypresian in age (sequence T40-T45; ~56.1-54.9 Ma) and is dominated by northerly-prograding sandstones and mudstones deposited in shallow-marine to terrestrial environments (Knox *et al.* 1997b; Smallwood & Gill 2002; Ritchie *et al.* 2011). The Balder Formation was deposited during the Ypresian (sequence T50; ~54.9-54.3 Ma) and comprises predominantly mudstones with frequent tuffaceous layers and interbedded sandstones (Knox *et al.* 1997b; Ritchie *et al.* 2011; Watson *et al.* 2017).

Paleogene Volcanism in the Faroe-Shetland Basin and Møre Marginal High

Volcanism in the Faroe-Shetland Basin and Møre Marginal High occurred as part of the NAIP, the total extent of which covers most of the North Atlantic margins, from the Rockall Basin to the Barents Sea, the Inner Hebrides Sea (west Scotland), Northern Ireland and East and West Greenland (e.g. Roberts et al., 1984b; Eldholm and Grue, 1994; Saunders et al., 1997; Naylor et al., 1999; Jolley and Bell, 2002; Ritchie et al., 2011; Hitchen et al., 2012; Figure 1). NAIP volcanism can be divided into two main phases: the first predating continental breakup, and the second occurring synchronously with breakup, signifying the transition from rift-to-drift tectonics and the start of seafloor spreading (Figure 2). This was followed by post-breakup volcanism at the site of seafloor spreading, lasting until the present day (Jolley & Bell 2002b).

Pre-breakup volcanism (~61.3-55 Ma; sequence T26 -T40; Vaila Fm. to Mid-Flett Fm.)

Tuffs and volcaniclastic material of Selandian age (sequences T26, T34 and T35; Vaila Fm.; 61.3-58.7 Ma), identified in cores and ditch cuttings, provide evidence for the earliest known volcanic activity in or near the Faroe-Shetland Basin (Jolley & Bell 2002b; Watson *et al.* 2017). During sequence T36 (Early Lamba Fm.; ~58.7-58.1 Ma) volcanism consisted of localised basaltic lava fields identified on the flank of Brendans Volcanic Centre (Jolley, 2009; Mclean et al. 2017) and in the Foula Sub-basin, plus regional deposition of ash preserved as the Kettla Tuff Member (Schofield *et al.* 2015; Watson *et al.* 2017). This was followed by a volcanic hiatus, during which deep marine sedimentation occurred in the Faroe-Shetland Basin (Jolley & Bell 2002b).

Widespread volcanism commenced in mid-sequence T40 (Flett Fm.; ~55.8 Ma), with basaltic volcanic successions (including a variety of lava flows, hyaloclastites and volcaniclastic facies) erupted across the entire North Atlantic margin from both widespread fissures and point-sourced volcanic edifices (e.g. Naylor et al., 1999; Jolley and Bell, 2002; Passey and Bell, 2007; Hansen et al., 2009; Passey and Jolley, 2009). This period represents the majority of volcanic activity observed within the Faroe-Shetland Basin.

Syn-breakup volcanism (~55-54 Ma; latest sequence T40-T50; Upper Flett Fm. to Balder Fm.)

Following a volcanic hiatus near the end of sequence T40 (Flett Fm.; ~55.2 Ma) a widespread unconformity developed, marked by regional deposition of sedimentary units (e.g. the coal-bearing Prestfjall Formation of the Faroe Island Basalt Group (Ellis *et al.* 2002; Passey & Jolley 2009) and fluvial and marine siliciclastics in the central Faroe-Shetland Basin (Schofield & Jolley 2013). When volcanism resumed in sequence T45 (Upper Flett Fm; 55.2-54.9 Ma), it was concentrated around the site of the proto-spreading axis, with emplacement of seaward-dipping reflectors (SDRs). SDRs are typically wedge-shaped packages of arcuate reflections dipping in a seawards direction, thought to represent the transition from continental to oceanic crust (e.g. Mutter et al., 1982; Roberts et al., 1984; Mutter, 1985; Franke, 2013). The syn-rift volcanic succession was erupted rapidly in sequence T45 (Jolley 2009; Passey & Jolley 2009) and includes localised lava fields, recognised only around the Corona Ridge, in the Judd Sub-basin and onshore in the Faroe Islands (Schofield & Jolley 2013).

In sequence T50 (Balder Fm.; 54.9-54.3 Ma) volcanism became localized at the presumed site of initial seafloor spreading in the North Atlantic. Flooding of this region resulted in widespread phreatomagmatic eruptions and deposition of Balder Formation tuffs across large areas of the North Atlantic (Roberts et al., 1984; Jolley and Bell, 2002; Watson et al., 2017).

Post-breakup volcanism (~54 Ma onwards; sequence T60-present)

This phase consisted primarily of further SDR emplacement passing into the formation of normal oceanic crust (Jolley & Bell 2002b). During this stage volcanism appears to have been (and still is) largely restricted to the zone of sea-floor spreading, with limited evidence of substantial intrusive or extrusive activity elsewhere. Fully marine conditions were established across the North Atlantic region ~1.5-2 Myrs after cessation of the sequence T50 (Balder Fm.) phreatomagmatic eruptions at ~54.3 Ma (Jolley & Bell 2002b).

Onshore volcanic stratigraphy: the Faroe Islands Basalt Group (FIBG)

Volcanic rocks of the NAIP have been studied in detail in outcrop and boreholes on the Faroe Islands, where they are termed the Faroe Islands Basalt Group (FIBG) (Figure 2). Passey and Bell (2007) and Passey and Jolley (2009) have subdivided the FIBG into 7 formations: 4 volcanic (Lopra, Beinisvørð, Malinstindur and Enni Formations) and 3 sedimentary (Prestfjall, Hvannhagi and Sneis Formations).

Lopra Formation

The Lopra Formation has been identified only in the Lopra-1/1A borehole, located onshore near the southern limit of the Faroe Islands, where it is at least ~1075 m thick (although the total thickness is still unknown). It comprises a lower unit of volcaniclastic sandstone that is at least 120 m thick, overlain by ~430 m of hyaloclastite, which is topped by ~60 m of volcaniclastic sandstone. The lower two-thirds of the formation are intruded by basalt sills and/or invasive lava flows (Ellis *et al.* 2002; Passey & Jolley 2009).

Beinisvørð Formation

The Beinisvørð Formation directly overlies the Lopra Formation within the Lopra-1/1A borehole. It has a gross stratigraphic thickness of ~3250 m, and is composed of laterally extensive tabular lava flows with an average thickness of ~20 m. The upper parts of lava flows are typically highly weathered, accompanied by the development of inter-lava palaeosols (Ellis *et al.* 2002; Passey & Bell 2007; Passey & Jolley 2009). Volcaniclastic lithologies, ranging from mudstones to conglomerates, become widespread within the upper 100-200 m of the formation (Passey & Jolley 2009).

The Beinisvørð Formation is overlain by the coal-bearing, mudstone-rich Prestfjall Formation, with an erosional contact between the two. This is locally conformably overlain by the Hvannhagi Formation, a syn-eruptive unit of pyroclastic and volcaniclastic rocks (Ellis *et al.* 2002; Passey & Jolley 2009).

Malinstindur Formation

The Malinstindur Formation is up to ~1350 m thick, and comprises almost exclusively compound basaltic lava flows (Passey & Bell 2007). Inter-lava beds of volcaniclastic sandstone and mudstone occur throughout the sequence, but are more common towards the top (Ellis *et al.* 2002; Passey & Jolley 2009). The Malinstindur Formation is disconformably overlain by the Sneis Formation, dominated by volcaniclastic sandstones and conglomerates. Sills are common within the Sneis Formation (Passey & Jolley 2009).

Enni Formation

The Enni Formation is the uppermost unit of the FIBG, and disconformably overlies the Sneis Formation. It has a minimum thickness of ~900 m, but up to ~1 km is estimated to have been eroded, based on zeolite studies (Waagstein *et al.* 2002; Jørgensen 2006; Passey & Jolley 2009). It comprises a mixture of tabular and compound lava flows, with common inter-eruption deposits of volcaniclastic mudstones and sandstones (Passey and Bell, 2007; Passey and Jolley, 2009; Millett et al., 2017)

Age and correlation of the FIBG

Palynological analysis of sedimentary units within the FIBG indicate that the Lopra Formation is equivalent to lower sequence T40 (early Flett Fm.; ~56.1 Ma), while the Beinisvørð Formation is upper sequence T40 (mid Flett Fm.; ~55.8 Ma). The Malinstindur and Enni Formations were both erupted within sequence T45 (upper Flett Fm.; ~55.2-54.9 Ma; Figure 2; Jolley, 2009; Passey and Jolley, 2009; Schofield and Jolley, 2013).

It has long been recognised that there is a significant discrepancy (up to ~5 Myrs) between the biostratigraphic and radiometric (K-Ar and Ar-Ar) age indications derived from the Beinisvørð Formation of the FIBG, with radiometric ages indicating an older age of eruption (e.g. Jolley *et al.* 2002; Waagstein *et al.* 2002; Storey *et al.* 2007; Jolley 2009; Passey & Jolley 2009; Ólavsdóttir *et al.* 2019). This has led to several authors proposing that volcanism of the FIBG occurred over a period of e.g. ~55-61 Ma, (e.g. Waagstein *et al.* 2002; Storey *et al.* 2007), although the analyses of Waagstein *et al.* (2002) have been disregarded as unreliable in a review of isotopic ages from the NAIP (Wilkinson *et al.* 2016). Ólavsdóttir *et al.* (2019) attempt to combine seismic-stratigraphic relationships and well data to constrain the age of the FIBG. They conclude that the Lopra and Beinisvørð Formations are equivalent to the Vaila (sequence T22-35; 62.9-58.7 Ma) and Lamba (sequence T36-38; 58.7-56.1 Ma) Formations, respectively, while the Malinstindur and Enni Formations correspond to the Flett Formation (sequence T40-45; 56.1-54.9 Ma). Such a discrepancy in ages obviously has important implications for any models that involve the timing of volcanism in the NAIP. Within this study we utilise the biostratigraphic framework of Passey

& Jolley (2009) and Schofield & Jolley (2013), that is consistently calibrated to offshore well penetrations within the Faroe-Shetland Basin.

Until issues surrounding the alteration state of samples analysed for Ar-Ar (both whole rock and feldspar mineral separates), from zeolite facies rocks are resolved, we do not consider the older isotopic ages from the FIBG unchallengeable given they contradict the clear climatic proxies for the PETM within this interval (Passey & Jolley 2009; Jolley *et al.* 2012). Explicitly, numerous concerns have been raised surrounding the Ar-Ar dating method for altered low-K tholeiitic basalt samples, even if macroscopically appearing fresh (e.g. Baksi 2013; Cramer *et al.* 2013; Wilkinson *et al.* 2016). Given no fresh plagioclase crystals were identified within the Lopra 1/1A borehole (Waagstein *et al.* 2002), within the interval that Storey *et al.* (2007) report older ages from plagioclase separates, we feel that the debate surrounding the age of the Beinisvørð Formation remains very much unresolved. From a seismic stratigraphic perspective, we acknowledge the extreme challenges faced by Ólavsdóttir *et al.* (2019), in seismically imaging the thick FIBG in Faroese waters based on the available seismic data. In this regard we maintain that the seismic evidence (not the interpretation of it) remains in line with the offshore calibrated biostratigraphic record, which does not support an older age for the Beinisvørð Formation, which we attribute to the Flett Formation.

The Beinisvørð Formation has been geochemically correlated with the pre-breakup Nansen Fjord Formation in East Greenland while the Malinstindur and Enni Formations have been correlated to the syn-breakup Milne Land Formation. The Lopra Formation does not appear to have a geochemical equivalent in East Greenland (Larsen *et al.* 1999; Passey & Jolley 2009; Søager & Holm 2009). The biostratigraphic and radiometric dates of the East Greenland formations also correspond well with their Faroese equivalents, thus supporting the correlation (Larsen *et al.* 1999; Storey *et al.* 2007; Passey & Jolley 2009).

Offshore volcanic stratigraphy in the northern Faroe-Shetland Basin – well penetrations

Within the Faroe-Shetland Basin 33 industry wells have been drilled into the extrusive volcanic succession, including 213/27-1z (Rosebank), 6104/21-1 (Brugdan), 214/04-1 (Tobermory) and 214/09-1 (Bunnehaven). The majority of these wells are located in the middle of the basin, particularly on and around the Corona High (Figure 3). In the northernmost Faroe-Shetland Basin a number of wells penetrate the extrusive volcanic sequence. These include Lagavulin (217/15-1/1Z) and wells drilled on nearby structural highs (Ben Nevis and the Erlend Volcanic Centre).

Lagavulin

Lagavulin, a wildcat exploration well drilled by Chevron between October 2010 and June 2011, is one of the best-studied volcanic boreholes in the Faroe-Shetland Basin. It is located close to the basin's northern margin with the Møre Marginal High, ~200 km due north of the Shetland Islands. The well penetrated just over 2.6 km of extrusive volcanic strata without exiting the base of the sequence and represents one of the thickest continuous drilled volcanic sections of the entire North Atlantic Igneous Province (Millett et al. 2015; Figure 4). Given that the nearest existing wells lie ~80 km to the E (Ben Nevis), ~100 km to the SW (Tobermory) and ~100 km to the SE (Erlend), Lagavulin provides an excellent opportunity to investigate the volcanic stratigraphy and development of the NAIP in this previously unexplored area of the basin.

Palynofloras recovered from volcaniclastic units throughout the volcanic sequence penetrated by Lagavulin are of latest Paleocene-early Eocene character. Two key taxa found in Lagavulin are *Caryapollenites circulus,* which is not recorded in sedimentary rocks older than the base of sequence T40 (base Flett Fm.; ~56.1 Ma) in the Faroe-Shetland Basin, and frequent *Alnipollenites verus*, a pollen that is common in the Faroe-Shetland Basin throughout the upper part of sequence T40 but becomes more rare after earliest sequence T45 (Upper Flett Fm.; 55.2 Ma) (Jolley 1997; Jolley *et al.* 2002). The association of these two taxa indicates that the entire volcanic succession of 2.6 km was probably erupted during sequence T40 (~56.1-55.2 Ma), in just under 1 Ma (Jolley 2009; Millett *et al.* 2015), and is therefore part of the pre-rift succession, equivalent to the Beinisvørð Formation of the FIBG (Jolley 2009; Millett *et al.* 2015).

A detailed stratigraphic and petrogenetic study of the drilled volcanic section of Lagavulin based on wireline logs and drill cuttings was undertaken by Millett et al., (2015). From this, two major volcanic sequences were identified within the well (Figure 4a). The lower sequence comprises ~280 m of highly reworked volcaniclastics passing up into ~300 m of hyaloclastite, which is overlain by ~500 m of tabular lava flows and an upper ~200 m unit of predominantly compound lava flows. Towards the top of the sequence the lavas are highly altered and contain reddened interbeds or boles, indicating a period of subaerial weathering. The amount of alteration and weathering suggest that this boundary has a good chance of being recorded regionally away from the wellbore. The upper sequence comprises ~400 m of fresh and relatively homogenous hyaloclastites overlain by ~550 m of tabular lava flows. A volcaniclastic interbed occurs ~200 m above the base of the tabular flows, indicative of a volcanic hiatus. This is succeeded by ~300 m of mixed volcaniclastic and epiclastic rocks, which is topped by a final tabular lava unit, ~40 m thick. These two sequences have been interpreted as prograding lava deltas that became emergent, separated by a period of volcanic quiescence and rapid relative subsidence (Millet et al., 2015). The topmost volcaniclastic-epiclastic unit indicates a second significant volcanic hiatus, which was dominated by the erosion of the emergent lava field; followed by the eruption of a final large lava flow, representing the end of eruptive activity in this region (Millett *et al.* 2015).

Thirteen geochemically defined stratigraphic packages were identified within the volcanic succession at Lagavulin, based on a combination of whole-rock major and trace element analysis (Millett *et al.* 2015). Clear geochemical variations occur between the two large-scale volcanic sequences. The lower sequence is dominated by samples with Mg# > 50, TiO_2 < 1.5 wt%, P_2O_5 < 0.15 wt%, Fe_2O_3 < 13.5 wt% and Zr < 150 ppm; while the upper sequence is dominated by samples with Mg# < 50, TiO_2 > 1.5 wt%, P_2O_5 > 0.15 wt%, Fe_2O_3 up to 15.7 wt% and Zr ≥ 150 ppm (Figure 4a), which are referred to as low-Ti (low TiO_2/low Zr) and high-Ti (high TiO_2/high Zr), respectively. In addition, the lower sequence typically has lower Zr/Y and ΔNb than the upper sequence, which Millett et al. (2015) interpreted as representing a significant change in the mantle melting regime feeding the lava pile. They suggest that the lower sequence was derived from large-degree partial melting of a depleted MORB-like source (mid-ocean ridge basalt), while the upper sequence was derived from smaller degrees of partial melting of a more enriched source. This decrease in the extent of melting may be related to falling mantle temperatures, or geographically distinct melting regions with different lithospheric thicknesses that fed the lava pile at different times. The distinct change in geochemistry between the two sequences coincides with the volcanic hiatus and relative subsidence event recorded by the lava delta successions, suggesting a genetic link between magmatism and basin development (Millett *et al.* 2015).

Ben Nevis

219/21-1 was drilled by Shell between August and September 2003 in order to test the Ben Nevis prospect, a large anticlinal structure at the SE limit of the Møre Marginal High. The well penetrated ~265 m of basaltic extrusives (a lower unit of ~78 m of hyaloclastite and an upper unit of ~170 m of lavas and sedimentary interbeds, separated by ~18 m of claystone). The sequence has been interpreted as a prograding lava delta which was flooded, resulting in deposition of marine claystone. This was then succeeded by subaerial volcanism (Jolley 2009). The volcanic succession is underlain by thick Cretaceous mudstones, heavily intruded by basic sills (Mclean *et al.* 2017).

Palynoflora from interbeds throughout the volcanic sequence, including *Momites* spp. And *Alnipollenites verus*, indicate that the entire succession can be attributed to sequence T36 (early Lamba Fm.; ~58.7-58.1 Ma), with no Flett Fm. volcanics present (Jolley 1997, 2009).

Erlend Volcanic Centre

The Erlend Volcanic Centre is located ~100 km due north of Shetland on the Erlend High (Figure 1). It was first interpreted as an igneous centre by Chalmers and Western (1979) based on its pronounced gravity and magnetic anomalies, and was later mapped by Gatliff et al. (1984) using 2D seismic data.

Unlike most igneous centres on the NE Atlantic Margin, Erlend has been drilled by three exploration wells. The earliest is 209/09-1, drilled between October 1979 and January 1980 on the SE flank of the edifice, where it penetrated 170 m of interbedded hyaloclastites and subaerial lavas, underlain by Cretaceous claystones atop a granitic basement high (Jolley & Bell 2002a). This was followed by 209/03-1 in August-September 1980, which was drilled ~4 km W of the volcano's crest, where it penetrated 823 m of basaltic lavas and sedimentary interbeds underlain by Cretaceous limestones and siltstones (Kanaris-Sotiriou *et al.* 1993; Jolley & Bell 2002a). 209/04-1A was drilled in September-December 1985 on the NE flank of the volcano, and encountered 475 m of hyaloclastites and lavas with siltstone and sandstone interbeds, underlain by Thanetian and Cretaceous claystones and shales (Jolley & Bell

2002a). All sedimentary units underlying the extrusive volcanics are heavily intruded by a mixture of basic and acidic igneous rocks (Kanaris-Sotiriou *et al.* 1993; Jolley & Bell 2002a).

Palynofloras from sediments within the volcanic sequence indicate a Late Paleocene-Early Eocene age (sequence T40; 56.1-55.2 Ma), a similar age to the Lagavulin sequence (Jolley & Bell 2002a; Millett *et al.* 2015).

Tobermory and Bunnehaven

Tobermory (214/04-1) and Bunnehaven (214/09-1) are located in the vicinity of the Corona High in the central Faroe-Shetland Basin. Tobermory, drilled between April and June 1999, encountered ~180 m of hyaloclastite (in which the well terminated) overlain by ~60 m of lavas and sedimentary interbeds. This sequence has been interpreted as a lava delta, thought to be equivalent in age to the Flett Fm. (end of well report). Bunnehaven was drilled between July and October 2000 and encountered ~20 m of subaerial basaltic lava flows, which are considered to be equivalent to the lavas penetrated by Tobermory. The lavas are underlain by claystones and sandstones of the Flett and Lamba Formations, which lie unconformably on a thick sequence of Jurassic-Cretaceous mudstones with minor sandstones. The well terminated in gneissic basement (end of well report).

Data and Methods

Seismic Data

This study was primarily undertaken using four 2D seismic surveys (Table 1; Figure 3) covering the study area, provided by PGS and Spectrum. Although the seismic surveys have different polarities (Table 1), the two with normal polarity were phase rotated to match those with reverse polarity to aid interpretation.

An important consideration when undertaking seismic interpretation of a thick subsurface succession is how the seismic resolution changes with depth. Vertical seismic resolution is the minimum thickness that a geological feature (e.g. a lava flow) must have to be visible as a discrete seismic event (Widess 1973; Brown 2011). Attenuation of the seismic wave results in a reduction in frequency with increasing depth, resulting in a decrease in vertical resolution. Within the volcanic pile, the average dominant frequency of the seismic data ranges from around 30 Hz near the upper surface (at depths ranging from 1.3-3.2 s TWT) to around 15 Hz near the base (depths ranging from 2.7 to 5.5 s TWT; Table 2). Using the average velocity for the volcanic succession in Lagavulin (~4.3 km/s), the average vertical seismic resolution decreases from ~36 m at the top to ~72 m at the base. This means that in the lower part of the lava pile individual features such as single lava flows will not be thick enough to produce a discrete seismic event; rather, observable seismic events will correspond to the combined response from several features.

Well Data

Well data (including composite logs and end of well reports) were utilised for 7 commercially-released wells – Lagavulin, Tobermory and Bunnehaven in the northern Faroe-Shetland Basin, 209/03-1, 209/04-1A and 209/09-1 on the Erlend High, and 219/21-1 on the Ben Nevis high (Figure 3). Although data were available for ~170 commercially-released wells in the Faroe-Shetland Basin, the majority of these lie outside the study area and therefore were not used. Checkshot and VSP data were used to tie the wells to the seismic data and allow interpretation of volcanic units away from the boreholes.

Mapping of the Volcanic Succession

The volcanic sequence was interpreted using the concept of seismic volcanostratigraphy as outlined in Planke et al. (2000), with the succession being bounded by the Top and Base Volcanic reflections. This was subdivided into several distinct packages based on the interpretation of seismic facies characteristics (Table 3) combined with petrophysical characteristics of well penetrations where available (as in Planke, 1994; Andersen et al., 2009; and Millett et al., 2015).

Top and Base Volcanics

The contrast in acoustic impedance between igneous and sedimentary lithologies results in high reflection coefficients and thus high-amplitude seismic reflections (Planke *et al.* 1999; Abdelmalak *et al.* 2016a). Across the study area the top of the volcanic sequence is therefore simple to identify (Table 3).

The base of the volcanic sequence, however, is more difficult to interpret, which has led to numerous incorrect pre-drill predictions in wells that have encountered extrusive volcanics (e.g. Lagavulin, 164/7-1 (Rockall "Dome" prospect), Brugdan, William). Difficulty in interpreting the base of the volcanic sequence along the Atlantic Margin is largely due to high acoustic impedance contrasts and compositional and textural heterogeneity within the volcanics (e.g. White et al., 2003; Nelson et al., 2009). This results in absorption and scattering of seismic energy (White *et al.* 2003; Abdelmalak *et al.* 2016b), so that the signal received from the base-volcanics is very weak and often leaves no obvious seismic reflection (Table 3), especially where the volcanics are particularly thick (>500 m). In addition, the base of the volcanic pile is often not a single surface, but comprises a transitional zone of extrusives mixed with sills and volcaniclastic sediments (Archer *et al.* 2005; Abdelmalak *et al.* 2016a). In places where there is no clear base-volcanic reflection, the shallowest appearance of sills (where visible) can be used as a rough guide to infer the location of sub-basalt sediments. Sills commonly intrude close to the sediment-basalt transition, as has been observed in wells (e.g. Brugdan, 209/03-1A, 209/04-1). In addition, the high acoustic impedance contrast between sills and their host lithology means that they typically have a distinctive seismic expression (Planke *et al.* 1999; Bell & Butcher 2002; Smallwood & Maresh

2002; Schofield *et al.* 2012, 2015; Mark *et al.* 2018). Sills and lavas, however, have a similar acoustic impedance so that where sills are intruded into lava packages they do not usually produce a significant seismic event. The visibility of sills therefore indicates that they were likely intruded into non-igneous lithologies beneath the extrusive volcanic pile. Special caution is required if substantial volcaniclastic lithologies are present, as the impedance contrast between these and sills can appear close to that of sills and clastic sediments.

Volcanic Facies

Volcanic sequences typically comprise a variety of volcanic facies (e.g. lava flows, hyaloclastites, volcaniclastics), which are related to differences in the style, environment of emplacement and effusion rate of eruptions, and which can be identified and interpreted from seismic data (Planke et al., 1999; Spitzer et al., 2008; Wright et al., 2012; Hardman et al., 2018a). In this study we focus on three main volcanic facies: tabular and compound lava flows and hyaloclastic lava deltas, each of which have distinct seismic characteristics (Table 3). Lava deltas can be distinguished from sedimentary clinoforms due to their higher amplitudes and the fact that they can commonly be observed to pass laterally and updip into sub-horizontal subaerial lava flows. Non-deltaic hyaloclastites also occur within the study area, but since they typically have a similar chaotic appearance to compound lava flows they are difficult to identify from seismic data alone, and are therefore not considered in detail (Planke *et al.* 2000; Nelson *et al.* 2009).

Lava deltas are particularly useful for palaeoenvironmental analysis, since they are a coastal feature formed by lava flowing into a body of water, and therefore give an indication of the position of palaeo-shorelines and variations in relative sea level (RSL), similar to sedimentary deltas (e.g. Fuller, 1931; Jones and Nelson, 1970; Porębski and Gradziński, 1990; Wright et al., 2012). Stationary RSL results in a purely progradational delta, which builds out from the shoreline with minimal vertical growth (Figure 5a). A fall in RSL results in a downstepping geometry, whereby the younger lava delta progrades from a position on the flanks on the original delta level with the new base level (Figure 5b). A rise in RSL can result in two very different delta geometries, depending on the supply of lava. In cases where volcanism is sufficiently voluminous to fill the newly-created accommodation, thus keeping pace with or even outpacing RSL rise, the delta will aggrade (i.e. it has significant vertical growth as well as horizontal progradation; Figure 5c). When volcanism is less voluminous, however, the rise in RSL outpaces the supply of lava, resulting in retrogradation of the shoreline and backstepping of the delta front. When RSL stop rising, the lava delta will build out from the new, higher shoreline, resulting in a progradational package on top of the original delta (Figure 5d).

Importantly, the volcanic packages identified in the Lagavulin well by Millett et al. (2015) can be shown to correspond well to seismic facies boundaries, allowing interpretation away from the wellbore (Figure 4). For the purposes of this study, three main packages are identified and referred to as V1-V3, from oldest to youngest (see Figures 4 and 6 for detail). Since the volcanic sequence in Lagavulin has been dated to sequence T40 (Flett Fm.; (~56.1-55.2 Ma), it follows that these three packages, penetrated by Lagavulin, were all erupted within sequence T40.

Thickness of volcanic units

Two-way time (TWT) thicknesses of volcanic units were converted into approximate stratigraphic thicknesses using the average seismic velocity from Lagavulin. The average velocity for the entire volcanic sequence in the well (calculated as ~4.3 km/s) was used in order to account for the likely presence of mixed volcanic and non-volcanic facies within each unit.

Uncertainty

Owing to the sparseness of well data in the study area and the spacing of the available 2D seismic lines, there is considerable uncertainty in mapping of the volcanic succession. Although Lagavulin provides an extremely detailed breakdown of the units and facies present within the volcanic pile, it is a single data point within a large area. The nearest wells (Ben Nevis, Erlend, Tobermory and Bunnehaven) penetrate much thinner volcanic successions, the ages of which are not precisely known. Combined with their distance from Lagavulin (80-100 km), this means that they are of limited use for mapping individual units across the basin, but they are useful for constraining the distribution and thickness of the volcanic pile as a whole. In addition to this, it has been shown that complexity within volcanic successions can make it difficult to correlate individual packages more than a few 10's of km, even at outcrop (Millett *et al.* 2017). This is especially true of intra-volcanic sedimentary and volcaniclastic units, which are lower amplitude than lavas and hyaloclastites and are extremely difficult to interpret for any distance away from a wellbore. Units mapped seismically are therefore likely to comprise several merged eruptive and intra-volcanic sequences. In order to deal with the uncertainties the volcanic units have been mapped using an iterative process in order to find the most sensible interpretation across the whole study area.

Seismic Interpretation of volcanic stratigraphy

Volcanic Unit 1 (V1)

Unit V1 is the lower lava delta unit identified in the Lagavulin well (Figure 4), where it comprises hyaloclastites overlain by subaerial lava flows. The top of the unit is highly weathered and oxidised (Millett *et al.* 2015).

Seismic character

The top reflection of unit V1 is typically higher-amplitude than the internal reflections of the overlying and underlying volcanics, making it relatively straightforward to interpret across the region (Figures 4 and 6). On the NE Møre Marginal High, ~150 km NE of Lagavulin, the top of V1 coincides with the regional top-volcanic reflection (Figure 7). Its base appears to coincide with the base of the volcanic pile across most of the study area, except in the vicinity of Ben Nevis, where a thick unit of older volcanics (sequence T36; Lamba Fm.; ~58.7-58.1 Ma) are observed beneath V1 (Figure 8).

Across the Faroe-Shetland Basin, the internal seismic character of unit V1 is variable. In some areas it appears largely noisy and chaotic, but in others sub-parallel, sub-horizontal reflections are observed (Figures 4 and 6). Imaging is better on the Møre Marginal High, with distinct reflections (also sub-parallel and sub-horizontal) often visible throughout the unit (Figures 6 and 8).

Numerous packages of clinoforms are visible within V1 across the study area. ~40 km NE of the Lagavulin wellbore a complex package is observed with clinoforms dipping to the NW, NE and SE, arranged in a radial pattern with the clinoforms dipping outward (Figures 9 and 10). SE-dipping clinoforms occur along the Fugloy Ridge for ~100 km south of Lagavulin. These are ~200-400 ms TWT (~430-860 m) thick, with the clinoforms typically partially overlain by sub-horizontal reflections (Figures 11 and 12). On the Lagavulin structure, close to the wellbore, a pair of adjacent packages ~300-400 ms TWT (~650-860 m) thick can be interpreted. The western package has clinoforms dipping to the SE, and the eastern package to the SW (Figure 8). A small package of apparently NW-dipping clinoforms, ~400 ms TWT (~860 m) thick, occurs on the northern edge of the Ben Nevis High (Figure 13).

Two distinct, laterally extensive packages of clinoforms also occur within V1 on the Møre Marginal High (Figure 14). The package to the north is 500 ms TWT (~1075 m) thick and ~30 km long with NE-dipping reflections; while the package to the south is ~400 ms TWT (~860 m) thick and ~20 km long with SE-dipping reflections. In both packages clinoforms are observed to connect updip to sub-parallel reflections.

Within the northern Faroe-Shetland Basin, all packages of V1 clinoforms appear to have a progradational geometry, with only occasional minor aggradation (e.g. Figures 9-12). Aggradation is much more pronounced on the Møre Marginal High, with packages commonly starting off progradational and becoming more aggradational through time (Figures 7 and 14). The amount of aggradation varies from ~400 ms TWT (~860 m) on the southern Møre Marginal High to ~800 ms TWT (~1.6 km) in the central part (Figure 14).

A dome-shaped feature, ~40 by 30 km in diameter and ~800 ms (~1720 m) high, with an irregular top surface is located ~25 km NE of Lagavulin within the packages of radially-dipping clinoforms (Figures 9 and 10). Its interior is largely chaotic, except towards the edges where distinct irregular sub-parallel layering is observed. In addition, truncated west-dipping reflections are visible on the central Møre Marginal High (Figure 7).

Distribution

Unit V1 pinches out to the east and south and is lost beneath SDR wedges to the north and west, covering an area of ~44 km^3 (Figure 15a). In the Faroe-Shetland Basin and on the Fugloy Ridge it has a typical thickness of ~500-700 ms TWT (~1.1-1.5 km), while on the Møre Marginal High it is ~800-1400 ms TWT (~1.7-3 km) thick. On the southern Møre Marginal High, ~35 km NE of Lagavulin, there is a 50 x 30 km region where unit V1 thins dramatically and may pinch out in places (Figures 6 and 16a).

Volcanic Facies Interpretation

Across the study area the volcanic facies interpreted in unit V1 differ from those identified within the Lagavulin wellbore (hyaloclastites overlain by subaerial lavas). The majority of the V1 lava field appears to be dominated by subaerial lava flows, indicated by widespread sub-parallel reflections. The coexistence of layered and chaotic seismic facies may represent a mixture of tabular and compound flows, consistent with the findings of Millett et al. (2015).

Clinoformal packages are interpreted as lava deltas comprising hyaloclastites, rather than sedimentary features, due to their high amplitudes and the fact that they commonly pass updip into sub-parallel lava flows. These deltas occur along almost the entire eastern margin of the lava field, suggesting that there was an extensive body of water present during eruption of unit V1 (Figures 15a and 17a). Previous Paleocene palaeogeographic studies of the region indicate that this may have been a major seaway, covering much of the Faroe-Shetland, Møre and Vøring Basins (Ziegler, 1982; Coward et al., 2003). Thus during eruption of V1 the northern Faroe-Shetland Basin and Møre Marginal High were largely subaerial, with deltas occurring at the coastline to the east. This coastline appears to have had a complex morphology resulting in varying orientations of the lava deltas.

The progradational to aggradational geometries of the lava deltas indicate that relative sea level (RSL) was stationary or rising through eruption of unit V1, with volcanism keeping pace with sea level rise (Figures 5a and 5c). RSL rise appears to have been greatest along the Møre Marginal High, particularly in the central part (up to ~1.6 km), resulting in more pronounced aggradation. The rate of RSL rise here also appears to have markedly increased during eruption of V1, causing the change from progradation to aggradation of lava deltas (Figures 7 and 14).

The truncation observed on the Møre Marginal High is indicative of a period of erosion resulting in an unconformity (Figure 7) and therefore our findings support a period of emergence and possible volcanic hiatus following the eruption of unit V1.

The dome-like feature near Lagavulin is likely to be a preserved volcano (Figures 9 and 17a). It is surrounded to the N, E and S by lava deltas, which all dip radially away from the edifice, suggesting that they originated from it. The deltas are overlain by layered facies that appear to be subaerial lava flows, indicating that the volcano was originally surrounded by shallow water (except to the W), but the lava deltas filled the accommodation and became emergent.

Volcanic Unit 2 (V2)

Unit V2 is the upper lava delta sequence identified in Lagavulin, where it comprises 400 m of hyaloclastites passing up into 550 m of tabular lava flows. A thin volcaniclastic interbed occurs ~200 m above the base of the sequence within the hyaloclastites (Millett *et al.* 2015).

Seismic character

V2 is typically better-imaged and less noisy than V1, so that featureless regions of seismic are rare (Figure 6). The unit is dominated by sub-parallel layering with several packages of clinoforms occurring in the eastern part of the lava field (Figure 10). SE-dipping clinoforms are observed along ~80 km of the Møre Marginal High and ~130 km of the Fugloy Ridge (Figures 12 and 18). The clinoforms are typically ~300-400 ms TWT (~650-860 m) high on the Møre Marginal High, increasing to ~600 ms TWT (~1290 m) on the Fugloy Ridge, and mostly occur above and/or adjacent to SE-dipping lava deltas interpreted in unit V1. On the northern Fugloy Ridge two distinct packages of clinoforms can be distinguished within V2, with the upper package occurring above and updip of the lower (i.e. backstepping in a landward direction; Figure 12). Clinoforms are observed on the northern and southern flanks of the Ben Nevis High, dipping to the NE and SW, respectively (Figures 11 and 13). These are ~200-300 ms TWT (430-650 m) high.

Easterly-dipping clinoforms are observed on the NE Møre Marginal High, ~400 ms (460 m) high, located on top of the NE-dipping lava delta in unit V1 (Figure 19). Clinoforms in unit V2 are commonly partially or fully overlain by sub-parallel reflections (e.g. Figures 12 and 18). With the exception of the northern Fugloy Ridge, all packages of clinoforms have a purely progradational or progradational-aggradational geometry. The amount of aggradation varies from ~100-200 ms TWT (~220-430 m) on the southern Møre Marginal High (Figure 19) to ~300-400 ms TWT (~650-860 m) on the Fugloy Ridge. The volcanic dome identified in V1 is onlapped and partially-overlain by unit V2 (Figure 9).

Distribution

Unit V2 appears to have a very similar distribution to the underlying V1 unit, covering much of the northern Faroe-Shetland Basin (Figure 15b). On the Fugloy Ridge, the top of V2 is truncated by the regional top-volcanic reflection (Figure 18). It can also be interpreted across the Møre Marginal High until it is lost beneath younger volcanics to the west and north.

Thickness is fairly consistent across the northern Faroe-Shetland Basin and southern Møre Marginal High at ~400-600 ms TWT (860-1300 m), thickening to up to ~1.1 s TWT to the south (2.35 km) before it is truncated (Figure 16b). It appears to pinch out against the underlying V1 unit on the NE Møre Marginal High (Figure 7), ~150 km NE of Lagavulin, but thickens again ~70 km further north (Figure 16b).

Volcanic facies Interpretation

Similar to unit V1, the volcanic facies interpreted regionally within V2 differ from those identified in Lagavulin (hyaloclastites overlain by subaerial lavas). The prevalence of sub-parallel reflections in unit V2 indicates that it is dominated by subaerial tabular lava flows, with lava deltas occurring near the eastern edge of the lava field along the coastline of the seaway interpreted in Coward et al. (2003), similar to V1 (Figure 17b). Most of the deltas identified in V2 appear to be continuations of those in V1 in that they develop by emplacement into the same major water way, but at a later time. The location of many V2 deltas directly above those in V1 (i.e. backstepping; Figures 5d, 9 and 12) suggests that RSL rose between emplacement of V1 and V2. The mainly progradational to aggradational geometries of the lava deltas indicate that RSL was stationary or rising (by up to 860 m) during eruption of V2 (Figures 5a and 5c). The backstepping geometry of the deltas on the northern Fugloy Ridge (Figure 12) suggests that here rising RSL temporarily outpaced volcanism (Figure 5d). Given that such backstepping does not occur elsewhere in the region, this indicates that either there was anomalously rapid RSL rise here, or that volcanics were temporarily not emplaced on this part of the Fugloy Ridge. When the emplacement of volcanics resumed, RSL had risen, and thus the second lava delta prograded over the top of the first.

The volcano interpreted in V1 appears to have become inactive prior to eruption of V2. It appears to have formed a topographic feature during emplacement of V2, resulting in an onlapping relationship between the two (Figure 9).

Volcanic Unit 3 (V3)

Unit V3 is the topmost unit within Lagavulin, where it comprises ~40 m of tabular lava, signalling the end of volcanic activity in the area. It is separated from package V2 by a ~300 m thick volcaniclastic and epiclastic sequence (Millett *et al.* 2015).

Seismic character

In the vicinity of the wellbore unit V3 is represented by a single reflection, but elsewhere it appears to be thicker and contain multiple high amplitude, continuous sub-parallel reflections (Figures 4 and 8). There are four narrow packages of clinoforms in the study area which all occur in association with lava deltas in unit V2: two SE-dipping packages along

~60 km of the Møre Marginal High and ~130 km of the Fugloy Ridge (Figures 11-13), and NE- and SW-dipping packages located on the northern and southern flanks of the Ben Nevis High (Figures 11 and 13). These packages typically appear to downstep by ~100 ms TWT (~215 m) in relation to the underlying V2 deltas (Figure 12).

Distribution

Distribution of V3 is similar to that of the underlying units, but it is more restricted on the Møre Marginal High (Figure 15c). Thickness appears to be consistent across most of the study area at ~100-200 ms TWT (~215-430 m), but thins in places (such as the Lagavulin structure) to less than 50 m (Figures 6 and 16c). It is thicker in the southern part of the study area, on the Fugloy Ridge (up to 500 ms TWT; ~1075 m) and thins markedly to the north. The unit becomes buried by younger volcanics off the NW edge of the Fugloy Ridge and Møre Marginal High (Figures 18 and 20) and is truncated on the NE Møre Marginal High (Figure 7).

Volcanic facies Interpretation

Similar to the underlying units, the volcanic facies within V3 differ regionally to those identified in the wellbore. Thickness variations suggest that unit V3 comprises multiple flows in areas away from Lagavulin. The continuous, parallel nature of the reflections indicates that the lava flows are tabular in nature. The packages of clinoforms are interpreted as lava deltas, which appear to be continuations of deltas within V2 (Figure 17c). The typical downstepping geometry of the deltas shows that RSL had fallen between emplacement of V2 and V3 (Figure 5d).

Summary of Volcanic Sequence

Across most of the northern Faroe-Shetland Basin and Fugloy Ridge the thickness of the entire volcanic sequence (i.e. units V1 to V3) is fairly consistent (Figures 6 and 16d), between ~1-1.5 s TWT (~2.15-3.2 km), thinning to zero 20-50 km east of the Fugloy Ridge and Møre Marginal High. It thickens slightly northwards onto the Møre Marginal High with an average of ~1.4 s TWT (~3 km), up to a maximum of ~2.3 s TWT (~5 km). The greatest thickening occurs ~120 km NNE of Lagavulin on the western edge of the Møre Marginal High, where the thickness approximately doubles (from ~1-2 s TWT; ~2.15-4.3 km) in a NE direction (Figures 16d and 20). The thickened volcanic strata appear confined to a curvilinear zone ~20 by 100 km, although the quality of seismic imaging means that it is uncertain whether the volcanics thicken further to the NE. Most thickening is observed in unit V1 and in younger volcanics overlying V3 (Figure 20).

Although the volcanic succession as a whole has a fairly consistent regional thickness, there are distinct difference between the individual packages. Unit V1 is much thicker in the northern part of the study area (Figures 6 and 16a), on the Møre Marginal High (~1.6 s TWT; 3.4 km), and thinner to the south (~700 ms TWT; ~1.5 km). Units V2 and V3, by contrast, are thicker in the southern part of the study area (V2: ~1 s TWT; 2.15km; V3: ~300 ms; ~650 m), and thin dramatically to the north (V2: ~500 ms; ~1.1 km; V3: ~100 ms TWT; ~215 m), even pinching out in places (Figures 6, 16b and 16c).

The rate of lava delta aggradation (and thus RSL rise) also varies between the volcanic units. Within V1, RSL rise was greatest on the Møre Marginal High (up to 1.6 km), and very minor within the Faroe-Shetland Basin. This contrasts with V2, where RSL rise appears to have been greatest on the Fugloy Ridge (up to ~860 m), and was typically less than half that on the Møre Marginal High. V3 is again different, with lava deltas recording a fall in RSL between the emplacement of V2 and V3.

Discussion

Emplacement Environments and Relative Sea Level

The volcanic units described in this study are dominated by subaerial lava flows, showing that during sequence T40 times (Flett Fm.; 56.1-55.1 Ma) the northernmost Faroe-Shetland Basin and Møre Marginal High were subaerially-exposed. The predominantly tabular nature of the lava flows indicates that they were formed by highly effusive eruptions (Table 1; Jerram, 2002; Nelson et al., 2009) which, combined with the fact that the individual volcanic units are continuous over more than 300 km (Figures 15-17), suggests that volcanism was mainly a result of fissure eruptions. Point-sourced eruptions (i.e. individual volcanic centres) appear to have been relatively uncommon.

The majority of lava deltas within all volcanic sequences are found at the eastern edge of the lava field, interpreted as prograding into a major Palaeocene-Eocene seaway. These deltas were long-lived (persisting for several hundred thousand years) and appear to have built out during multiple stages throughout development of the volcanic pile. Given the extremely rapid emplacement of the volcanic succession (more than 2 km in less than 1 Myr), the lava deltas provide an extremely high-resolution record of the movements of palaeo-shorelines, and thus changes in relative sea level, through time. It is evident that, although the overall trend is of rising RSL resulting in progradation and aggradation of the lava delta sequences, there are finer-scale variations within this trend. Within unit V1 RSL appears to have been largely stationary, although towards the later stages of emplacement on the Møre Marginal High RSL increased rapidly, resulting in significant aggradation of the lava deltas here. Given that this aggradation does not occur elsewhere, it is likely that the Møre Marginal High underwent rapid localised subsidence (of ~800-1600 m) during eruption of unit V1. Widespread backstepping of the deltas between units V1 and V2 indicates that there was a regional rise in RSL which outpaced volcanism. This change in base level was also recorded in the volcanic sequences within Lagavulin, which has been proposed to be due to transient uplift related to the proto-Iceland plume (Hartley *et al.* 2011; Millett *et al.*

2015). Unit V2 appears to have a similar pattern to V1, with lava deltas being predominantly progradational at first and becoming more aggradational through time, particularly in the southern part of the study area. This indicates that the increase in RSL was greater in the Faroe-Shetland Basin (up to ~860 m) than on the Møre Marginal High (up to ~430 m). On the northern Fugloy Ridge there is a pronounced backstep rather than aggradation of the lava delta, showing that here RSL rise outpaced volcanism. This is likely to have been due to a temporary lack of volcanism on this part of the Fugloy Ridge (though what may have caused this is not known), since such a localised rapid rise in RSL compared to nearby areas is unfeasible. Downstepping of V3 lava deltas compared to those in V2 indicates that RSL fell by ~200 m between emplacement of these two units.

According to the eustatic sea level curve of Haq et al. (1987), during sequence T40 (Flett Fm.; 56.1-55.2 Ma) sea level fell by ~40 m. The increases in RSL described above therefore cannot be attributed to variations in eustatic sea level, and must be entirely related to crustal subsidence of the northern Faroe-Shetland area (with the exception of the final fall in RSL between units V2 and V3, which can be at least partly attributed to eustatic changes). The volcanic succession therefore records a series of spatially-variable subsidence events throughout emplacement of V1 and V2. Subsidence was greatest on the Møre Marginal High during the later stages of V1, while during V2 it was greatest towards the south, in the vicinity of the Fugloy Ridge. If the volcanic sequence was older than the age we have assumed in this study (as implied by radiometric dating of the FIBG), then the emplacement of units V1-V3 may have coincided with a rise in eustatic sea level during sequence T38 (Lamba Fm.; 58.1-56.1 Ma) of ~200 m (Figure 2; Haq *et al.* 1987). However, this could not account for the large rises in RSL recorded by the lava deltas (up to 1600 m), suggesting that in any case a significant amount of subsidence must have occurred during emplacement of the volcanic pile.

With the exception of the final fall in RSL between units V2 and V3, which may be attributed to either eustatic sea level fall, tectonic uplift, or a combination of the two, the volcanic succession in the study area appears to record only subsidence throughout sequence T40 (Flett Fm.; 56.1-55.2 Ma). This contrasts with the sedimentary record in the southern part of the Faroe-Shetland Basin, which has been interpreted to show 600-1000 m of uplift during the lower part of the Flett Fm., with peak uplift occurring between 55.8 and 55.35 Ma (Shaw Champion *et al.* 2008; Hartley *et al.* 2011).

Thickness Variations

The much greater thickness of V1 on the Møre Marginal High than in the south (Figures 6 and 16a) suggests that it may have been primarily erupted from a volcanic fissure to the north. Similarly, the greater thickness of V2 and V3 on the Fugloy Ridge suggests that they may have been erupted mainly to the south (Figures 6 and 16b-c), possibly in the vicinity of the Faroe Islands. There were probably also localised eruptions of all units elsewhere along the margin.

The cause of the significant localised thickening at the western edge of the Møre Marginal High (Figures 16a, 16d and 20) is difficult to determine. The amount of thickening (>2 km) makes it unlikely that the lavas simply ponded in a topographic low, since the lavas are all subaerial and an air-filled basin of >2 km is improbable. Reflections within the wedge do not steepen in a basinward direction, suggesting that the volcanics are not simply SDRs, and the thickening occurred too rapidly (since V1-V3 were erupted in <1 Myrs) to be due to thermal subsidence, which typically takes place over hundreds of Myrs (e.g. Newman *et al.* 1999). The most likely possibility is that the thickening is fault-controlled. The fault is likely to be located just off the western edge of the Møre Marginal High, and although its exact position is obscured due to poor imaging beneath a seamount it is unlikely to be located further basinward than shown in Figure 20. In either case, the amount (>2 km) and rapidity

(<1 Myrs) of thickening indicates that this syn-volcanic faulting was significant. Fault movement was probably greatest during eruption of unit V1, since that experienced the most thickening.

Geochemical characteristics and comparison with the onshore FIBG

Examples of the facies interpreted in the study area are extensively characterized from the Faroe Islands (e.g Passey and Jolley, 2009; Jolley et al., 2012; Millett et al., 2017), with some key differences. On the Faroe Islands low-Ti lavas are interspersed with high-Ti lavas in mixed facies (simple and compound) during the latter stages of eruption e.g. the Malinstindur and Enni Formations (Millett et al., 2017). The high-Ti dominated sequences, however, comprise almost exclusively simple tabular lavas in the lower part of the sequence in the Beinisvørð Formation (Jolley et al., 2012). These observations appear to suggest that the evolution of the Lagavulin succession (where the low-Ti volcanics are found in the lower sequence, and high-Ti volcanics in the upper) was broadly reversed in terms of eruption dynamics and chemistry relative to the Faroe Islands sequence. There are two possible explanations for this. First, it is possible that the geochemical differences are related to the thickness variations and potential different eruption locations for the Lagavulin sequence. If unit V1 was mainly erupted in the vicinity of the Møre Marginal High it is conceivable that it would be extremely thin or missing from at least parts of the onshore Faroe Islands succession. In this case the Beinisvørð Formation on the Faroe Islands would be equivalent only to units V2 and V3, explaining why no low-Ti lavas are observed. There are no equivalents to the Malinstindur or Enni Formations in Lagavulin (Millett *et al.* 2015), which is why no low-Ti lavas occur towards the top of the sequence. The second possibility is that the geochemistry at Lagavulin is a local occurrence and varies laterally, which would suggest that the interpreted volcanic unitd had multiple source areas and probably multiple eruption sites, implying an important role for significant lateral magma transport in the basin

(Schofield *et al.* 2015). The volcanic sequence at Lagavulin and on the Faroe Islands may therefore simply represent coeval eruptions from separate sources.

Unconformities

Following eruption of V1 there appears to have been a period of erosion and development of an unconformity on the NE Møre Marginal High, forming the truncated reflections visible in Figure 7. Cuttings from Lagavulin also show that the top of V1 was subaerially exposed in the northern Faroe-Shetland Basin, forming reddened boles (Millett *et al.* 2015). The unconformity therefore appears to be present across much of the study area, suggesting that there was a significant period of regional intra-Flett Fm. (intra-Sequence T40; 56.1-55.2 Ma) uplift and erosion, which Millett et al. (2015) interpret as coinciding with a widespread volcanic hiatus.

Relation to Rifting

It is generally thought, based on palynological and geochemical evidence, that rifting and breakup did not begin in the Faroes region until ~ 55 Ma, with syn-breakup volcanics being erupted during sequence T45 (55.2-54.9 Ma; Naylor *et al.* 1999; Ellis *et al.* 2002; Jolley & Bell 2002b; Passey & Jolley 2009). This implies that the entire Lagavulin sequence (and its equivalents across the Faroe-Shetland Basin and Møre Marginal High), which erupted during sequence T40 (56.1-55.2 Ma), comprises only pre-breakup volcanics. The apparently syn-tectonic significant thickening of the Lagavulin sequences observed at the western edge of the Møre Marginal High (Figures 16d and 20) suggests that at least localised rifting was taking place during sequence T40. The amount of thickening (~1 s TWT; ~2.15 km) indicates that fault movement was significant and may have been part of a wider rifting event. Given the variable thickening of the volcanic units (i.e. greatest thickening in V1, and less thickening in V2 and V3), it appears that this rifting event began prior to or during V1 times and continued throughout eruption of the younger units, but with decreasing fault movement through time. Such syn-volcanic faulting is not observed elsewhere in the study area, suggesting that rifting may have begun earlier in the north than further south. This cannot be confirmed with the data available for this study, but it is broadly supported by the findings of Gernigon *et al.* (2009, 2019), who interpreted magnetic data across the Norway Basin to conclude that seafloor spreading initiated along the Aegir Ridge near the central part of the Møre Marginal High at ~53.3 Ma, before propagating regionally from NE to SW. SW propagation of the Aegir Ridge is also proposed by Ellis and Stoker (2014). Although seafloor spreading is obviously younger than rifting, it is plausible that a similar NE-SW diachroneity existed along the margin during the pre-spreading rift phase, resulting in an earlier rifting event near the Møre Marginal High.

The initiation of rifting in the central Møre Marginal High region provides a possible explanation for the presence of low-Ti volcanics (derived from large-degree melting) in the lower sequence in Lagavulin prior to such melts being available in the Faroe Islands area to the SW (Figure 4a; Millett et al., 2015). A period of rifting within sequence T40 (56.1-55.2 Ma) would have resulted in decompression melting and eruption of large volumes of MORB-like basalts, forming unit V1 (Figure 21). This rifting could also have had an influence on relative uplift during this time interval, however, given the evidence for rapid uplift and subsidence within the FSB during this time on a higher frequency than can easily be explained by rifting, any potential contribution from diachronous rifting remains poorly constrained at present (Shaw Champion *et al.* 2008; Millett *et al.* 2015; Hardman *et al.* 2018b).

Units V2 and V3 appear to have originated by different melting conditions than unit V1. The smaller degrees of melting compared to V1 may be due to a shorter melting column alongside a potential difference in source mantle lithology, most likely as a result of thicker overlying lithosphere (Figure 21). This could result either from an increase in lithospheric thickness in the region of melting between eruption of the different packages, or from each package originating from a geographically separate melting region, with differing lithospheric thicknesses. Given the relative ages of the packages, there would not be enough time for the lithosphere to "re-thicken" between the cessation of V1 volcanism and the eruption of V2/V3 lavas, as this process requires several million years rather than the available <1 Myr (Millett *et al.* 2015). The most likely explanation therefore is that units V1 and V2/V3 had geographically distinct source regions and therefore melting columns. A similar model has been proposed for the intercalation of high-Ti and low-Ti lavas in the Enni Formation in the Faroe Islands (Millett *et al.* 2017), whereby the different lavas were sourced from distinct geographically separated melting zones at the same time during locally diachronous rift segmentation. In our model, the low- Ti lavas formed in a long melting column beneath thinned lithosphere to the north, while the high-Ti lavas formed in a shorter melting column beneath thicker lithosphere to the south. This is supported by geophysical data indicating that the Faroe Islands are underlain by 35-40 km thick cratonic basement (e.g. Bott et al., 1974; Richardson et al., 1998).

The application of this model to the study area is supported by thickness variations in volcanic units across the Faroe-Shetland Basin and Møre Marginal High (Figures 6 and 16d), which indicate that unit V1 may have been primarily erupted (and probably sourced) to the north, whereas units V2 and V3 may have been mainly erupted in the vicinity of the Faroe Islands. Subsidence variations recorded by lava deltas also support the rifting model, with greatest subsidence during V1 occurring on the Møre Marginal High, compared to greater subsidence on the Fugloy Ridge during V2. An older age for the volcanism, as implied by radiometric dating, would not change the results of our mapping, but would suggest that rifting in the vicinity of the Møre Marginal High would have occurred even earlier than we have proposed, during sequence T36-38 rather than sequence T40.

Implication for Exploration on the NW Atlantic Margin

The presence of thick lava sequences continues to provide an inhibitor to exploration along the NW Atlantic Margin, both in terms of inhibiting imaging in sub-basaltic sequence and also drilling related difficulties (e.g. Gallagher & Dromgoole 2007; Varming *et al.* 2012; Millett *et al.* 2016). Several wells including Lagavulin (217/15-1z), Brugden (6104/21-1) and William (6005/13-1A) have targeted prospects beneath thick volcanic cover and failed to exit the encountered volcanic sequences (Millett et al., 2015). Thicker than anticipated volcanic sequences were encountered in each case, highlighting the challenges of both imaging and interpretation in such settings. Continued work to improve imaging through volcanic sequences, alongside the development of improved drilling technologies to allow successful targeting of sub-basaltic targets are required. Through the integration of well data, seismic-stratigraphic mapping and updated geological models for volcanic facies development, we present new findings that can be used to aid future basin modelling and exploration.

Conclusions

By combining detailed facies and geochemical analysis of the volcanic sequence penetrated by the Lagavulin well (217/15-1Z) with regional high-quality 2D seismic data, we have interpreted the volcanic stratigraphy across the northern Faroe-Shetland Basin and Møre Marginal High, allowing for important insights into the temporal and spatial evolution of the sequence T40 pre-breakup lava field.

Three key volcanic units (V1-V3, oldest-youngest) are traced away from the Lagavulin wellbore; these can be interpreted with varying degrees of confidence from the Fugloy Ridge to the central Møre Marginal High. Each unit is dominated by subaerial lava flows, with significant accumulations of hyaloclastites along the eastern edge of the lava field. Lava deltas appear to have existed along the eastern edge of the Faroe-Shetland Escarpment and Møre Marginal High for the duration of sequence T40 volcanism, due to the presence of a persistent seaway across the Faroe-Shetland and Møre Basins. Aggradation of the lava deltas indicates that a significant amount (up to 1600 m) of spatially-variable subsidence occurred during eruption and emplacement of the volcanics. Although the thickness of the volcanic pile as a whole is fairly consistent across the study area at ~1-1.5 s TWT (~2.15-3.2 km), analysis of individual units shows that V1 is thickest in the north, on the central Møre Marginal High, and thinner to the south, while V2 and V3 have the opposite pattern. Significant thickening on the central Møre Marginal High appears to have been fault-related and potentially indicates localised rifting during eruption of V1.

These units can be grouped into two geochemically distinct volcanic sequences (Millett *et al.* 2015): a low-TiO_2 sequence (V1), formed by large degrees of partial melting beneath thinned lithosphere, and an upper sequence of high-TiO_2 volcanics (V2 and V3), formed by smaller degree melting of an enriched mantle source beneath thicker lithosphere than V1. Combined with the thickness variations outlined above, this suggests that the lower sequence erupted in the vicinity of the Møre Marginal High, sourced from large-degree decompression melting in response to early rifting in the north; while the upper sequence was sourced from smaller degrees of melting beneath thicker lithosphere and erupted further to the south.

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Tables

Survey Company	Survey Name	Survey Type	Date	Shot-point Interval (m)	Streamer length (m)	Size (sq. km)	Line Spacing	Polarity
PGS	CRRG 09/10/11	2D Geo-streamer (Broadband)	Shot 2009-2011	25	8100	7,515	8km 20km	Reverse
Spectrum	NUK-98	Reprocessed 2D	Shot 1998 Reprocessed 2003	25	6000	4,080	10km 18km	Reverse
Spectrum	DGSF97	2D	Shot 1997	25	4500	4,951	5km 8km	Normal
Spectrum	SF99	Reprocessed 2D	Shot 1999 Reprocessed 2007	25	6000	1,262	19km 10km	Normal

Table 1 Acquistion and processing data for the seismic surveys used in this study. Reverse polarity = a downwards increase in acoustic impedance corresponds to a negative-amplitude reflection (shown as blue); normal polarity = a downwards increase in acoustic impedance corresponds to a positive-amplitude reflection (shown as red).

Survey	Target of interest	Depth (ms TWT)	Typical Frequency (Hz)	Resolution (m)
CRRG 09/10/11	Upper	~3000-3200	20 - 40	27 - 54
	Lower	~5000-5500	10 - 20	54 - 108
NUK-98	Upper	~3000-3300	25 - 40	27 - 43
	Lower	~4300-4600	15 - 25	43 - 72
DGSF97	Upper	~1300-1600	20 - 30	36 - 54
	Lower	~2800-3300	15 - 20	54 - 72
SF99	Upper	~1600-1900	20 - 30	36 - 54
	Lower	~2700-3000	10 - 20	54 - 108

Table 2 Calculated vertical resolutions for the upper and lower part of the volcanic pile in each seismic survey. Average frequency measured from the relevant seismic data. Velocity is assumed to be constant throughout the volcanic sequence, at 4.3 km/s (the average velocity measured from Lagavulin). Resolution is then calculated using the formula $Res = \frac{\lambda}{4} = \frac{V/F}{4}$ *where* λ *= wavelength.*

Volcanic Feature	Volcanological Description	Seismic Characteristics		Emplacement Environment	Key References
		Bounding Reflections	Internal Reflections		
Top Basalt	Surface of volcanic pile	Top: High-amplitude, positive, continuous to semi-continuous	N/A	N/A	(Planke *et al.* 1999)
Base Basalt	Base of volcanic pile	Base: Commonly low amplitude to non-existent, negative, discontinuous.	N/A	N/A	(Planke *et al.* 1999; Abdelmalak *et al.* 2016b)
Tabular Lavas	Single flow lobe, sheet-like geometry. Flows commonly stacked on top of one another, separated by inter-eruption sediments or palaeosols.	Top: High-amplitude, positive, smooth, continuous. Base: Low-amplitude, negative, commonly discontinuous and obscured.	High-amplitude, parallel to sub-parallel, sub-horizontal	Subaerial, commonly thought to result from high-effusion eruptions	(Jerram 2002; Nelson *et al.* 2009)
Compound Lavas	Flows comprise numerous anastomosing lobes of pahoehoe lava.	Top: High-amplitude, positive, disrupted. Base: Low-ampliude, negative, , often obscured	Often intermediate to low-amplitude, often disrupted with chaotic appearance	Subaerial, thought to result from lower-effusion eruptions	(Jerram 2002; Nelson *et al.* 2009; Hardman *et al.* 2018a)
Lava Delta	Hyaloclastite (volcanic breccia) and fragmented quenched lava with steeply-dipping (up to 35°) foresets. Often capped by subaerial lava flows.	Top: Positive, intermediate to high-amplitude. Often truncates underlying reflections. Base: Low-amplitude, negative. Overlying reflections typically truncated or terminate.	Progradational clinoforms, which often connect updip to sub-parallel reflections	Coastal, formed when lava flows into a body of water, quenches and fragments. Over time and with successive eruptions the delta builds out away from the shoreline.	(Moore *et al.* 1973; Planke *et al.* 1999; Nelson *et al.* 2009; Wright *et al.* 2012; Watton *et al.* 2014)
SDRs	Basaltic lava flows basalts	Top: Typically high-amplitude, positive, smooth. Base: Seldom observed	Dip and diverge in a seawards direction, often arcuate, steepening with depth	Subaerial, emplaced in a subsiding basin	(Gibson & Love 1989; Planke *et al.* 1999, 2000; Pindell *et al.* 2014; Paton *et al.* 2017)
Edifice	Point-sourced lava flows and volcaniclastics, built up and away from a central conduit to form a mound.	Top: High-amplitude, positive, typically dome-shaped. Base: Typically low-amplitude, often obscured.	1. Reflections that dip radially away from the crest and converge at the base of the edifice. 2. Extremely chaotic, often no visible reflections.	Variable	(Calvès *et al.* 2011; Gaina *et al.* 2017)

Table 3 Volcanological and seismic characteristics of the various volcanic features and facies interpreted in the study area

Figure Captions

Figure 1. Structural map of the Faroe-Shetland Basin and Møre Marginal High. Purple box shows the location of the study area. After Blystad et al. (1995) and Ritchie et al. (2011). Black lines show the locations of seismic lines throughout this paper; grey line shows the location of the regional composite line shown in Figure 5. Inset is a regional map of the North Atlantic Igneous Province showing its geographical extent. Purple box indicates the area of interest for this study. After Brooks (2011) and Wilkinson et al., (2016).

Figure 2. Chart showing variations in lithostratigraphic nomenclature across the study area, and how these relate to the T-sequences of Ebdon et al. (1995). The stratigraphy of the Faroe Island Basalt Group (highlighted in red and pink) is shown in the breakout. Lithostratigraphy taken from: Norwegian North Sea – Halland et al. (2014b); UK North Sea – Jones et al. (2003); Champion et al. (2008); Faroe-Shetland Basin – Ritchie et al. (2011). FIBG from Jolley (2009). Eustatic sea level curve taken from Haq et al. (1987).

Figure 3. Map of the study area showing wells within the Faroe-Shetland Basin and 2D seismic surveys utilised in this study.

Figure 4. Seismic line CRRG11-1074 across the Faroe-Shetland Basin. Interpreted section shows the volcanic sequences identified by Millet et al. (2015) and this study; these sequences are here referred to as V1 (oldest) to V3 (youngest). In the wellbore, V1 and V2 comprise sequences of hyaloclastites passing up into subaerial lava flows, indicative of lava delta progradation; and V3 consists of a single large lava flow. Inset box shows the location of fig. 4a. See Figure 1 for line location.

Figure 4a. Geochemical data from the Lagavulin well, as published in Millet et al (2015), compared to the seismic interpretation. There is a distinct change from lower values of TiO2 and Zr in the lower half of the well (sequences V1) to higher values in the upper part (sequences V2 and V3).

Figure 5. Schematic diagrams showing the response of lava deltas to changes in relative sea level (RSL). Stationary RSL results in a purely progradational geometry of the delta. A fall in SL results in downstepping of the delta front. A rise in RSL, where the rate of volcanism is sufficient to keep pace with the rate of RSL rise, results in an aggradational geometry. A rise in RSL that is rapid enough to outpace volcanism results in backstepping of the delta front, followed by progradation of a second delta over the top of the first.

Figure 6. Regional seismic line showing thickness variations in volcanic units V1-V3 from south to north across the Faroe-Shetland Basin and Møre Marginal High. Unit V1 is thicker in the northern part of the study area and thins southwards, while units V2 and V3 show the opposite trend. See Figure 1 for line location.

Composite line comprises, from left to right: (Spectrum) DGSF97-005; (PGS) CRRG11-1032, CRRG09-203, CRRG11-1045, CRRG10-2035, CRRG11-1057, CRRG10-2037, CRRG11-1074, CRRG09-205, CRRG11-1115, CRRG10-206B, CRRG09-118.

Figure 7. Seismic line CRRG09-118 across the eastern edge of the Møre Marginal High, showing a long-lived lava delta sequence within unit V1. The delta changes from purely progradational on the

left to aggradational on the right. An angular unconformity is clearly visible at the top of unit V1, resulting in truncation of volcanic reflections. Units V2-V3 and younger onlap this surface, thinning and pinching out against unit V1. See Figure 1 for line location.

Figure 8. Seismic line CRRG11-1085 across the Faroe-Shetland Basin from Lagavulin (217/15-1Z) to Ben Nevis (219/21-1), showing the interpretation of volcanic units V1-V3 from Lagavulin across the basin, plus broad sub-volcanic structure and stratigraphy. The line is slightly offset from both wells (Lagavulin is located 15 km to the SE, and Ben Nevis is 8 km to the SE). A significant wedge of pre-Lagavulin volcanics (dated at Lamba Fm., sequence T36, by Jolley (2009) occur to the NW of Brendan's Dome. See Figure 1 for line location.

Figure 9. Seismic line CRRG09-205, showing interpretation of volcanic stratigraphy. A dome-shaped feature with a chaotic interior is interpreted as a volcanic edifice, and several packages of lava delta clinoforms are observed. See Figure 1 for line location.

Figure 10. Seismic line CRRG11-1095, showing interpretation of volcanic units V1 to V3 plus an underlying package of older volcanics. NE-prograding lava deltas are present in all three volcanic units. Note downstepping of V3 compared to V2. See Figure 1 for line location.

Figure 11. Seismic line CRRG11-1067, showing interpretation of units V1 to V3 plus an underlying package of older volcanics. Lava deltas prograding towards each other (to the SE and to the SW) are visible. The SW-prograding deltas, present only within V2 and V3, originate from the Ben Nevis High to the east. See Figure 1 for line location.

Figure 12. Seismic line CRRG11-2000 on the northern end of the Fugloy Ridge, showing interpreted volcanic stratigraphy. Lava deltas are present within all three units. Note the significant backstepping between the two deltas in unit V2, and the downstepping of V3 compared to V2. See Figure 1 for line location.

Figure 13. Seismic line CRRG10-206B in the Faroe-Shetland Basin, showing interpretation of units V1 to V3 plus an underlying package of older volcanics. A NW-prograding lava delta is present in V1, and SE- and NE-prograding deltas in V2 and V3. The northerly-prograding deltas originate from the Ben Nevis High to the south. See Figure 1 for line location.

Figure 14. Seismic line CRRG10-207 across the Møre Marginal High, showing interpreted volcanic stratigraphy and sub-volcanic structure. Two lava deltas prograding away from each other are observed in unit V1. Unit V2 pinches out against the top of V1, and V3 is missing. Note the change from progradational to aggradational geometries of the lava deltas. See Figure 1 for line location.

Figure 15. TWT maps showing the depth and distribution of packages a) V1 b) V2 and c) V3. The extent and progradation directions of lava deltas are also shown, along with the southern limit of SDR packages.

Figure 16. TWT thickness maps showing the thickness of a) V1 b) V2 c) V3 and d) the entire volcanic succession. The extent and progradation directions of lava deltas are also shown, along with the southern limit of SDR packages.

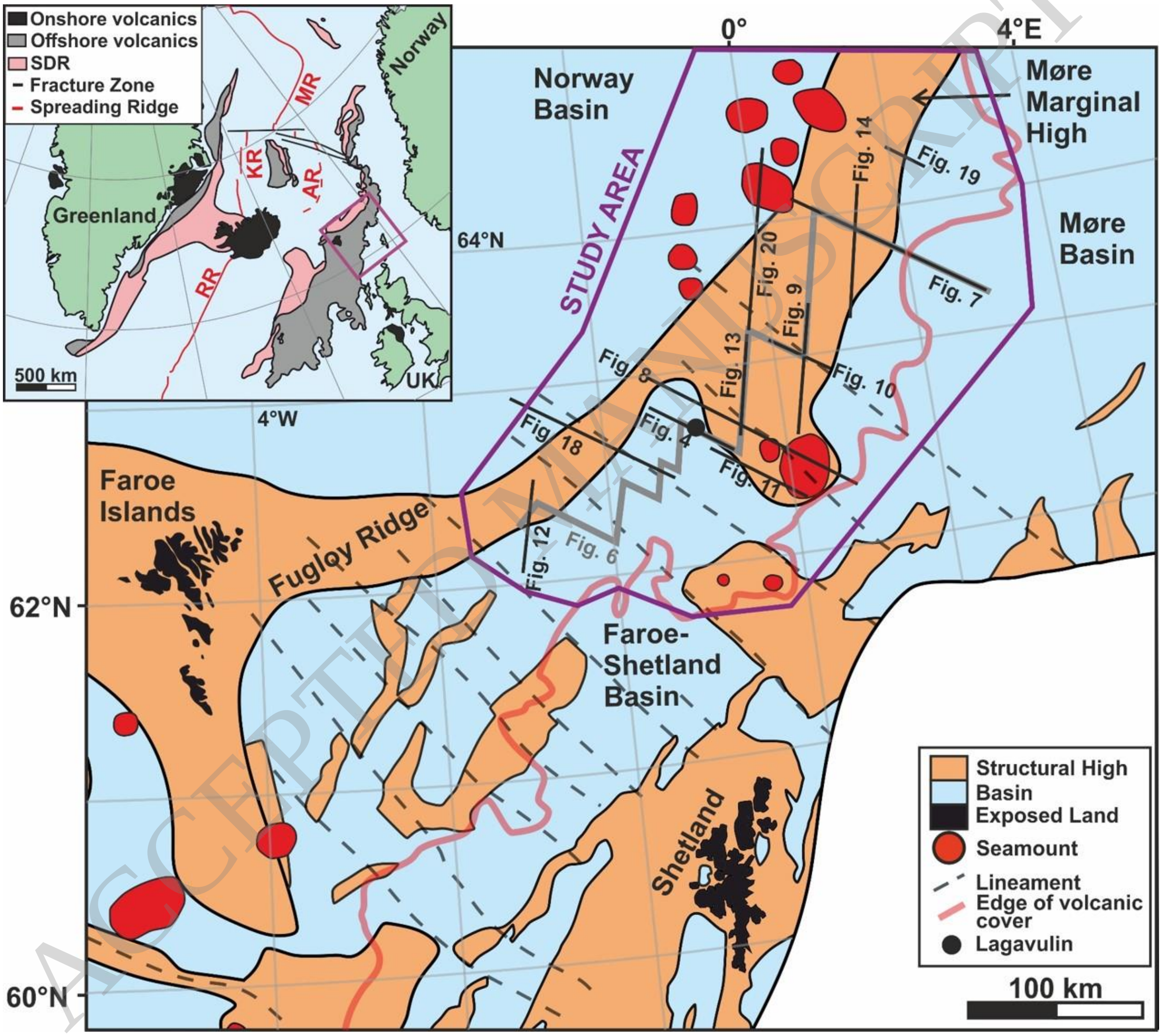
Figure 17. Block diagrams illustrating the interpreted extents and facies distributions of the volcanic sequences identified in Lagavulin. Base map adapted from Mudge (2014). a) V1 b) V2 c) V3. Inset map shows the location of the block diagrams (purple box).

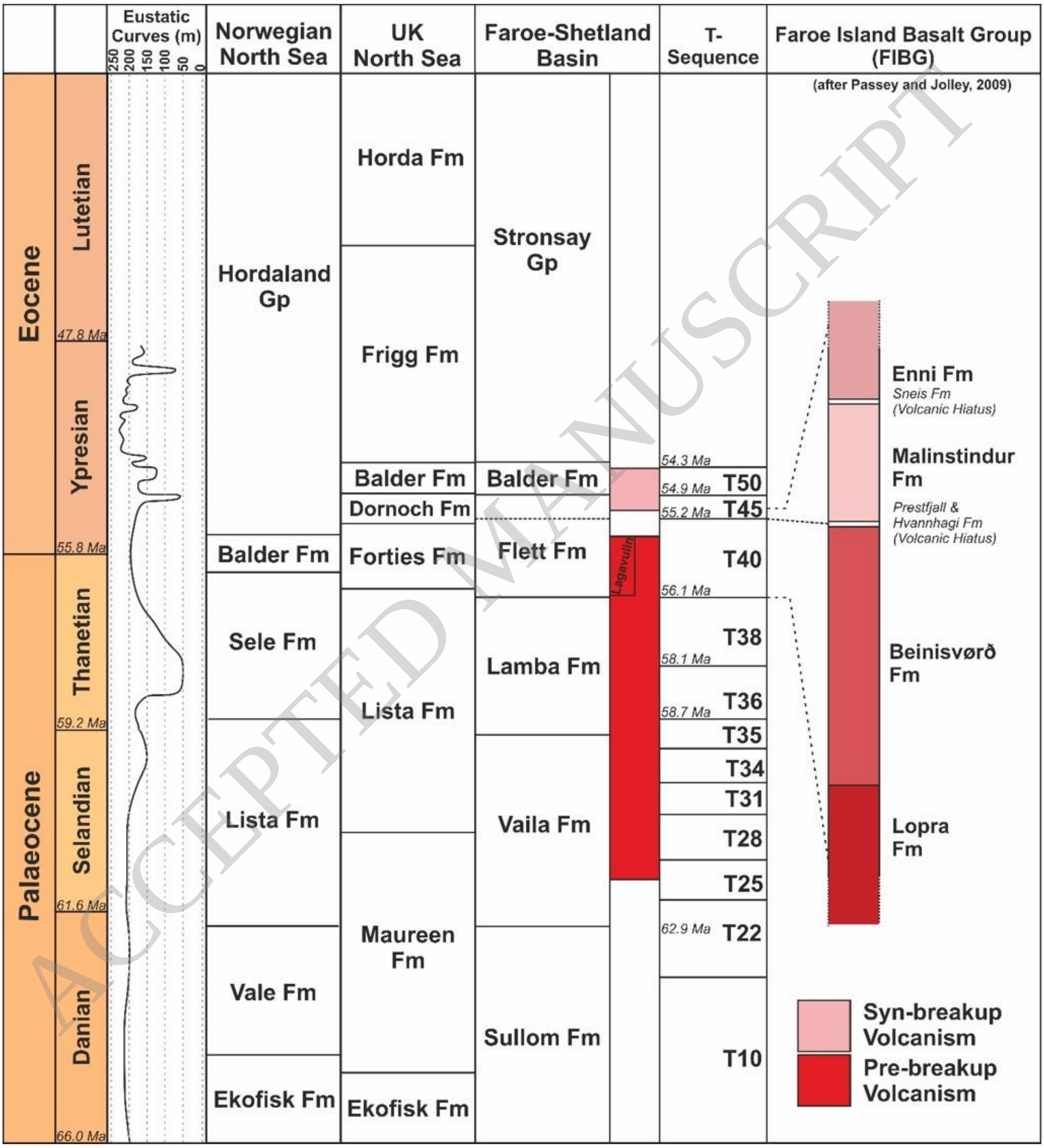
Figure 18. Seismic line CRRG11-1057 across the northern Fugloy Ridge, showing interpreted volcanic and sub-volcanic stratigraphy. An extensive SE-prograding lava delta is observed within unit V2. Units V2 and V3 are both truncated on the west of the Fugloy Ridge. V3 is overlain by a package of younger volcanics, which thickens significant in a seawards direction. Note reflections dipping steeply in a landward direction which may represent dykes intruding the volcanic pile. See Figure 1 for line location.

Figure 19. Seismic line CRRG09-120 across the Møre Marginal High, showing interpreted volcanic stratigraphy. Note the extensive lava delta within unit V2. See Figure 1 for line location.

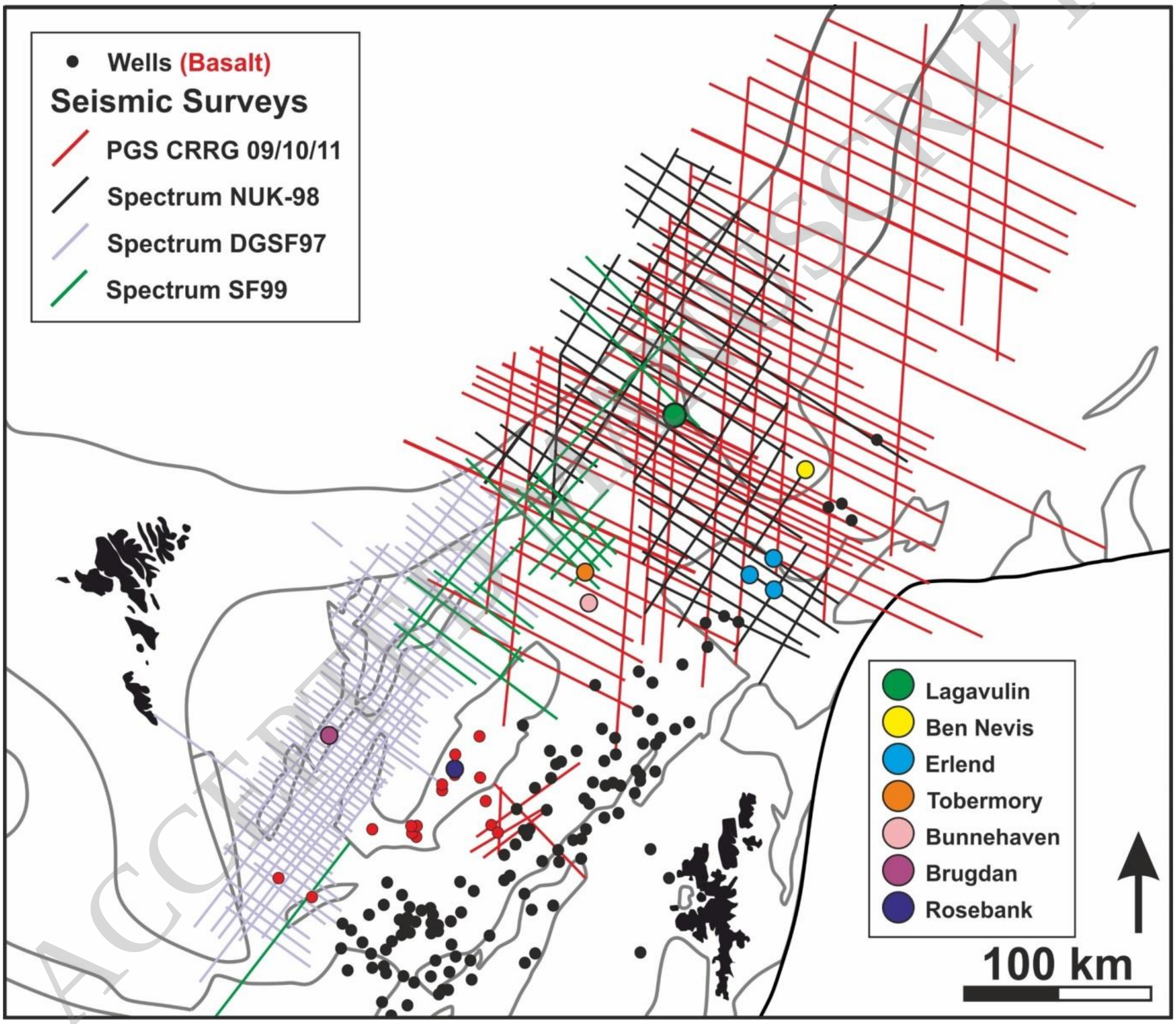
Figure 20. Seismic line CRRG09-205 across the Møre Marginal High showing interpreted volcanic stratigraphy and sub-volcanic structure. Note the significant thickening of volcanic sequences towards the north, where a syn-volcanic fault is interpreted adjacent to a seamount that post-dates the Lagavulin sequence. The exact position of this fault is uncertain, but it is likely to be within the hatched area beneath the seamount. See Figure 1 for line location.

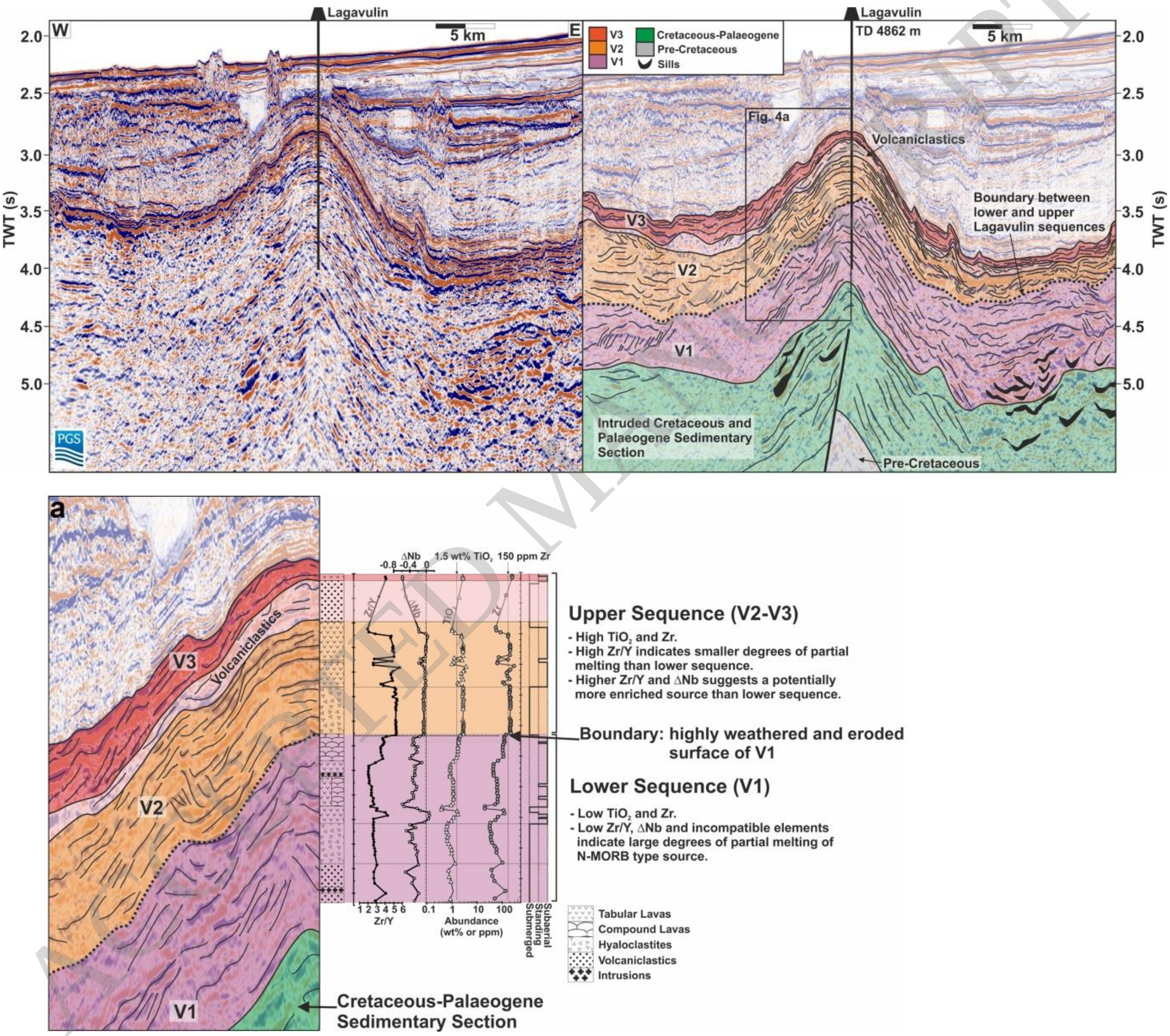
Figure 21. Simplified conceptual cross section showing possible differences in melt production between the Faroe Platform and Møre Marginal High in order to explain the variable geochemical signatures of V1 and V2-3 in Lagavulin.

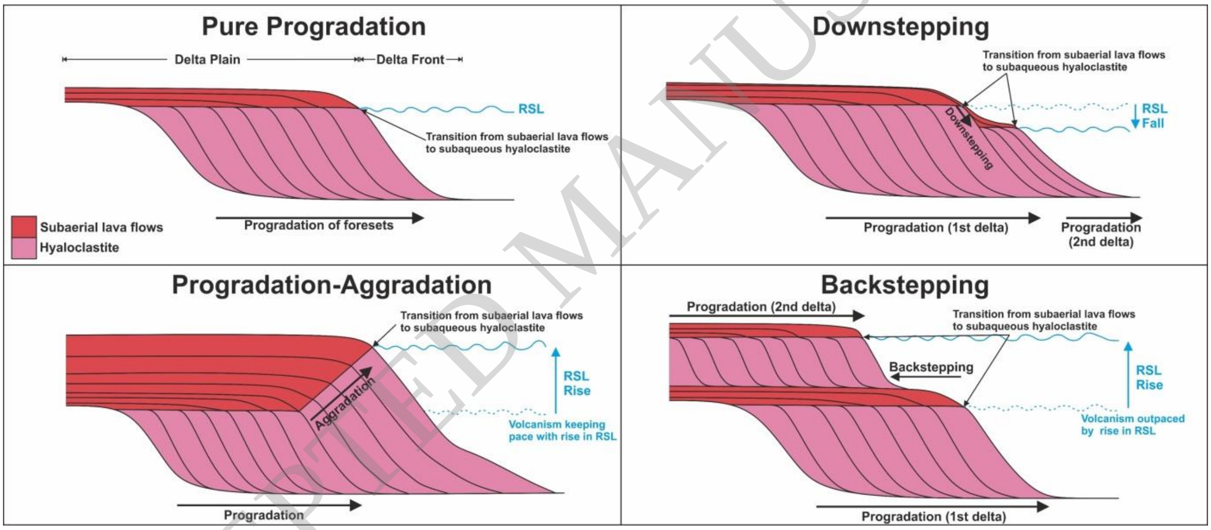

Onshore volcanics
Offshore volcanics
SDR
Fracture Zone
Spreading Ridge
Greenland
Norway
UK
MR
KR
AR
RR
500 km
0°
4°E
4°W
64°N
62°N
60°N
Norway Basin
Møre Marginal High
Møre Basin
STUDY AREA
Faroe Islands
Fugloy Ridge
Faroe-Shetland Basin
Shetland
Fig. 14
Fig. 19
Fig. 20
Fig. 9
Fig. 7
Fig. 13
Fig. 10
Fig. 8
Fig. 4
Fig. 18
Fig. 11
Fig. 12
Fig. 6
Structural High
Basin
Exposed Land
Seamount
Lineament
Edge of volcanic cover
Lagavulin
100 km

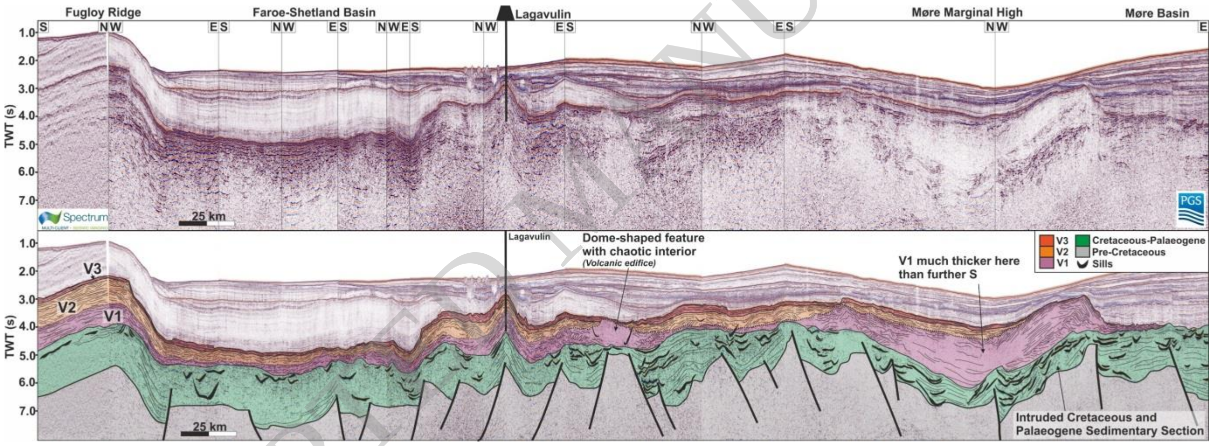


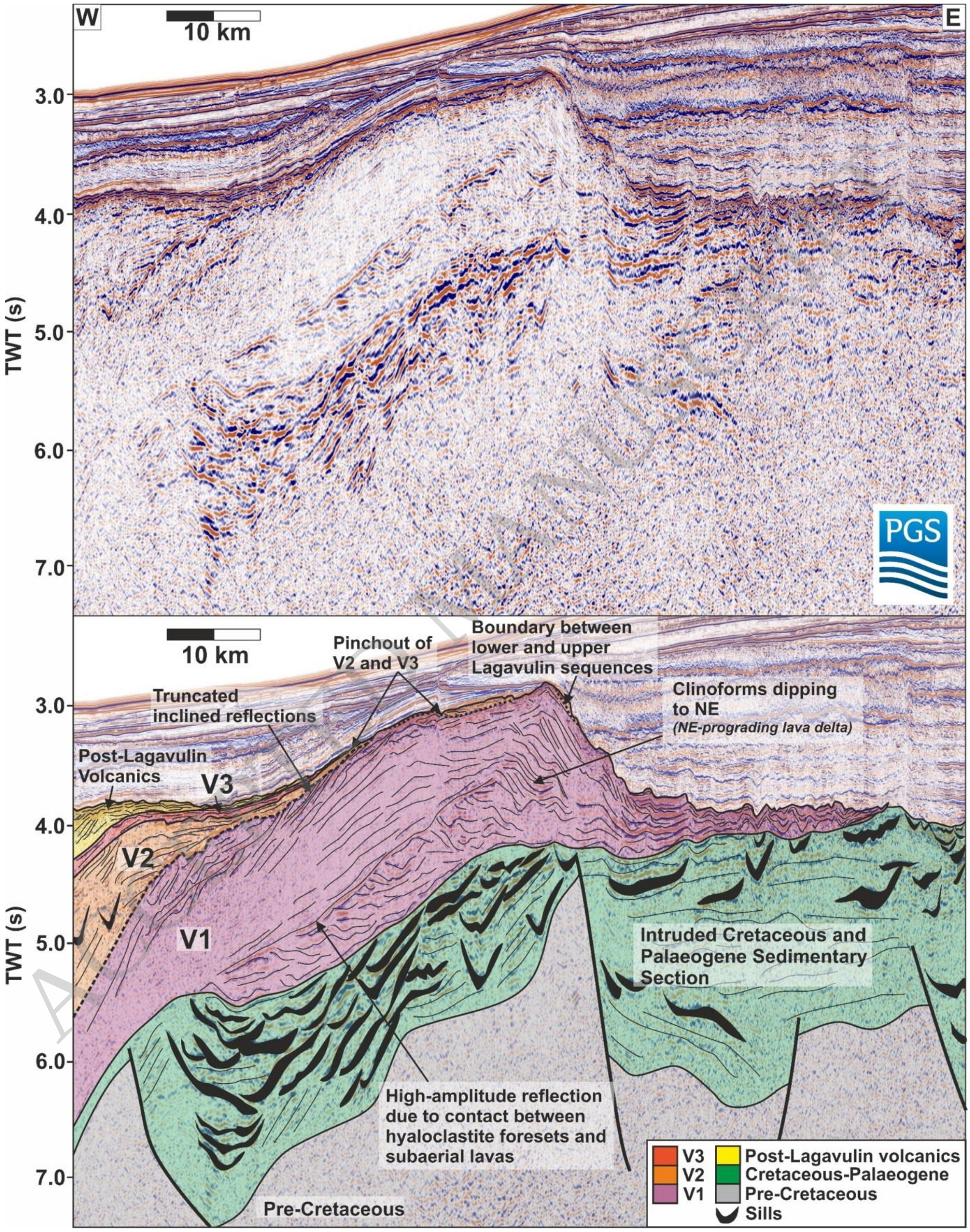
Eustatic Curves (m)
250 200 150 100 50 0
Norwegian North Sea
UK North Sea
Faroe-Shetland Basin
T-Sequence
Faroe Island Basalt Group (FIBG)
(after Passey and Jolley, 2009)
Eocene
Palaeocene
Lutetian
Ypresian
Thanetian
Selandian
Danian
47.8 Ma
55.8 Ma
59.2 Ma
61.6 Ma
66.0 Ma
Hordaland Gp
Balder Fm
Sele Fm
Lista Fm
Vale Fm
Ekofisk Fm
Horda Fm
Frigg Fm
Balder Fm
Dornoch Fm
Forties Fm
Lista Fm
Maureen Fm
Ekofisk Fm
Stronsay Gp
Balder Fm
Flett Fm
Lamba Fm
Vaila Fm
Sullom Fm
Lagavulin
54.3 Ma
54.9 Ma
T50
55.2 Ma
T45
T40
56.1 Ma
T38
58.1 Ma
58.7 Ma
T36
T35
T34
T31
T28
T25
62.9 Ma
T22
T10
Enni Fm
Sneis Fm (Volcanic Hiatus)
Malinstindur Fm
Prestfjall & Hvannhagi Fm (Volcanic Hiatus)
Beinisvørð Fm
Lopra Fm
Syn-breakup Volcanism
Pre-breakup Volcanism

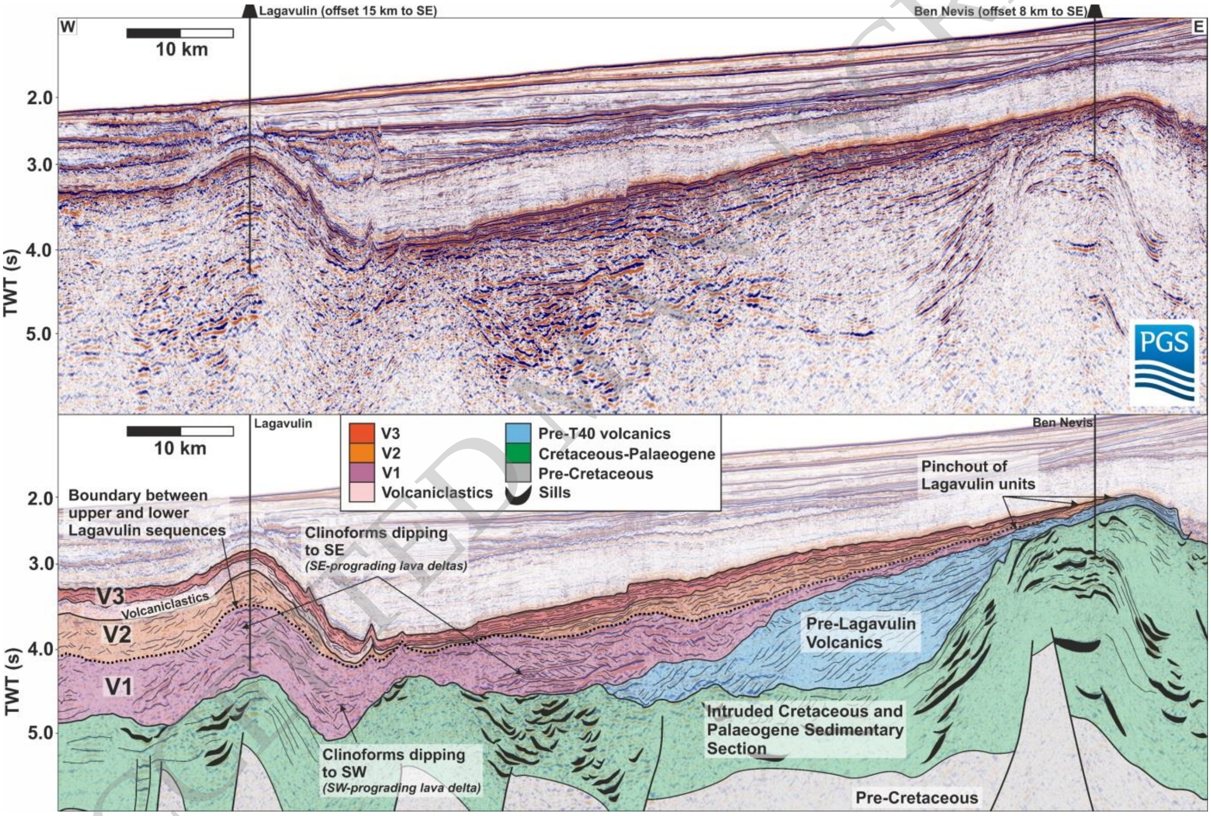

Wells (Basalt)
Seismic Surveys
PGS CRRG 09/10/11
Spectrum NUK-98
Spectrum DGSF97
Spectrum SF99
Lagavulin
Ben Nevis
Erlend
Tobermory
Bunnehaven
Brugdan
Rosebank
100 km

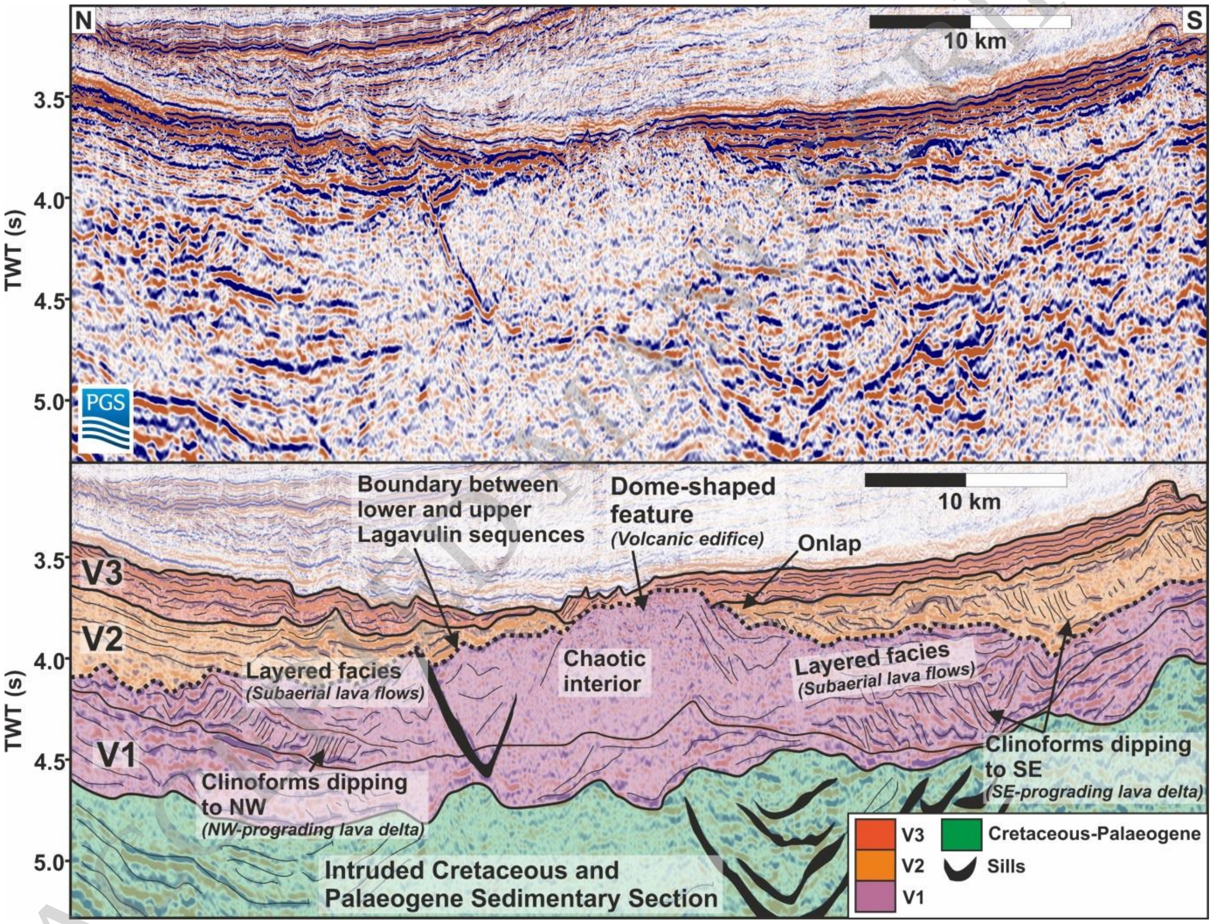

Lagavulin
W
E
5 km
V3
V2
V1
Cretaceous-Palaeogene
Pre-Cretaceous
Sills
Lagavulin
TD 4862 m
Fig. 4a
Volcaniclastics
Boundary between lower and upper Lagavulin sequences
TWT (s)
Intruded Cretaceous and Palaeogene Sedimentary Section
Pre-Cretaceous
PGS
a
Volcaniclastics
Upper Sequence (V2-V3)
- High TiO2 and Zr.
- High Zr/Y indicates smaller degrees of partial melting than lower sequence.
- Higher Zr/Y and ΔNb suggests a potentially more enriched source than lower sequence.
Boundary: highly weathered and eroded surface of V1
Lower Sequence (V1)
- Low TiO2 and Zr.
- Low Zr/Y, ΔNb and incompatible elements indicate large degrees of partial melting of N-MORB type source.
Zr/Y
Abundance (wt% or ppm)
Subaerial
Standing
Submerged
Tabular Lavas
Compound Lavas
Hyaloclastites
Volcaniclastics
Intrusions
Cretaceous-Palaeogene Sedimentary Section


Pure Progradation
Delta Plain
Delta Front
RSL
Transition from subaerial lava flows to subaqueous hyaloclastite
Subaerial lava flows
Hyaloclastite
Progradation of foresets
Downstepping
Transition from subaerial lava flows to subaqueous hyaloclastite
RSL Fall
Downstepping
Progradation (1st delta)
Progradation (2nd delta)
Progradation-Aggradation
Transition from subaerial lava flows to subaqueous hyaloclastite
RSL Rise
Aggradation
Volcanism keeping pace with rise in RSL
Progradation
Backstepping
Progradation (2nd delta)
Transition from subaerial lava flows to subaqueous hyaloclastite
Backstepping
RSL Rise
Volcanism outpaced by rise in RSL
Progradation (1st delta)

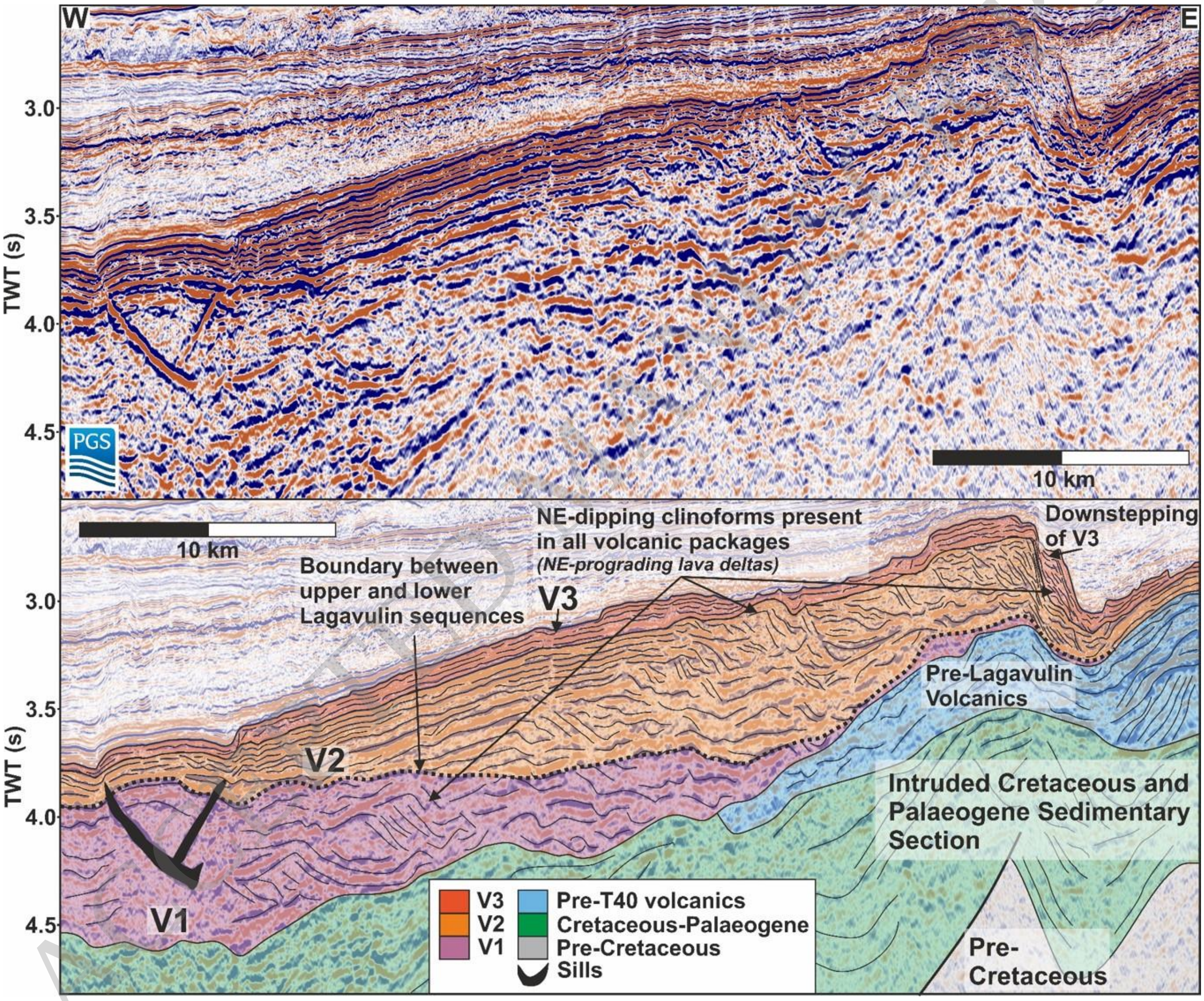

Fugloy Ridge
Faroe-Shetland Basin
Lagavulin
Møre Marginal High
Møre Basin
S
NW
ES
NW
ES
NWES
NW
ES
NW
ES
NW
E
TWT (s)
1.0
2.0
3.0
4.0
5.0
6.0
7.0
25 km
Spectrum
PGS
Lagavulin
Dome-shaped feature
with chaotic interior
(Volcanic edifice)
V1 much thicker here
than further S
V3
V2
V1
Cretaceous-Palaeogene
Pre-Cretaceous
Sills
Intruded Cretaceous and
Palaeogene Sedimentary Section

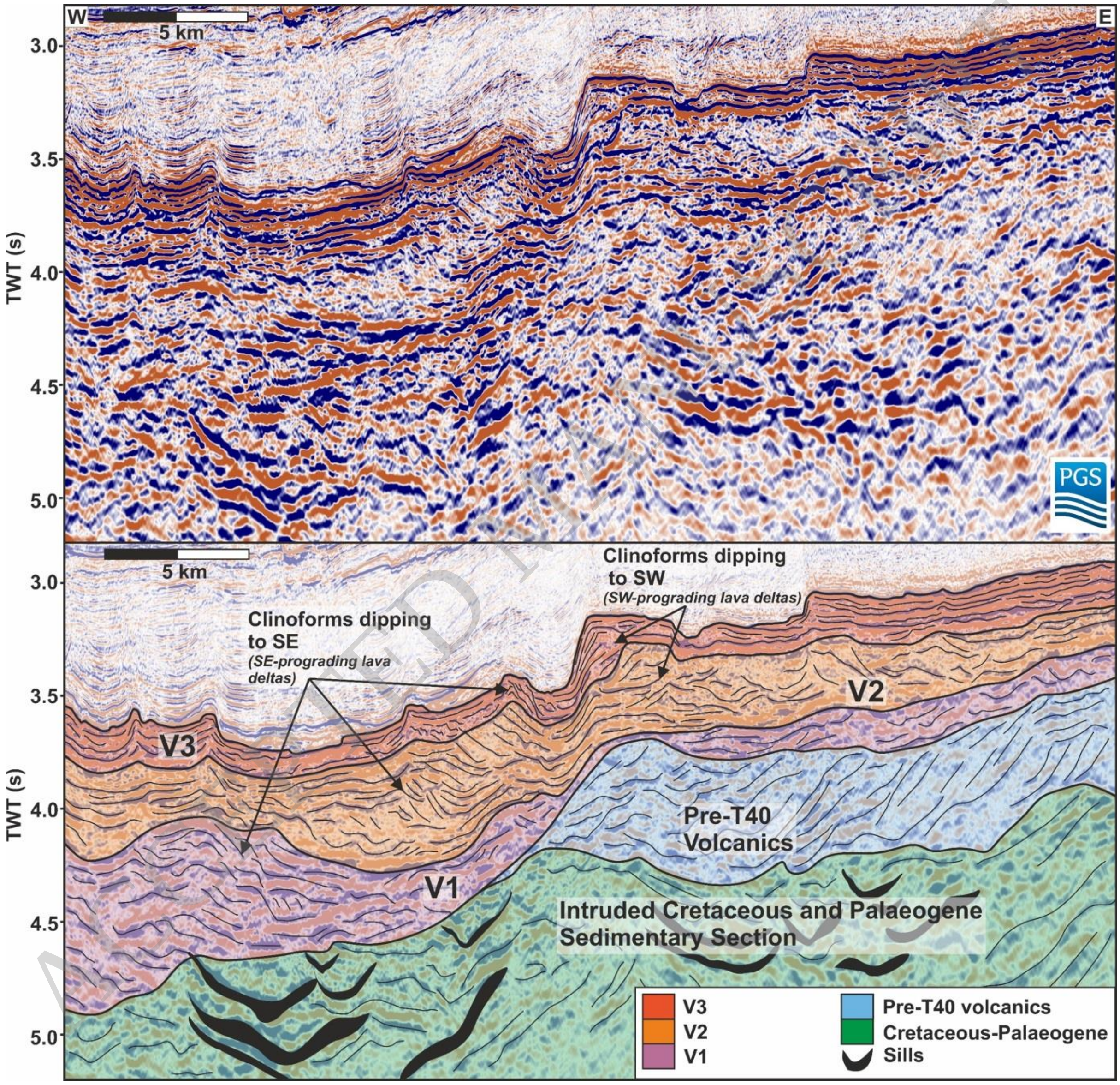

W
E
10 km
3.0
4.0
5.0
6.0
7.0
TWT (s)
PGS
10 km
Pinchout of V2 and V3
Boundary between lower and upper Lagavulin sequences
Truncated inclined reflections
Clinoforms dipping to NE
(NE-prograding lava delta)
Post-Lagavulin Volcanics
V3
V2
V1
Intruded Cretaceous and Palaeogene Sedimentary Section
High-amplitude reflection due to contact between hyaloclastite foresets and subaerial lavas
Pre-Cretaceous
V3
V2
V1
Post-Lagavulin volcanics
Cretaceous-Palaeogene
Pre-Cretaceous
Sills

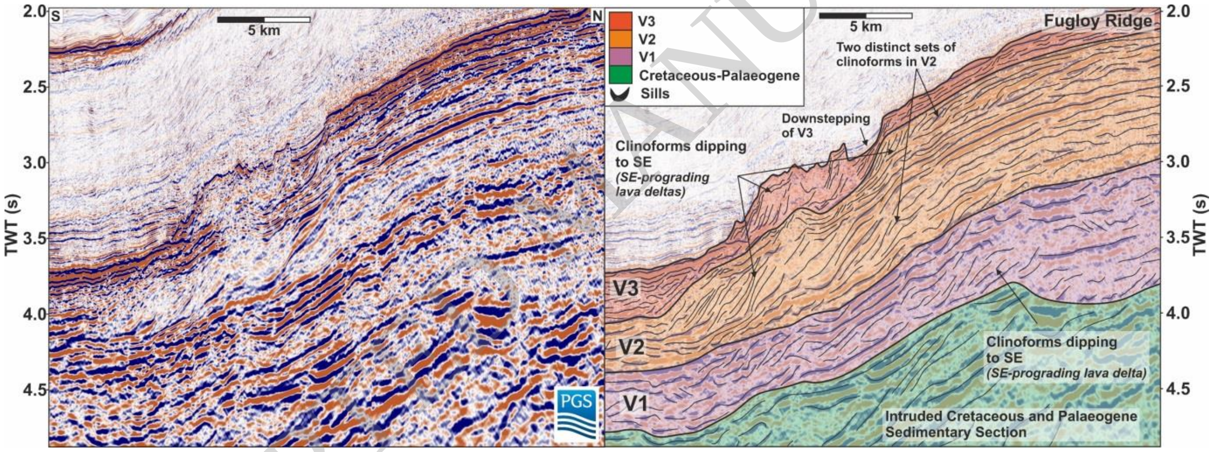

Lagavulin (offset 15 km to SE)
Ben Nevis (offset 8 km to SE)
W
E
10 km
2.0
3.0
4.0
5.0
TWT (s)
PGS
Lagavulin
Ben Nevis
V3
V2
V1
Volcaniclastics
Pre-T40 volcanics
Cretaceous-Palaeogene
Pre-Cretaceous
Sills
Boundary between upper and lower Lagavulin sequences
Clinoforms dipping to SE
(SE-prograding lava deltas)
Pinchout of Lagavulin units
Volcaniclastics
Pre-Lagavulin Volcanics
Intruded Cretaceous and Palaeogene Sedimentary Section
Clinoforms dipping to SW
(SW-prograding lava delta)
Pre-Cretaceous



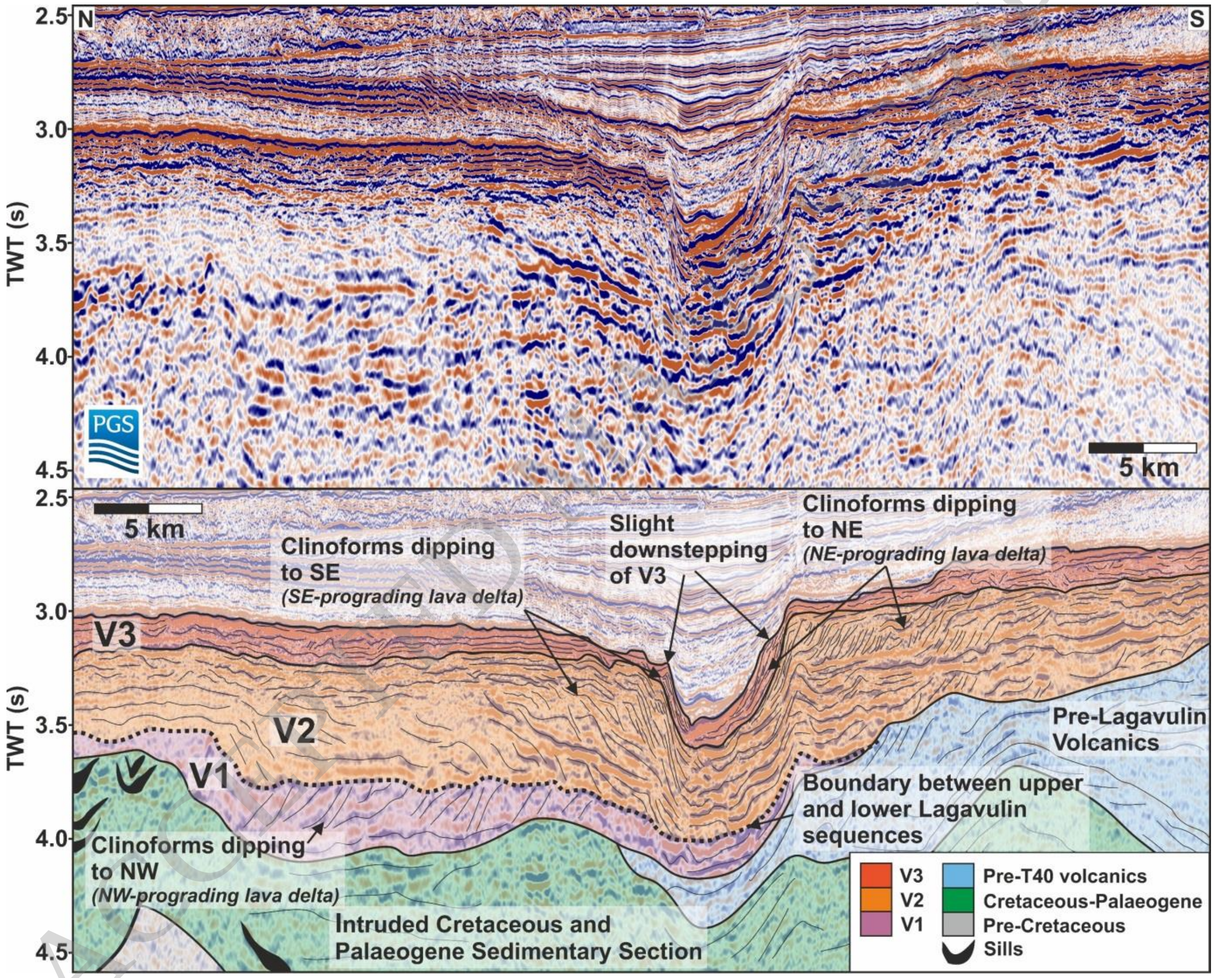
N
S
10 km
3.5
4.0
4.5
5.0
TWT (s)
PGS
Boundary between lower and upper Lagavulin sequences
Dome-shaped feature
(Volcanic edifice)
Onlap
V3
V2
V1
Layered facies
(Subaerial lava flows)
Chaotic interior
Clinoforms dipping to NW
(NW-prograding lava delta)
Clinoforms dipping to SE
(SE-prograding lava delta)
Intruded Cretaceous and Palaeogene Sedimentary Section
Cretaceous-Palaeogene
Sills


W
E
TWT (s)
3.0
3.5
4.0
4.5
PGS
10 km
10 km
NE-dipping clinoforms present
in all volcanic packages
(NE-prograding lava deltas)
Downstepping
of V3
Boundary between
upper and lower
Lagavulin sequences
V3
V2
V1
Pre-Lagavulin
Volcanics
Intruded Cretaceous and
Palaeogene Sedimentary
Section
Pre-
Cretaceous
V3
V2
V1
Pre-T40 volcanics
Cretaceous-Palaeogene
Pre-Cretaceous
Sills

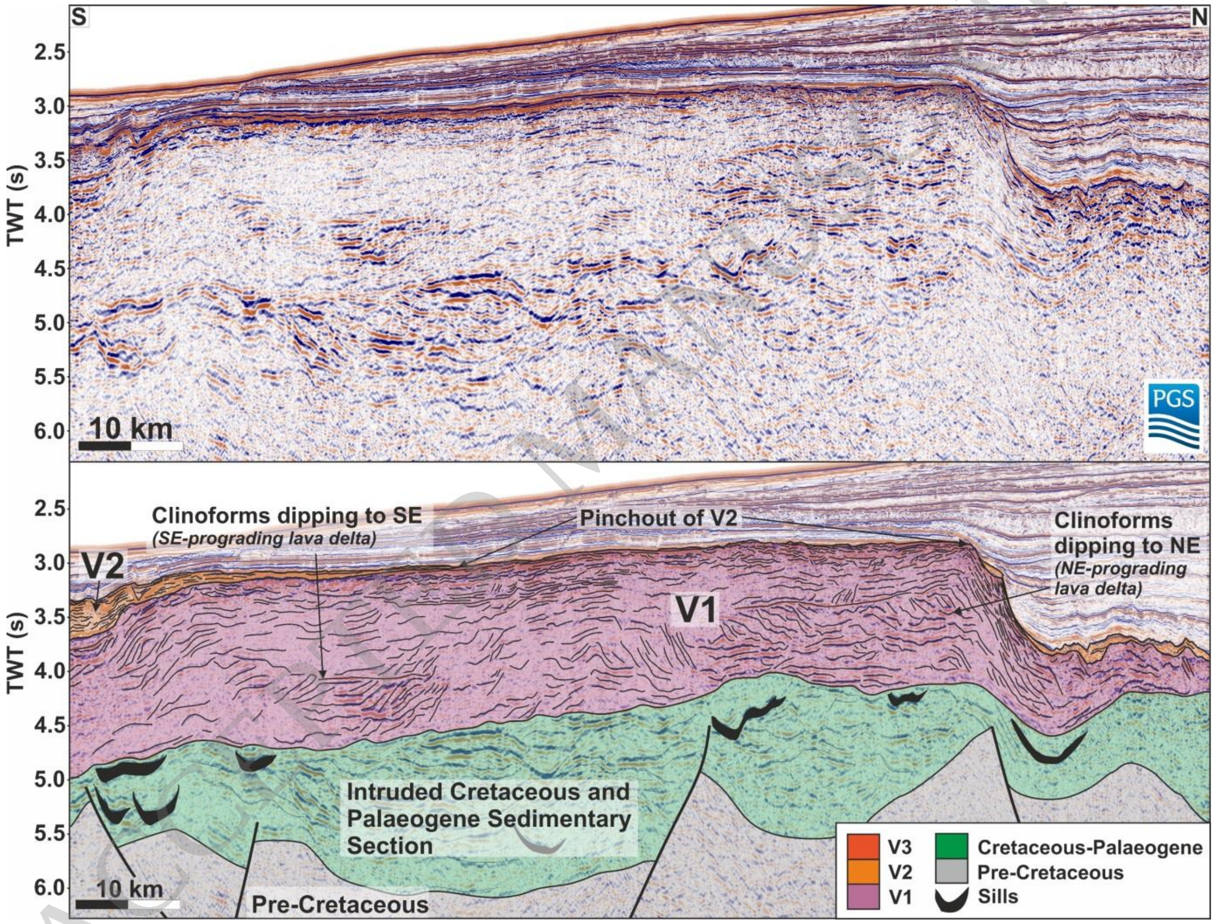

W
E
5 km
3.0
3.5
4.0
4.5
5.0
TWT (s)
PGS
5 km
Clinoforms dipping
to SW
(SW-prograding lava deltas)
Clinoforms dipping
to SE
(SE-prograding lava
deltas)
V3
V2
V1
Pre-T40
Volcanics
Intruded Cretaceous and Palaeogene
Sedimentary Section
V3
V2
V1
Pre-T40 volcanics
Cretaceous-Palaeogene
Sills

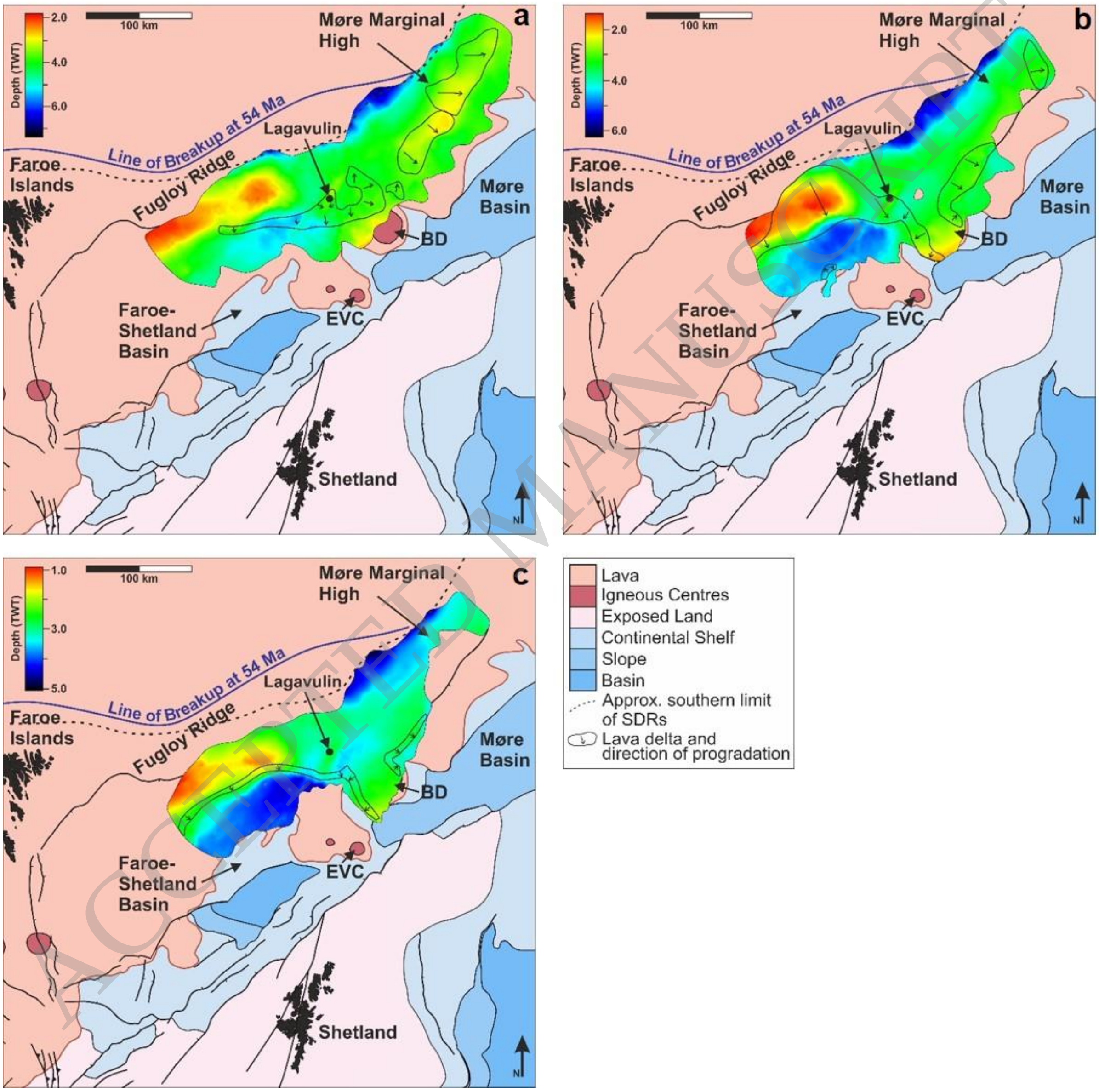


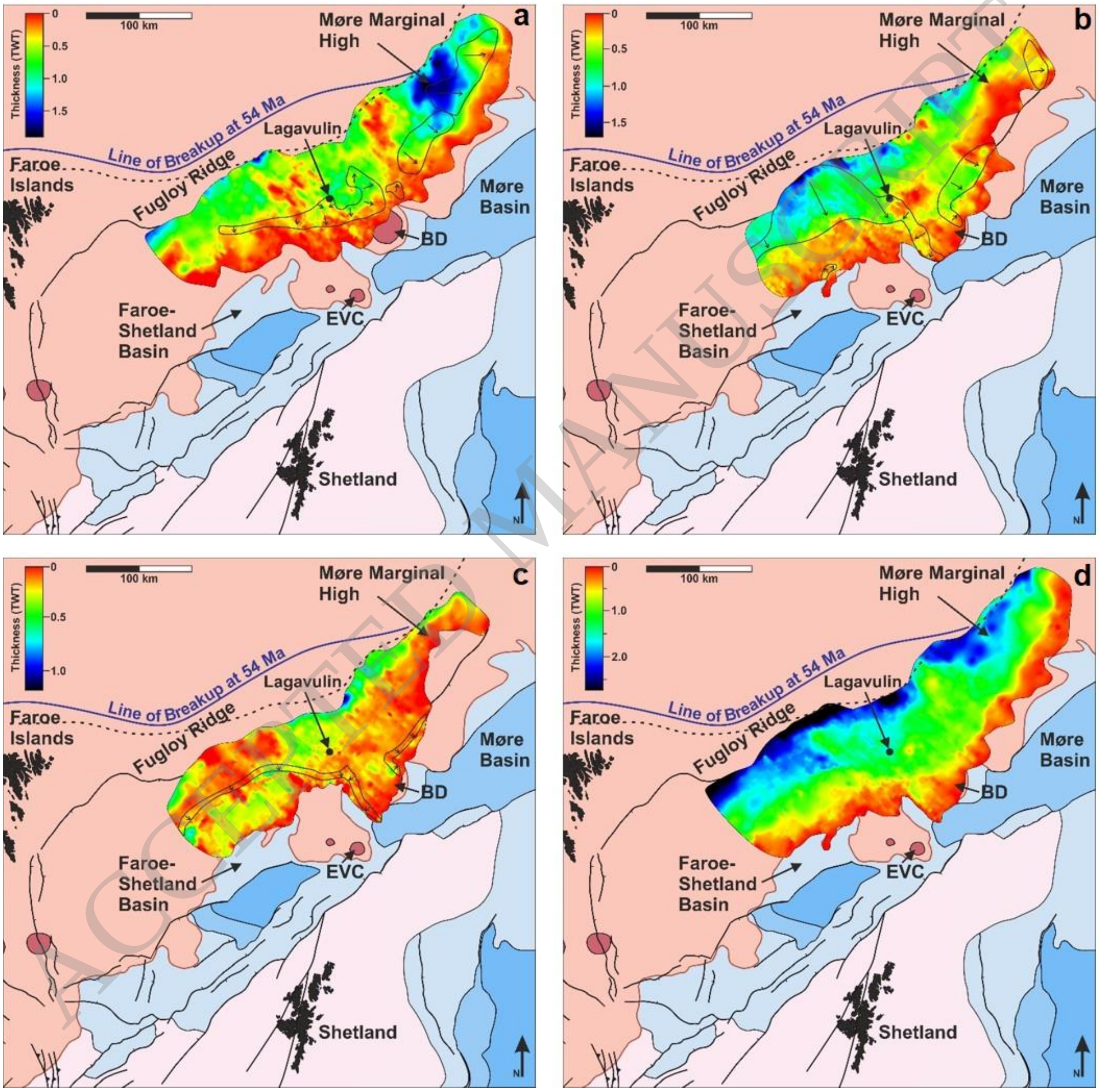
S
N
5 km
2.0
2.5
3.0
3.5
4.0
4.5
TWT (s)
PGS
V3
V2
V1
Cretaceous-Palaeogene
Sills
Fugloy Ridge
Two distinct sets of clinoforms in V2
Downstepping of V3
Clinoforms dipping to SE
(SE-prograding lava deltas)
Clinoforms dipping to SE
(SE-prograding lava delta)
Intruded Cretaceous and Palaeogene Sedimentary Section

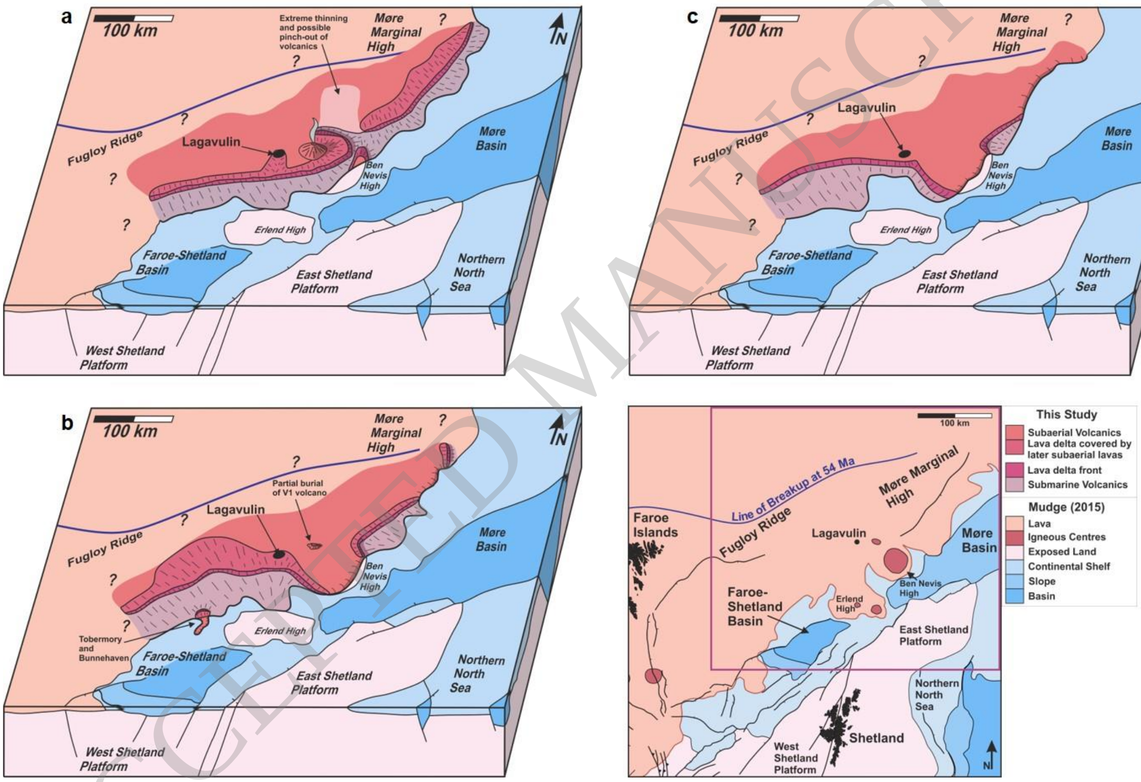


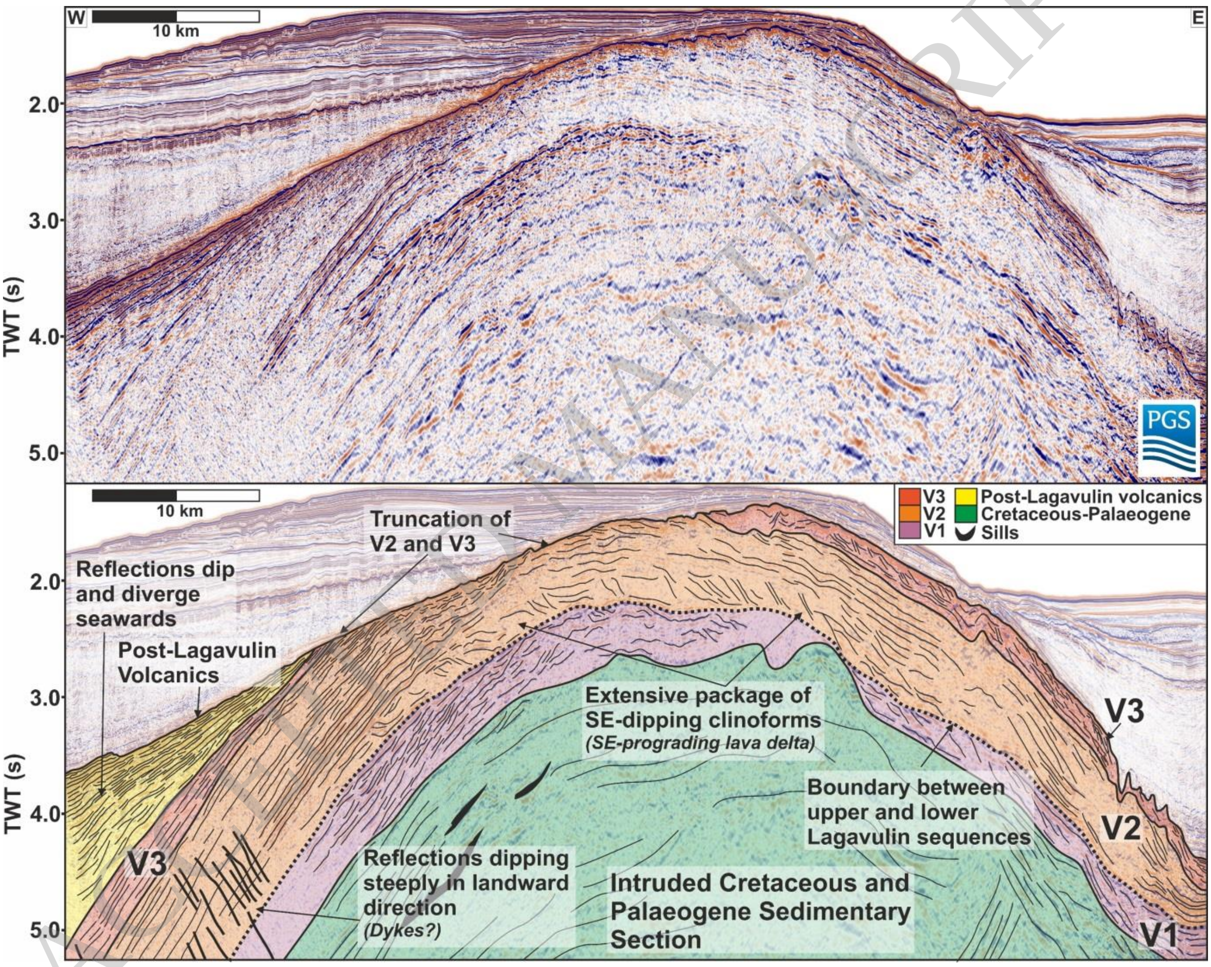
TWT (s)
2.5
3.0
3.5
4.0
4.5
N
S
PGS
5 km
Clinoforms dipping to SE
(SE-prograding lava delta)
Slight downstepping of V3
Clinoforms dipping to NE
(NE-prograding lava delta)
V3
V2
V1
Pre-Lagavulin Volcanics
Boundary between upper and lower Lagavulin sequences
Clinoforms dipping to NW
(NW-prograding lava delta)
Intruded Cretaceous and Palaeogene Sedimentary Section
Pre-T40 volcanics
Cretaceous-Palaeogene
Pre-Cretaceous
Sills

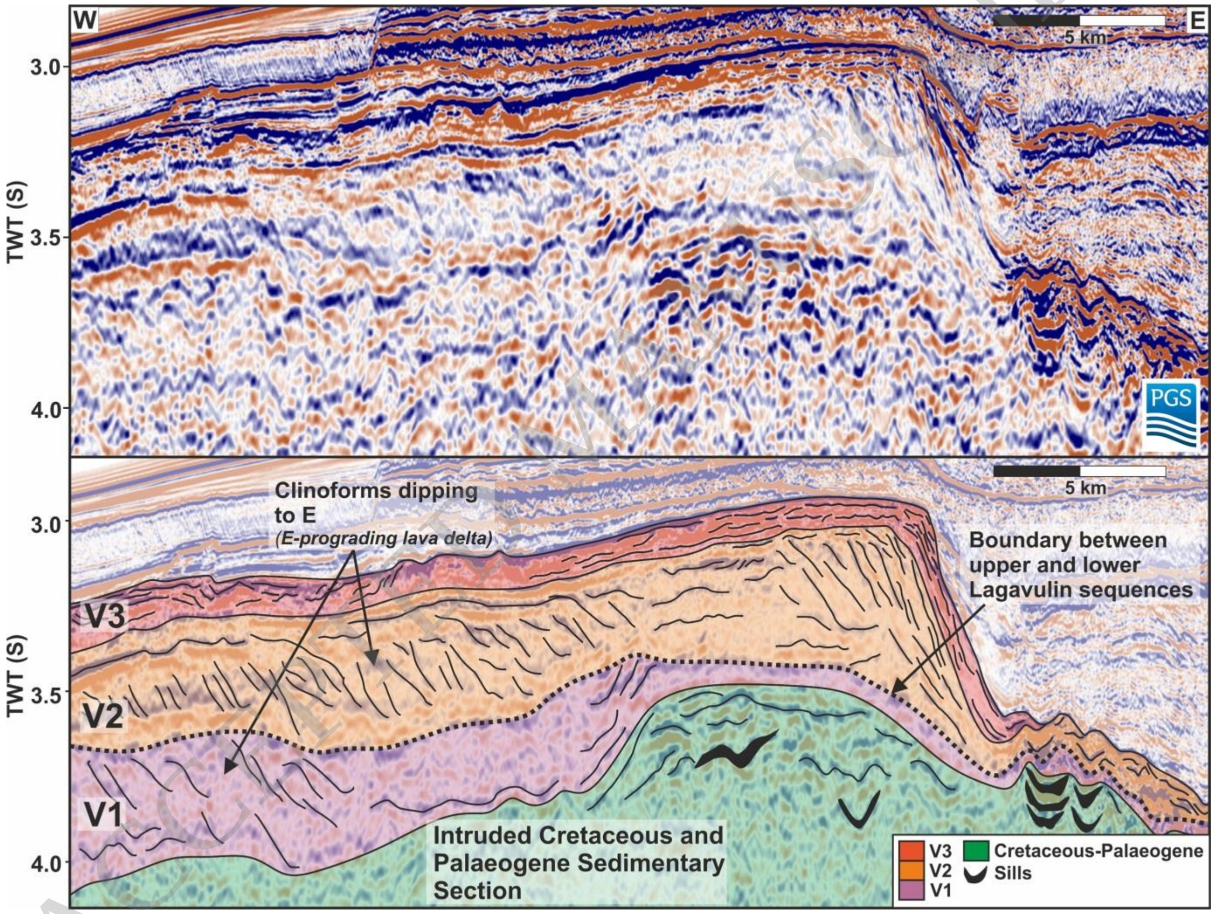

S
N
2.5
3.0
3.5
4.0
4.5
5.0
5.5
6.0
TWT (s)
10 km
PGS
Clinoforms dipping to SE
(SE-prograding lava delta)
Pinchout of V2
Clinoforms dipping to NE
(NE-prograding lava delta)
V2
V1
Intruded Cretaceous and Palaeogene Sedimentary Section
Pre-Cretaceous
V3
V2
V1
Cretaceous-Palaeogene
Pre-Cretaceous
Sills


a
b
c
Depth (TWT)
2.0
4.0
6.0
1.0
3.0
5.0
100 km
Møre Marginal High
Line of Breakup at 54 Ma
Lagavulin
Faroe Islands
Fugloy Ridge
Møre Basin
BD
Faroe-Shetland Basin
EVC
Shetland
N
Lava
Igneous Centres
Exposed Land
Continental Shelf
Slope
Basin
Approx. southern limit of SDRs
Lava delta and direction of progradation

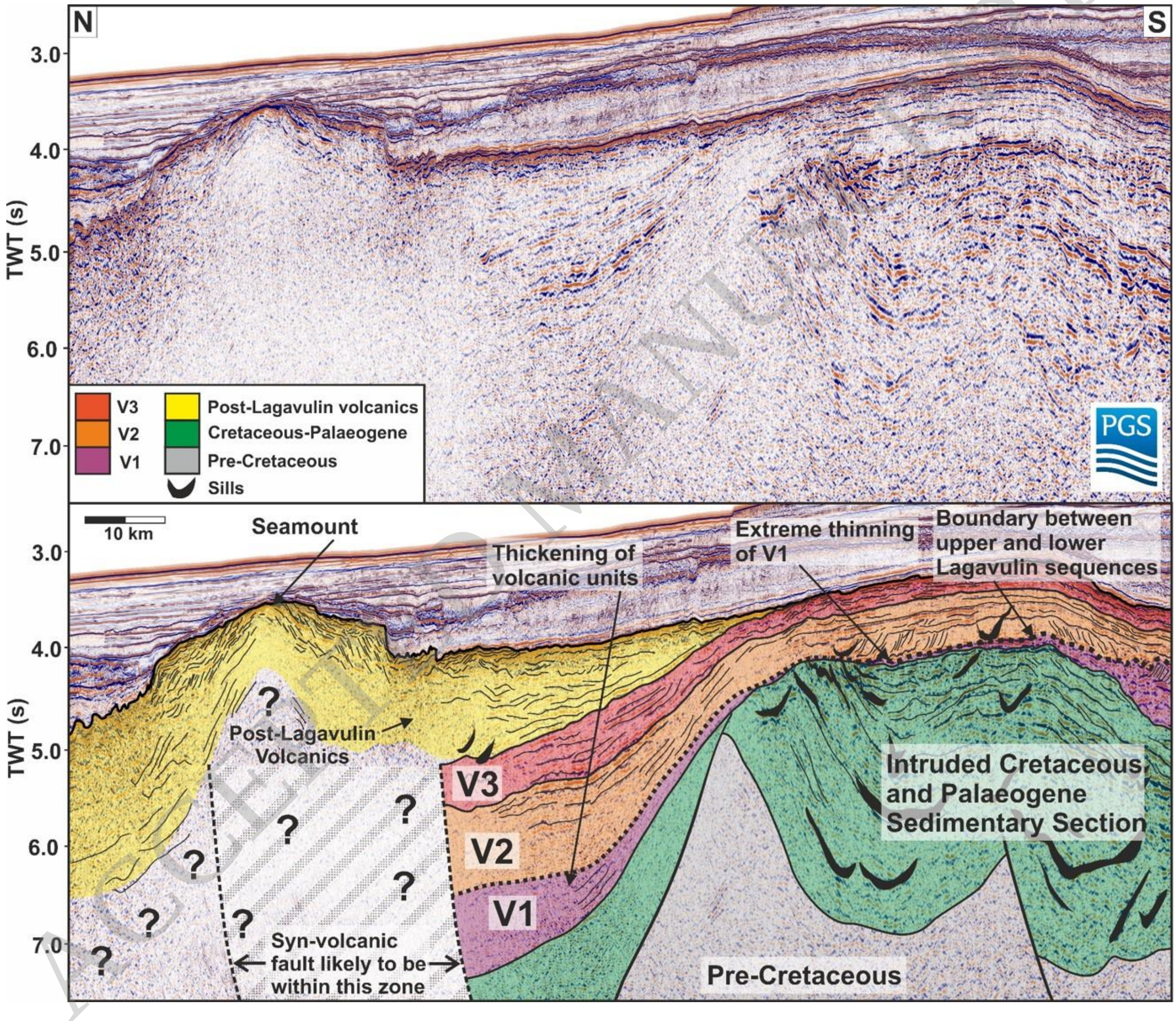

a
Thickness (TWT)
0
0.5
1.0
1.5
100 km
Møre Marginal High
Line of Breakup at 54 Ma
Lagavulin
Faroe Islands
Fugloy Ridge
Møre Basin
BD
Faroe-Shetland Basin
EVC
Shetland
N
b
Thickness (TWT)
0
0.5
1.0
1.5
100 km
Møre Marginal High
Line of Breakup at 54 Ma
Lagavulin
Faroe Islands
Fugloy Ridge
Møre Basin
BD
Faroe-Shetland Basin
EVC
Shetland
N
c
Thickness (TWT)
0
0.5
1.0
100 km
Møre Marginal High
Line of Breakup at 54 Ma
Lagavulin
Faroe Islands
Fugloy Ridge
Møre Basin
BD
Faroe-Shetland Basin
EVC
Shetland
N
d
Thickness (TWT)
0
1.0
2.0
100 km
Møre Marginal High
Line of Breakup at 54 Ma
Lagavulin
Faroe Islands
Fugloy Ridge
Møre Basin
BD
Faroe-Shetland Basin
EVC
Shetland
N

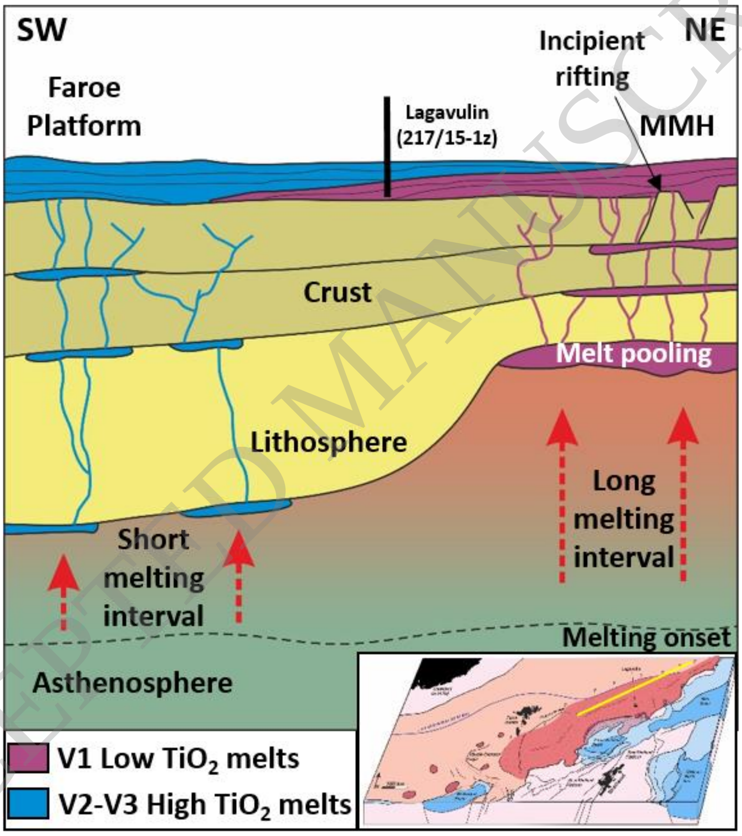

a
100 km
Extreme thinning and possible pinch-out of volcanics
Møre Marginal High
N
Fugloy Ridge
Lagavulin
Ben Nevis High
Møre Basin
Erlend High
Faroe-Shetland Basin
East Shetland Platform
Northern North Sea
West Shetland Platform
b
100 km
Møre Marginal High
N
Partial burial of V1 volcano
Lagavulin
Fugloy Ridge
Ben Nevis High
Møre Basin
Tobermory and Bunnehaven
Erlend High
Faroe-Shetland Basin
East Shetland Platform
Northern North Sea
West Shetland Platform
c
100 km
Møre Marginal High
N
Lagavulin
Fugloy Ridge
Ben Nevis High
Møre Basin
Erlend High
Faroe-Shetland Basin
East Shetland Platform
Northern North Sea
West Shetland Platform
100 km
Line of Breakup at 54 Ma
Møre Marginal High
Faroe Islands
Fugloy Ridge
Lagavulin
Møre Basin
Ben Nevis High
Faroe-Shetland Basin
Erlend High
East Shetland Platform
Northern North Sea
West Shetland Platform
Shetland
N
This Study
Subaerial Volcanics
Lava delta covered by later subaerial lavas
Lava delta front
Submarine Volcanics
Mudge (2015)
Lava
Igneous Centres
Exposed Land
Continental Shelf
Slope
Basin


W
E
10 km
2.0
3.0
4.0
5.0
TWT (s)
PGS
10 km
V3
V2
V1
Post-Lagavulin volcanics
Cretaceous-Palaeogene
Sills
Truncation of V2 and V3
Reflections dip and diverge seawards
Post-Lagavulin Volcanics
Extensive package of SE-dipping clinoforms
(SE-prograding lava delta)
Boundary between upper and lower Lagavulin sequences
Reflections dipping steeply in landward direction
(Dykes?)
Intruded Cretaceous and Palaeogene Sedimentary Section
V3
V2
V1



W
E
5 km
3.0
3.5
4.0
TWT (S)
PGS
Clinoforms dipping to E
(E-prograding lava delta)
Boundary between upper and lower Lagavulin sequences
V3
V2
V1
Intruded Cretaceous and Palaeogene Sedimentary Section
V3
V2
V1
Cretaceous-Palaeogene
Sills


N
S
3.0
4.0
TWT (s)
5.0
6.0
7.0
V3
V2
V1
Post-Lagavulin volcanics
Cretaceous-Palaeogene
Pre-Cretaceous
Sills
PGS
10 km
Seamount
Thickening of volcanic units
Extreme thinning of V1
Boundary between upper and lower Lagavulin sequences
Post-Lagavulin Volcanics
Intruded Cretaceous and Palaeogene Sedimentary Section
Syn-volcanic fault likely to be within this zone
Pre-Cretaceous


SW
NE
Faroe Platform
Lagavulin (217/15-1z)
Incipient rifting
MMH
Crust
Melt pooling
Lithosphere
Long melting interval
Short melting interval
Melting onset
Asthenosphere
V1 Low TiO_2 melts
V2-V3 High TiO_2 melts