# Seismic Interpretation of Sill-Complexes in Sedimentary Basins: The 'Sub-Sill Imaging Problem'

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**Running head:** Seismic interpretation of sill-complexes

**Supplementary materials:** None included

## Abstract

Application of 3D-seismic reflection-data to igneous systems in sedimentary basins has led to a revolution in the understanding of mafic sill-complexes. However, there is considerable uncertainty on how geometries and architecture of sill complexes within the subsurface relates those imaged in seismic reflection-data. To provide constraints on how sill complexes in seismic data should be interpreted, we present synthetic seismograms generated from a seismic-scale (22x0.25 km) outcrop in East Greenland constrained by abundant field-data.

This study highlights how overlying igneous rocks adversely affect imaging of underlying intrusions and rocks by decreasing seismic amplitude, frequency and making steeply dipping features near-impossible to image. Furthermore, seismic modelling shows that because of the high impedance contrast between siliciclastic host rock and dolerites, thin (< 5m) intrusions should in principle be imaged at relatively high amplitudes. This is contrary to many published 'rules of thumb' for seismic detectability of sill intrusions. However, actual seismic data combined with well-data shows significant amounts of un-imaged sill intrusions, and this is likely due to limited resolution, overburden complexity, poor velocity-models, and interference between closely spaced sill-splays. Significant improvements could be made by better predicting occurrence and geometry of sill intrusions and including these in velocity models.

# 1. Introduction

Igneous sill intrusions are common at volcanic rifted margins, in rifted basins and in large igneous provinces (e.g. Eldholm and Coffin, 2000; Skogseid, 2001; Bryan and Ferrari, 2013). Understanding the role of, and processes controlling, sills in such systems is essential to understand monitoring of volcanoes (Galland, 2012), understand magma-propagation in the crust and to volcanoes (Cartwright and Hansen, 2006a; Magee et al., 2013; Muirhead et al., 2016) and to understand hydrocarbon systems and plan hydrocarbon exploration in intruded basins (Rateau et al 2013; Schofield et al., 2015, 2016, 2017; Millett et al., 2016; Senger et al., 2017). Because of the high density- and velocity-contrasts between mafic intrusions and sedimentary host rock, intrusions are in principle readily imaged within seismic reflection datasets (e.g. Smallwood and Maresh, 2002). Models and understanding of igneous systems in sedimentary basins have seen significant progress following the application of 3D seismic data to study sill intrusions (e.g. Cartwright and Huuse, 2003; Thomson and Hutton, 2004; Hansen and Cartwright, 2006a; Thomson and Schofield, 2008, Schofield et al. 2012b; Magee et al., 2013; Jackson et al., 2013; Jerram and Bryan, 2015; Schofield et al. 2015). As hydrocarbon exploration moves into increasingly challenging basins, and as oil and gas fields are being targeted and discovered in the vicinity of intrusions (e.g. Tormore; Schofield et al., 2015), it is essential to have a good understanding of how real geometries of intrusions are imaged in the subsurface.

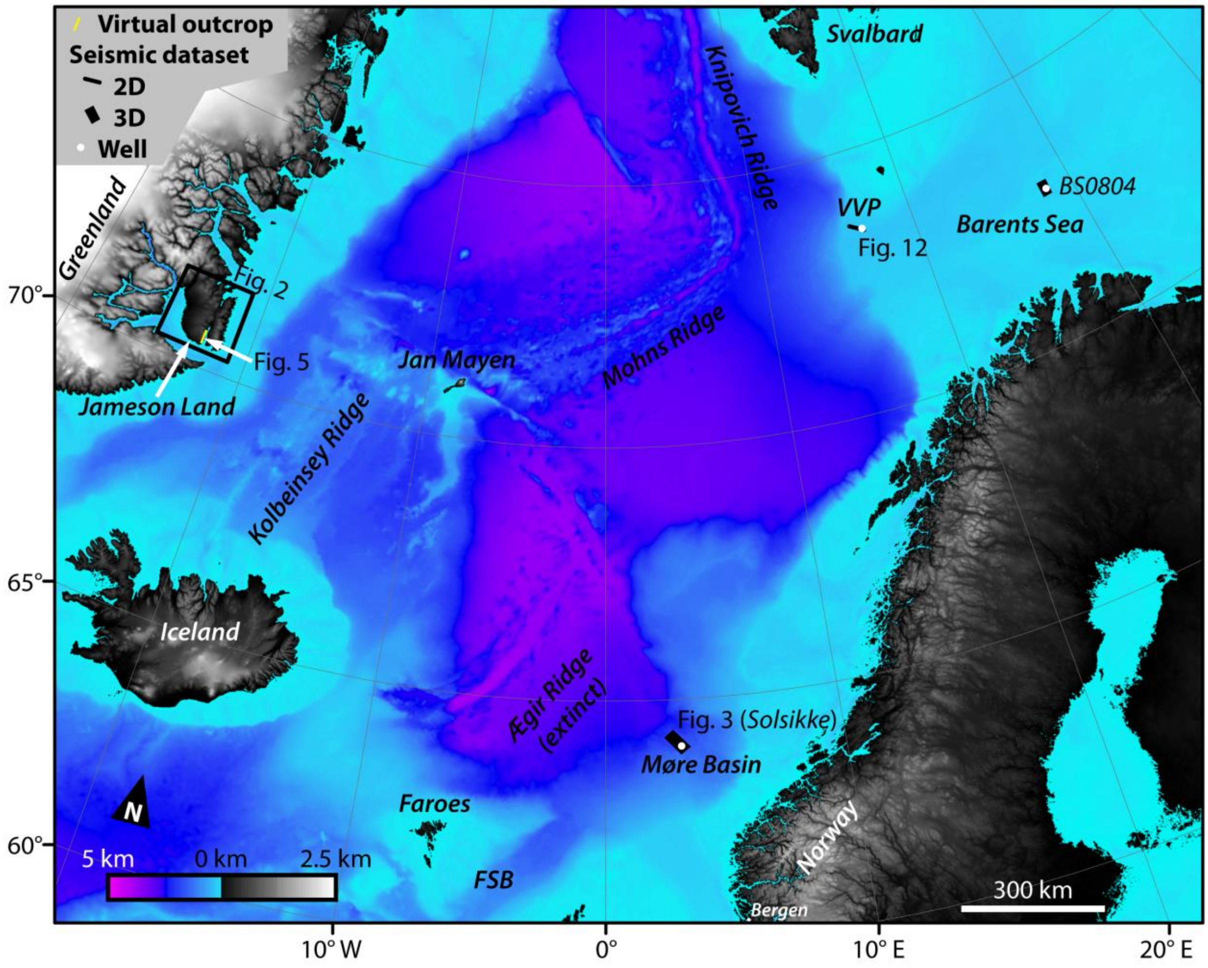


***Fig. 1**: Map of the conjugate margins around the Norwegian-Greenland Sea, showing locations of datasets used for this study. FSB, Faroe-Shetland Basin, VVP, Vestbakken Volcanic Province.*

However, studies of large-scale architectures of igneous intrusions are commonly based on seismic data. Studies of architecture of igneous intrusions in outcrops that have a scale comparable to seismic data (100s of m high, tens of km long) are rare, apart from one study by Eide et al (2017, Figs. 1, 2). This is a problem, because it is difficult to know exactly how properties of sill complexes imaged in seismic reflection data relate to actual geometries of sill complexes within the subsurface (Magee et al., 2015). This is an issue due to three inherent limitations in the 3D seismic method (Fig. 3): (1) decrease of seismic quality and resolution with depth due to absorption of high frequencies, seismic energy and downward increase in seismic velocity (e.g. Brown, 2011; Planke et al., 2005); (2)

overburden effects, where the seismic signal is affected by complex overburden which can be a considerable problem in basins with igneous rocks (e.g. Ziolkowski et al., 2003; Gallagher and Dromgoole, 2007; Flecha et al., 2011; Holford et al., 2013; Schofield et al., 2016); and (3) the inability of the reflection seismic method to image steeply dipping and vertical interfaces (e.g. Lecomte et al., 2016). Furthermore, reflection seismic data from areas with sill complexes are commonly not processed in ways that are optimal to preserve important details in such systems.

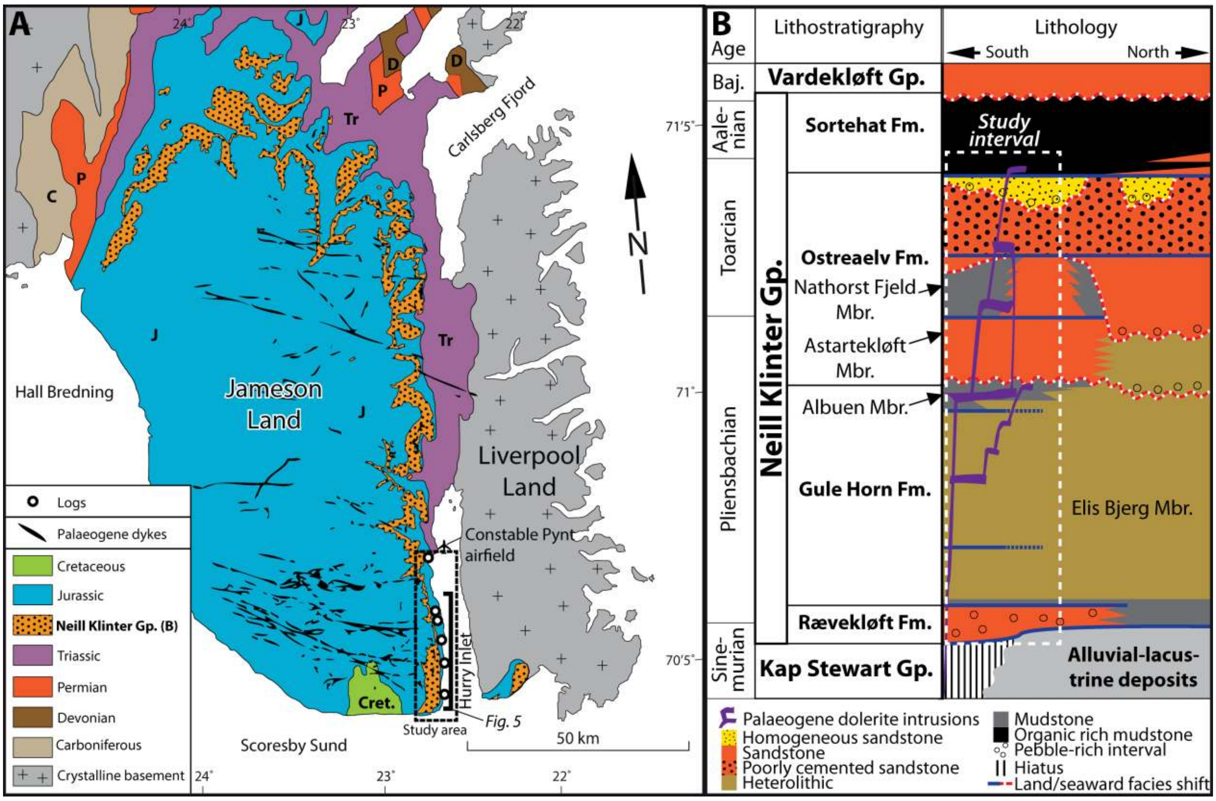


***Fig. 2****: Location and setting of the main outcrop dataset. See Fig. 1 for location. A) Geological map of the study area, modified from Ahokas et al., 2014. Note occurrence and orientation of dykes. B) Lithostratigraphy of the studied Jurassic Neill Klinter Gp. modified from Ahokas et al. (2014) and Eide et al. (2016).*

In this paper, we investigate how an outcropping example of a Paleogene sill complex intruded at c. 3 km depth in Jurassic sedimentary rock (Fig. 2; Eide et al., 2016; Eide et al.,

2017) would be imaged in seismic data at various simulated scenarios. Synthetic seismic data was generated using actual sill geometries, using methods that explicitly model both limited vertical and lateral resolution and take the inability to image steep beds into account (Lecomte et al., 2016). We also compare our results to actual seismic from the conjugate Norwegian Margin (Fig. 1), which has experienced significant volcanism both related to initial opening and during later readjustment (Talwani & Eldholm, 1977; Saunders et al 1997; Faleide et al 2008).

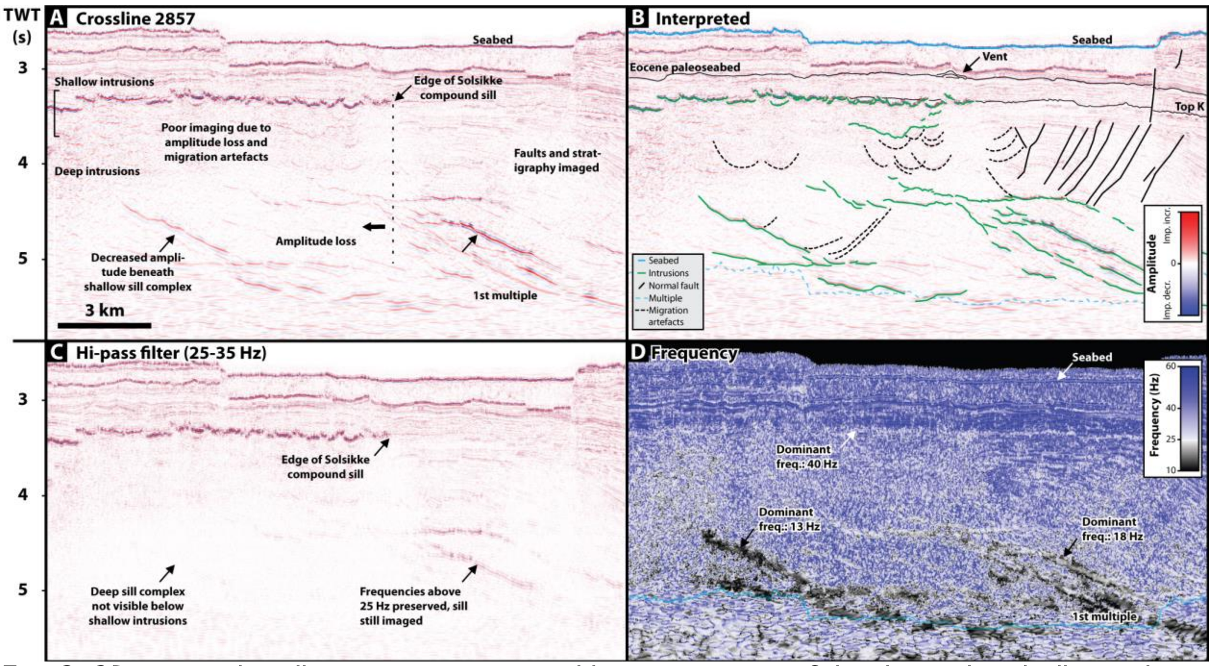


***Fig. 3**: 3D seismic line illustrating common problems in imaging of deeply emplaced sill complexes. For location, see Fig. 1. A) Uninterpreted seismic line. Note the shallow intrusions, which mainly appear as compound saucer-shapes, and the deeper intrusions, which show more layer-parallel and branching geometries. Also note the poor imaging below the shallow intrusions, and the general decrease of resolution and definition with depth. B) Interpreted version of A. K, Cretaceous. C) High-pass filtered version of (A) with a filter sloping from 35 to 25 Hz. Note loss of high frequencies below the shallow intrusions. D) Frequency plot of (A). Note the high dominant frequencies in the*

*upper part of the survey, lower frequencies in the lower part, and lowest frequencies below the shallow sill.*

The goals of this study are threefold: (1) To investigate how deeply emplaced sill complexes are imaged in seismic data; (2) to investigate how overburden complexity affect imaging of sill complexes; and (3) to discuss what these results imply for the interpretation of sill complexes in seismic data

## 1.1 Synthetic seismograms of sill complexes - issues and previous work

In order to investigate how actual sill geometries would be imaged in seismic data, synthetic seismograms can be generated from actual or conceptual models. Synthetic seismic data are routinely generated by constructing a reflectivity model, extracting one-dimensional (1D) vertical traces along this model, convolving these traces with a wavelet to obtain seismic traces (referred to as 1D convolution), and visualizing these traces as synthetic seismograms (Fig. 4c; e.g. Magee et al., 2015). However, although such methods give a good, first-order impression of seismic imaging in areas of little lateral variation, such methods are unsuited to investigate seismic imaging of complex geology because they vastly overestimate lateral seismic resolution, (e.g. Lecomte et al., 2016), such as in the case of salt diapirs, fold-and-thrust belts, or in this case: igneous sill complexes. The 1D convolution method leads to three particularly noteworthy artefacts that would not be present in actual 'real-world' reflection seismic data (Fig. 4). Firstly, reflector terminations of geological units (e.g. sills) will be much more pronounced than they will be in actual seismic data. Secondly, steeply dipping interfaces will be imaged perfectly (i.e. dykes), in contrast to actual seismic which is not adept at imaging steeply dipping interfaces. Thirdly, tuned reflector packages (e.g. thin sills) will exhibit pronounced amplitude variations which relate to thickness-variations in 1D convolution methods (e.g. Fig 4; Magee et al 2015), but these effects will be much more diffuse in real seismic data, because the resolution is limited and the response from a single sampled point will be influenced by adjacent points. This could lead to significant misunderstandings if lessons learned from analysis of such synthetic seismic data are used to guide interpretation of real seismic

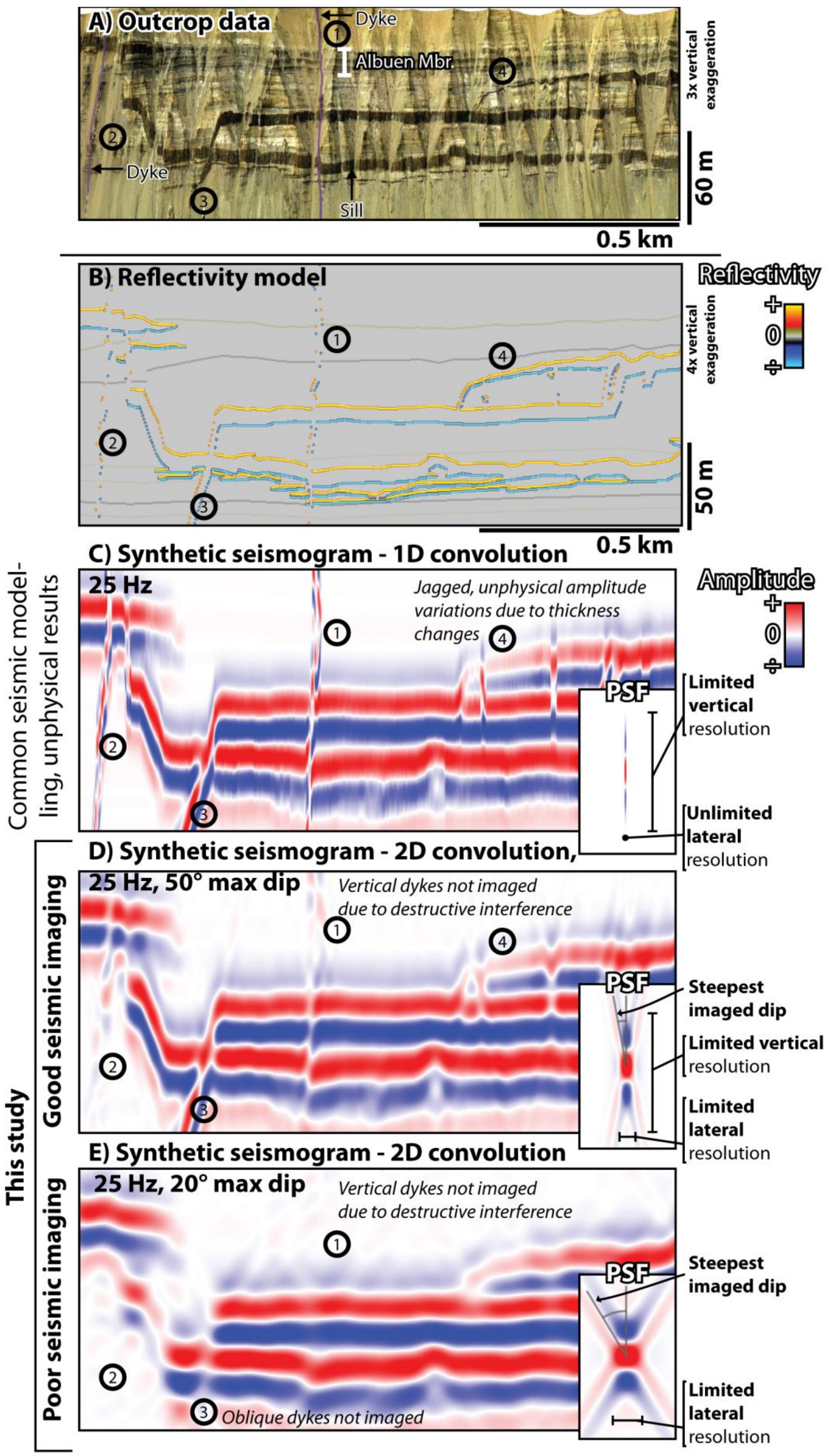
A) Outcrop data
Dyke
Albuen Mbr.
Dyke
Sill
3x vertical exaggeration
60 m
0.5 km
B) Reflectivity model
Reflectivity
4x vertical exaggeration
50 m
0.5 km
C) Synthetic seismogram - 1D convolution
25 Hz
Jagged, unphysical amplitude variations due to thickness changes
Amplitude
PSF
Limited vertical resolution
Unlimited lateral resolution
Common seismic modelling, unphysical results
D) Synthetic seismogram - 2D convolution,
25 Hz, 50° max dip
Vertical dykes not imaged due to destructive interference
PSF
Steepest imaged dip
Limited vertical resolution
Limited lateral resolution
Good seismic imaging
This study
E) Synthetic seismogram - 2D convolution
25 Hz, 20° max dip
Vertical dykes not imaged due to destructive interference
Oblique dykes not imaged
PSF
Steepest imaged dip
Limited lateral resolution
Poor seismic imaging

***Fig. 4*** *(previous page): Figure illustrating the seismic convolution method, the difference between 1D and 2D convolution, and the concept of point-spread functions (PSF). A) Outcrop data from Jameson land. Intrusive rocks appear as dark bands in the cliffs. B) Reflectivity model, generated from outcrop architectures and acoustic properties from relevant wells (Table 1). C) Synthetic seismogram generated using the reflectivity model in (A) convolved with the 1D wavelet shown in the inset. Using this method, the lateral resolution is vastly overestimated and equal to the trace spacing. Dipping interfaces are shown to be imaged even though they would not be in actual seismic. D) Synthetic seismogram generated using 2D convolution and high (50°) maximum imaged dip-values. This simulates good-quality seismic with the ability to image relatively steeply dipping interfaces. E) Synthetic seismogram generated using 2D convolution and low (20°) maximum imaged dip-values. This simulates poor-quality seismic which does not have the ability to image steeply dipping interfaces. In particular, note these four main issues: (1) Near-vertical dykes are well-imaged in synthetic seismic based on 1D convolution, but are not imaged well due to destructive interference in the 2D-convolution cases. (2, 3) Steeply dipping (c. 45°, 30°) dykes are not imaged in poor-quality seismic. (4) Because adjacent traces do not influence each other in 1D convolution, the reflectors show a jagged and unphysical amplitude variation, which are not be present in the 2D convolution models.*

Shallowly emplaced (<1.5 km) intrusions, which often show saucer-shaped morphologies (Fig. 3; Hansen and Cartwright, 2006a; Galland et al., 2009), are generally well-imaged and well understood, as these often occur shallowly without overlying high-velocity layers. However, architectures of deeply emplaced (> 1.5 km) sill complexes have been subject of much fewer studies than their shallow counterparts, and their architectures thus are less understood (Schofield et al., 2012a; Eide et al., 2017). Furthermore, these are often more poorly imaged in seismic data due to a thicker overburden and overlying intrusions (Fig. 3). This leads to reflection and absorption of seismic energy and attenuation of the seismic signal particularly at high frequency ranges (Fig. 3c, d; Gallagher and Dromgoole, 2007). Additionally, diversion and spreading of large-incidence ray paths occur, leading to both lower resolution and inability to image steeply dipping features at depth.

These imaging issues commonly lead to intrusions which are not imaged at all, intrusions that are imaged as tuned packages making thickness estimates difficult, multiple intrusions imaged as a single one due to interference effects, and unimaged steeply dipping features. Such effects become more pronounced with increasing depth.

## 2. Geological setting

The synthetic seismic generated in this study is based on architectures of igneous intrusions and host rock from a field area near the SE margin of the Jameson Land Basin (Fig. 2) in East Greenland (Eide et al., 2017). The host-rock was deposited in the Early-Middle Jurassic during a post-rift thermal sag phase (Surlyk, 2003), and comprises a variety of shallow-marine tide-influenced deposits, deposited close to the basin margin (Dam and Surlyk, 1998; Ahokas et al., 2014; Eide et al., 2016). The Jameson Land Basin was a minor sub-basin in the Mesozoic seaway between Norway and Greenland, and the sedimentary host rocks are time- and facies equivalent to the Båt Group, which is a prolific reservoir interval on the conjugate Haltenbanken area on the Norwegian Continental Shelf (Martinius et al., 2001; Ichaso and Dalrymple, 2014; van Capelle et al., 2017). Late Jurassic-Cretaceous rifting and Paleogene volcanism led to deposition of c. 3 km of sediment and lava flows above the study interval (Brooks, 1973, 2011; Larsen and Marcussen, 1992; Mathiesen et al., 2000; Surlyk, 2003). Deposition of a significant overburden led to extensive quartz-cementation prior to emplacement of igneous intrusions (Hald and Tegner, 2000). The detailed sill architecture and emplacement of these intrusives are covered by Eide et al (2017).

The studied sills are part of the mafic Jameson Land Suite, and consist of nonvesicular, aphyric tholeiitic dolerites with ophitic texture (Hald and Tegner, 2000), and vary in thickness from c.1 – 18 m, with an average thickness of 9 m (Eide et al., 2017). The architecture and morphology of the sills is clearly influenced by the host-rock lithology, as they favour mudstone units, avoid well-cemented homogeneous sandstone units, and show clear ductile/non-brittle emplacement features in poorly-cemented sandstone intervals (Eide et al., 2017). For the purpose of this study, the intrusives and Jurassic host-rock can be divided into seven facies associations (Figs. 2B, 5; Table 1): 1: Paleogene *dolerite intrusives*, 2: *Homogeneous sandstone*, 3: Bedded *sandstone* with thin mudstone intervals, 4: *Poorly cemented sandstone*, due to the presence of chlorite overgrowths (c.f. Ahokas et al., 2014), 5: *Heterolithic* beds, comprising interbedded 1-50 cm thick sand- and mudstone beds with a variable but overall approximately equal proportion of sandstone and mudstone, 6: Regionally extensive *mudstone* intervals, and 7: *organic rich mudstone* intervals (c.f. Krabbe et al.,1994).

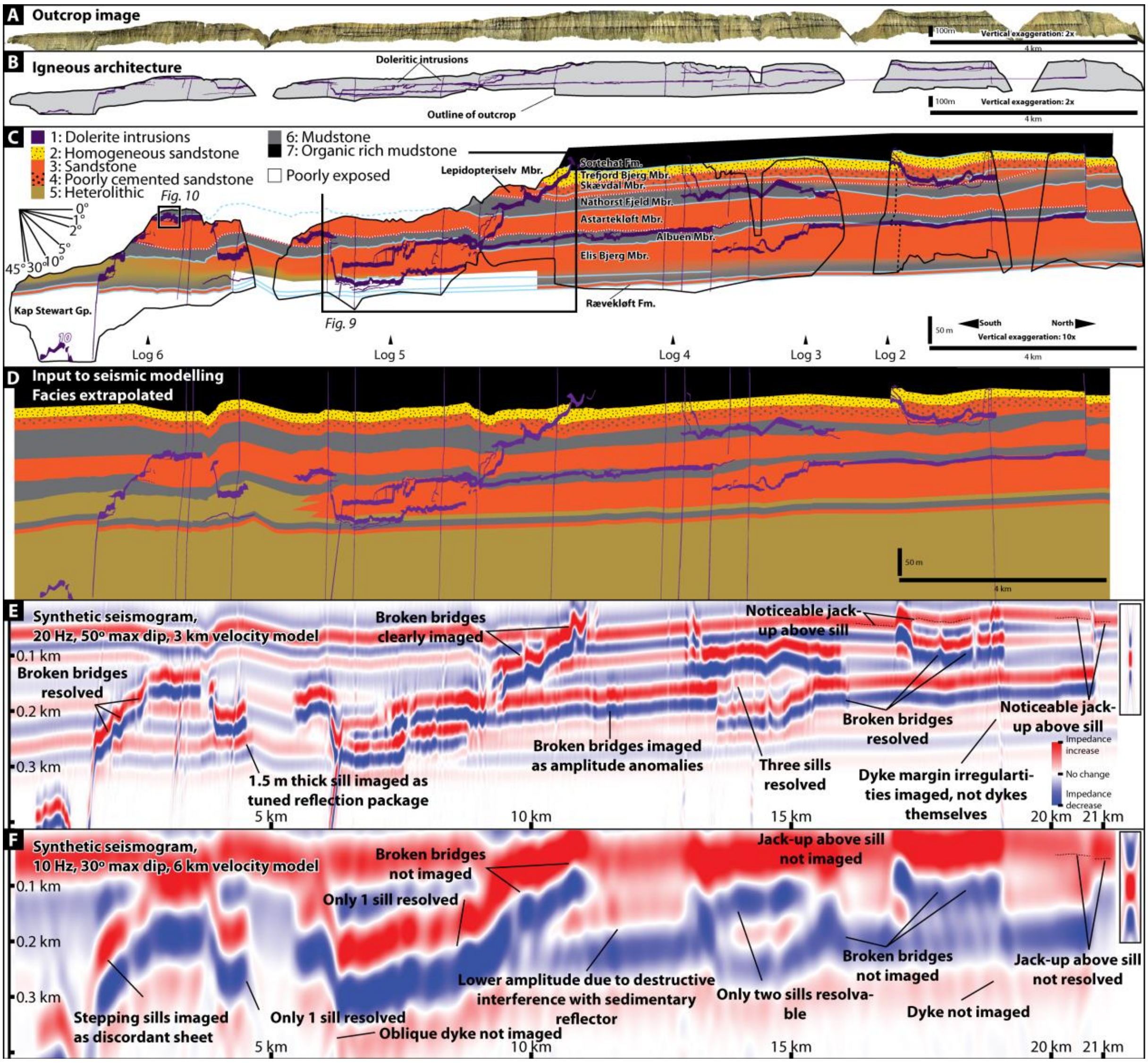


***Fig. 5:*** *Outcrop data and synthetic seismograms. For location, see Figs 1, 2. A) Studied outcrop. B) Igneous architecture from outcrop. C) Stratigraphic and igneous architecture. Note the tendency for intrusions to follow mudstone units, the consistent upwards-stepping of intrusions in other lithologies than mudstone, and the jack-up of host-rock above sills. D) Input model to seismic modelling, populated with acoustic properties from Table 1. E) Synthetic seismogram corresponding to model A in Fig. 7. F) Synthetic seismogram corresponding to model C in Fig. 7.*

The sills in Jameson Land are examples of a small-volume (c. 10% of studied outcrop sections) sill complex intruded deeply (3 km) into what was mainly brittle, layered host rock at the time of emplacement, which has subsequently been uplifted (Mathiesen et al., 2000;

Eide et al., 2017). This is in contrast to many exposed examples of sill complexes around the world. For example, the sills in the Karoo Basin of South Africa are associated with somewhat larger intrusion volumes, and were emplaced at shallower levels, where the host rock was likely not well consolidated (Schofield et al., 2010; Svensen et al., 2015). Other relatively well-studied examples, such as the sills emplaced in the Theron and Transantarctic Mountains of Antarctica were emplaced at a similar depth  and in similar lithologies, but are the results of much higher intrusion volumes as these attain thicknesses of up to 330 m (Hersum et al., 2007; Jerram et al., 2010). The architecture and depth of intrusion the Jameson Lands Suite makes it broadly comparable to intrusions seen at deep levels (~ 3-4 km), around the Base Cretaceous/Top Jurassic, in the present day contemporaneous basin fill along the NE Atlantic Margin (c.f. Schofield et al., 2015).

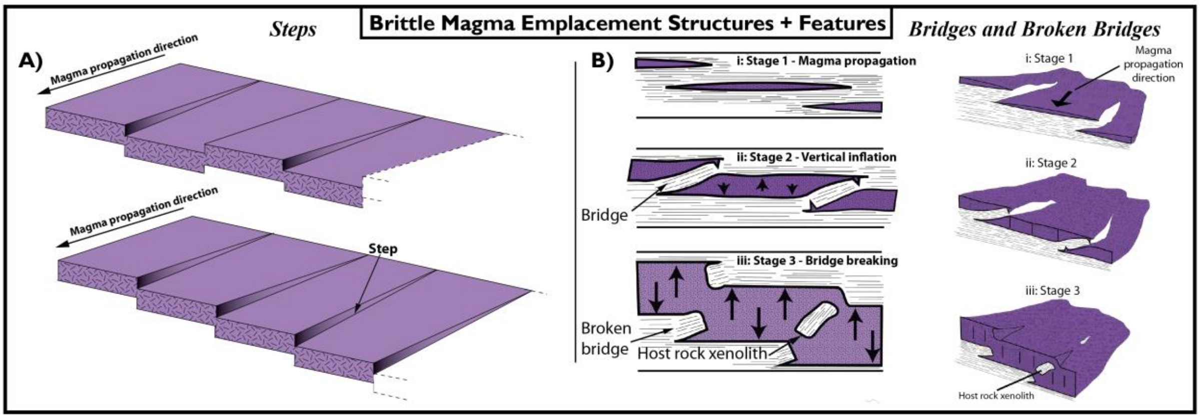


***Fig. 6:*** *Overview of sill structures and features formed during emplacement and progressive inflation of sill intrusions in brittle host rock. A) Expression of steps on sill margins and how they relate to the magma propagation direction. B) Development of broken bridges through vertical inflation of sills and breaking of host-rock bridges between sill segments. Modified from Schofield et al. (2012a).*

## 3. Sill architecture in studied outcrop section

The sedimentary and igneous architecture of the studied outcrop in Jameson Land is shown in Fig. 5. A variety of features noteworthy for the present study occur in this section, and these are summarized from Eide et al. (2017): Individual sills are broadly layer-parallel and constitute c. 10% of the material in the outcrop section. *Main sills* make up 70% of the intruded volume, and these are 7-12 m thick, with average thickness of 9 m. These are associated with *sill splays* which are less than 2 m thick minor sills which occur close to the main sills where the sills are propagating in other lithologies than thicker mudstone interval.

The type of host rock (*mudstone*, *brittle interbedded sandstone and mudstone*, *brittle homogeneous sandstone*, and poorly cemented sandstone) is a major controlling factor on the architecture of sills. The majority of sills (c. 60%) are emplaced within mudstones (FA6), and sills show great vertical stability and attain lengths of up to 6 km when emplaced in regional mudstones. Sills emplaced in well-cemented interbedded sandstone (FA 2) and mudstone (FA 5) make up c. 25% of sill volume, and show abundant *broken bridge*-structures (Fig. 6) and cross-cut stratigraphy and propagate towards regional mudstones (Fig. 5C). Broken bridge structures form as two vertically offset sills inflate, bend the host rock “bridge” formed between them, which and finally break and are filled with magma (Hutton, 2009; Fig. 6). These are important indicators for magma propagation directions (Schofield et al., 2012a), and unbroken bridges may form pathways for hydrocarbon migration. Sills in other lithologies constitute a minority (15%) of the intrusive material in the outcrop, but show complex splaying geometries in homogeneous sandstones (FA2), and both ductile and brittle emplacement features in poorly cemented sandstones (FA4).

E-W-trending, near-vertical (dip > 80°) dykes occur in the study area (Figs. 2A; 5C), and these vary from 1-10 m wide, and only a small number of these are interpreted to postdate the main intrusive phase based on cross cutting relationships, the rest are believed to be coeval with the dykes as radiometric ages are coherent and cross-cutting relationships are not observed in these (Hald and Tegner, 2000). Together, sills and dykes compartmentalize the host rock into intrusive-bounded blocks c. 0.2-4 km wide and 20-120 m thick.

Intrusion of c. 10% of igneous material into the outcrop volume appears to be solely accommodated by 1:1 uplift of host-rock above sills, as no deformation of the rock apart from sill intrusion is observed. This is proven as stratigraphy restores without error by removing the sills from images taken orthogonal to the plunge of the beds (Eide et al., 2017, their fig. 10). This indicates that deeply emplaced sills (> c. 1.5. km) formed as fractures which inflated vertically in a piston-like fashion.

The sum of these observations show that sill complexes have complicated architectures and that details of the architecture might have significant implications for sill emplacement, reservoirs and hydrocarbon systems (c.f. Rateau et al. 2013; Senger et al., 2017). How these details are imaged in seismic data will be studied in the remainder of this contribution.

## 4. Methods and dataset

### 4.1 Outcrop data

The synthetic seismic models are based on analysis of the Jameson Land Suite mafic sill complex and surrounding host-rock of the Neill Klinter Group, imaged in a 22 km long and 250 m high virtual outcrop model (Figs. 2A, 5A) acquired using oblique helicopter-mounted LIDAR-scanning. The outcrop model was acquired using the Helimap System (Vallet and Skaloud, 2004), using a laser scanner, a digital medium format camera with a 35 mm lens, and an inertial navigation system. Processing of these data (see Buckley et al., 2008; Rittersbacher et al., 2014; Eide et al., 2016) yielded a high-quality virtual outcrop model devoid of errors with 0.3 m point spacing and a pixel resolution of c. 7 cm. The outcrop model is constrained by a set of six sedimentary logs with a total thickness of 1040 m, acquired along the outcrop with the objective to sample the entire section. The outcrop is oriented north-south, and yields a strike-section through the sedimentary systems (Dam and Surlyk, 1998). Apart from a general proximal-to-distal trend leading to more fine grained deposits southwards in the outcrop throughout all stratigraphic units, the lateral lithological variability is low (Dam and Surlyk, 1998; Eide et al., 2016), leading to a good constraint on lithology, and ultimately petrophysical properties used for modelling, in the entire dataset. The outcrops are generally unfaulted, tectonic dip is on average 3° towards the west, and the sills are generally parallel to sedimentary layering.

### 4.2 Synthetic seismic modelling

The synthetic seismic models of the studied section yields data that simulate zero-phase, pre-stack depth-migrated (PSDM) reflection seismic data (Lecomte et al., 2015). The methodology applied for modelling is based on ray-tracing modelling for study of the overburden effects (Gjøystdal et al., 2007; Fig. 7), and on 2D convolution (Fig. 4) for the generation of synthetic seismic models of the target interval (c.f. Lecomte et al., 2016). Detailed reviews of these methods are out of scope of this paper and therefore only briefly summarised here.

Four main parameters are critical for the seismic modelling presented in this study: rock P-wave velocity, rock density, seismic frequency at the target depth, and maximum imageable dip (*max dip*) at the target depth. In reality, these parameters vary with the depth of the modelling target, the geological history of the target and overburden, and the overburden architecture. P-wave-velocity and density of host-rock and overburden are taken from relevant depths in time-equivalent formations on the Norwegian Continental Shelf (Table 1), and P-wave-velocity and density for igneous intrusions is from Smallwood and Maresh (2002). Seismic frequency at the target depth is measured from a publically available 3D seismic dataset from the Møre Basin (NH003, “*Solsikke”*) in the base case, and from the BG0804-survey in the Barents Sea in the case of an eroded basin (Fig. 8). It is worth pointing out that vertical resolution in seismic data is a function of the signal frequency and host rock velocity, while the lateral resolution is a function of signal frequency, host-rock velocity and max dip (c.f. Figs 4d, e). In each modelled scenario, the lateral resolution has been ascertained using ray-tracing methods (Fig. 6) to investigate the max dip and investigation of seismic frequency in comparable datasets to investigate the seismic frequency (Fig. 7).

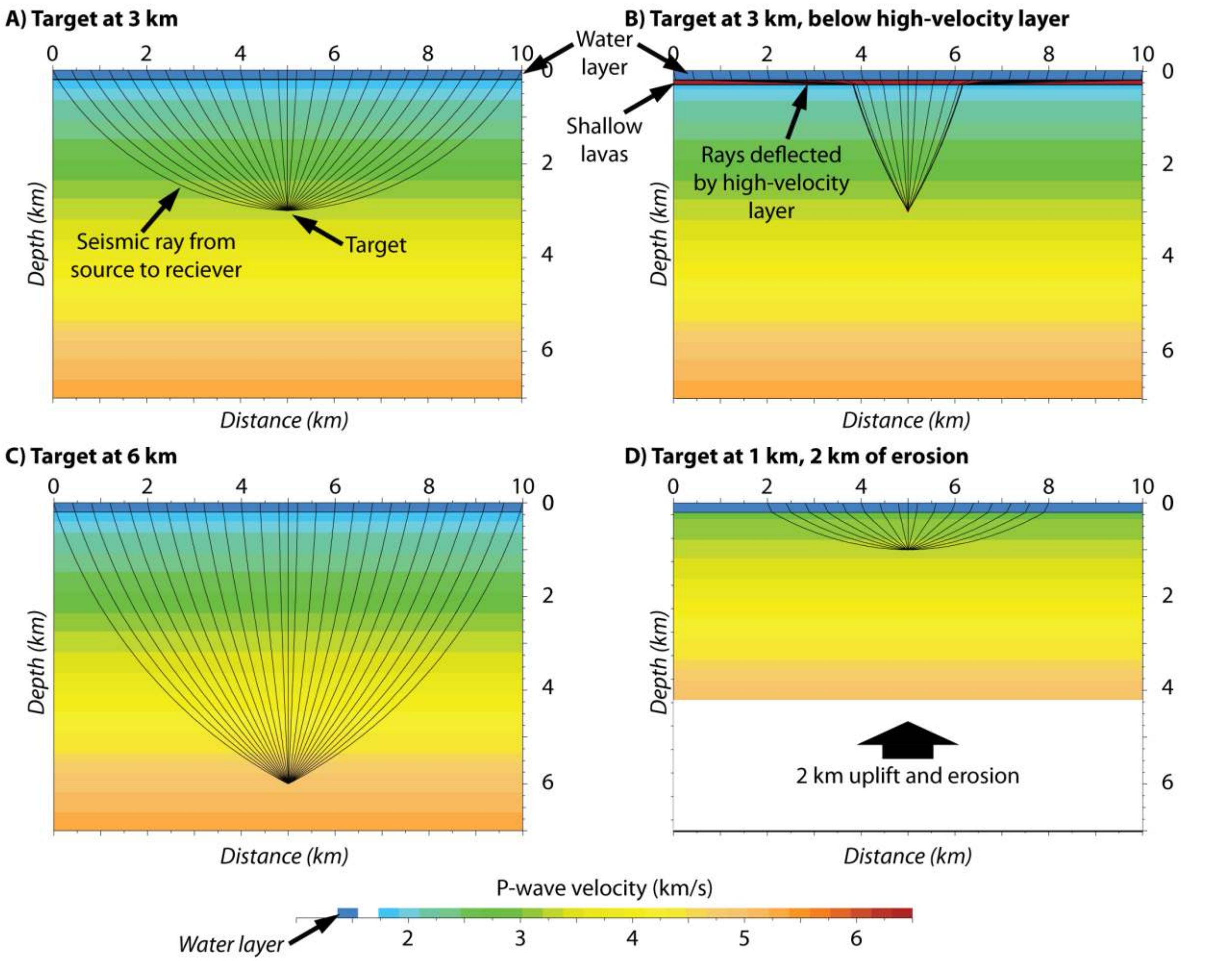


***Fig. 7:*** *Overburden models, generated to investigate ray propagation and resulting lateral resolution for each of the modelling cases. Interfaces hit by rays with a large variety of incidence angles will*

*have high lateral resolution, and more steeply dipping interfaces are imaged. A) Model with linear P-wave velocity gradient based on well data from the Norwegian Continental Shelf. B) Model with a linear P-wave velocity gradient and an overlying high-velocity layer (i.e. shallow intrusions, lavas). Note the steep incidence of rays reaching the target, resulting in low lateral resolution. C) Model with linear P-wave velocity gradient and target at 6 km. Note the steeper incidence of rays compared with (A). D) Model with the upper 2 km of stratigraphy eroded, resulting in a hard sea-floor. Note the large variety of incoming angles.*

Interpretations of sedimentary units and intrusions on the virtual outcrops from Jameson Land yielded scaled, georeferenced lines which have been projected onto a plane parallel to the outcrop and orthogonal to the plunge of the tectonic dip of layers. As the lateral lithological variability is quite low (Eide et al., 2016), extrapolation of stratigraphy and sill architecture into unexposed areas was straightforward (Fig. 5D).

The seismic modelling procedure utilized in this paper consists of the following steps: (1) Four different conceptual overburden models are created using specified layer interfaces, velocity gradients and petrophysical properties derived from relevant wells from the Norwegian Continental Shelf (Fig. 7, Table 1). (2) Wave-propagation from seismic source, through the overburden model, to the target depth, and then back to receivers is simulated based on ray-based algorithms in order to investigate max dip, which is used to define a two-dimensional spatial "wavelet" (termed Point-Spread Function (PSF), c.f. Fig. 4) based on dominant frequency of the signal at depth of interest. (3) Creation of a target reflectivity model using layer architectures obtained from outcrop analysis (Fig. 5D) and petrophysical properties from appropriate subsurface analogues in wells (Table 1). (4) 2D convolution of the target reflectivity model with the PSF obtained in Step 2.

## 4.3 Overburden models

This section contains an overview of imaging parameters (max dip and dominant frequency) based on the four overburden scenarios presented in Fig. 7). All models are constructed with a 0.2 km thick water layer, and a linear P-wave-velocity gradient from 1.8 km/s at the seabed, 3.3 km/s at 3 km, and 4.8 km/s at 6 km.

**Model A (simple overburden)** investigates a base-case, where the target intrusions are simply overlain by 3 km of sedimentary host rock (Fig. 7A).

**Model B (overlying igneous rocks)** investigates the influence of an overlying high-velocity layer, here modelled as a 100 m layer of basalt at the seabed (Fig. 7B). Several other geometries of overlying high-velocity layers have also been modelled, and gave similar results.

**Model C (deep burial)** tests the effects of further burial of the target interval to 6 km (Fig. 7C), resulting in increased host-rock velocity at the target interval (Table 1b).

**Model D (uplifted and eroded basin)** tests the effect of 2 km of erosion of the overburden (Fig. 7D), yielding a P-wave velocity of 2.9 km/s at the seabed and a target at 1 km depth where the average seismic velocity is 3.3 km/s.

### 4.3.1 Seismic frequency

Frequency attenuation with depth is difficult to model well, and signal frequency versus depth is therefore taken from the *Solsikke* 3D seismic survey from the Møre Basin (Fig. 1) for models without erosion (Fig. 7A-C). Shallow sills, presently at c. 1 km below the seabed, are imaged at c. 40 Hz. Deeper intrusions at c. 3 km below the seabed, are imaged at c. 18 Hz, which is similar to what is observed in other localities on the N Atlantic Margin, such as in the Faroe-Shetland Basin (Schofield et al. 2015), where the sills occur at relatively deep levels within the contemporaneous basin fill. Significant attenuation of high frequencies (>25 Hz) occur just below shallow sills (c.f. Fig. 3). At depths of c. 3 km in areas underlying shallow sills, the dominant frequency is c. 13Hz, a significant reduction compared to 18 Hz in areas without overlying sill intrusions (Fig. 8B). For model D, where 2 km of overburden has been eroded, frequency versus depth has been taken from the BG0804-survey in the Barents Sea (Fig. 1) where the depth of erosion is similar (Henriksen et al., 2011; Baig et al., 2016), and a number of exploration wells provide excellent depth control. At c. 1 km depth, the dominant frequency is 40 Hz (Fig. 8B).

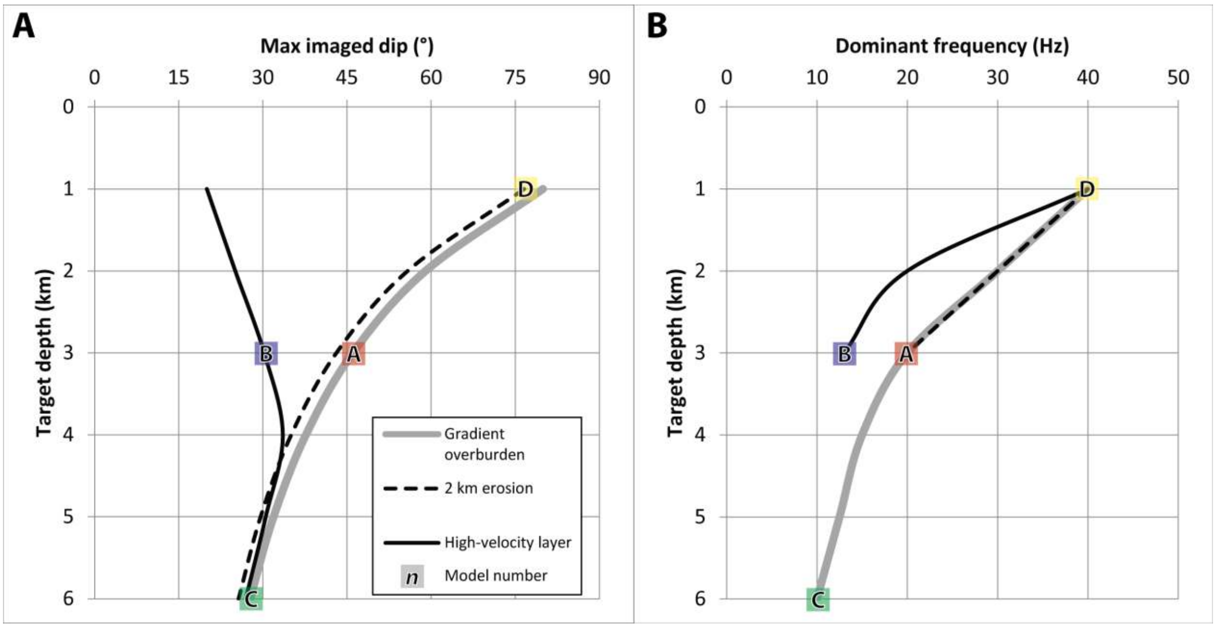


**Fig. 8**: Imaging properties used for seismic modelling. Inset letters show properties at target depth corresponding to cases in Fig 7. A) Max imaged dip calculated from models in Fig. 7. B) Dominant frequency at depth, derived from analysis of 3D seismic datasets.

### 4.3.2 Illumination and overburden effects

The main goal of the overburden modelling is to investigate the max dip and thus also the lateral resolution which can be expected to occur in different cases and at different depths. Using a ray-tracing approach, lateral resolution in seismic data is a function of the max dip, and this is illustrated in Fig. 4. The modelling shows that the models can be grouped into three groups which behave similarly (Fig. 7):

Models with a simple overburden (Fig. 7A,C) have high max dip (and hence high lateral resolution) at shallow depths (c. 80° at 1 km), and decrease rapidly to c. 45° at 3 km and 30 ° at 6 km.

In the models with high-velocitiy layers above the target, only rays with steep incidence angles may reach the target, as lower-incidence angle rays are refracted away from the target (Fig. 7b). This leads to a significant reduction in max dip, to c. 30° at 3 km depth (Fig. 8A)

In the models with 2 km erosion, the maximum imageable dip is close to, but slightly higher than models with a simple velocity-gradient in the overburden (Fig. 7A). The strong velocity contrast at the seabed caused by a hard sea-floor will increase the potential for strong multiples. This would lead to potentially worse imaging than is accounted for here, and costlier processing.

### 4.3.3 Summary of overburden effects

In summary, the modelling of overburden effects and inspection of publically available 3D-seismic data from relevant basins indicate that for models with a simple gradient overburden, seismic frequency and max dip decrease gradually with depth (Fig. 7). For models with overlying high-velocity layers such as lavas and/or shallow intrusions, a sharp decrease in max dip and high-frequency content occurs once the seismic waves hit the high-velocity layers. This leads to poorer imaging of underlying features such as sill complexes. The eroded models show a moderate worsening in lateral resolution, and a frequency-loss with depth comparable to models with simple gradient.

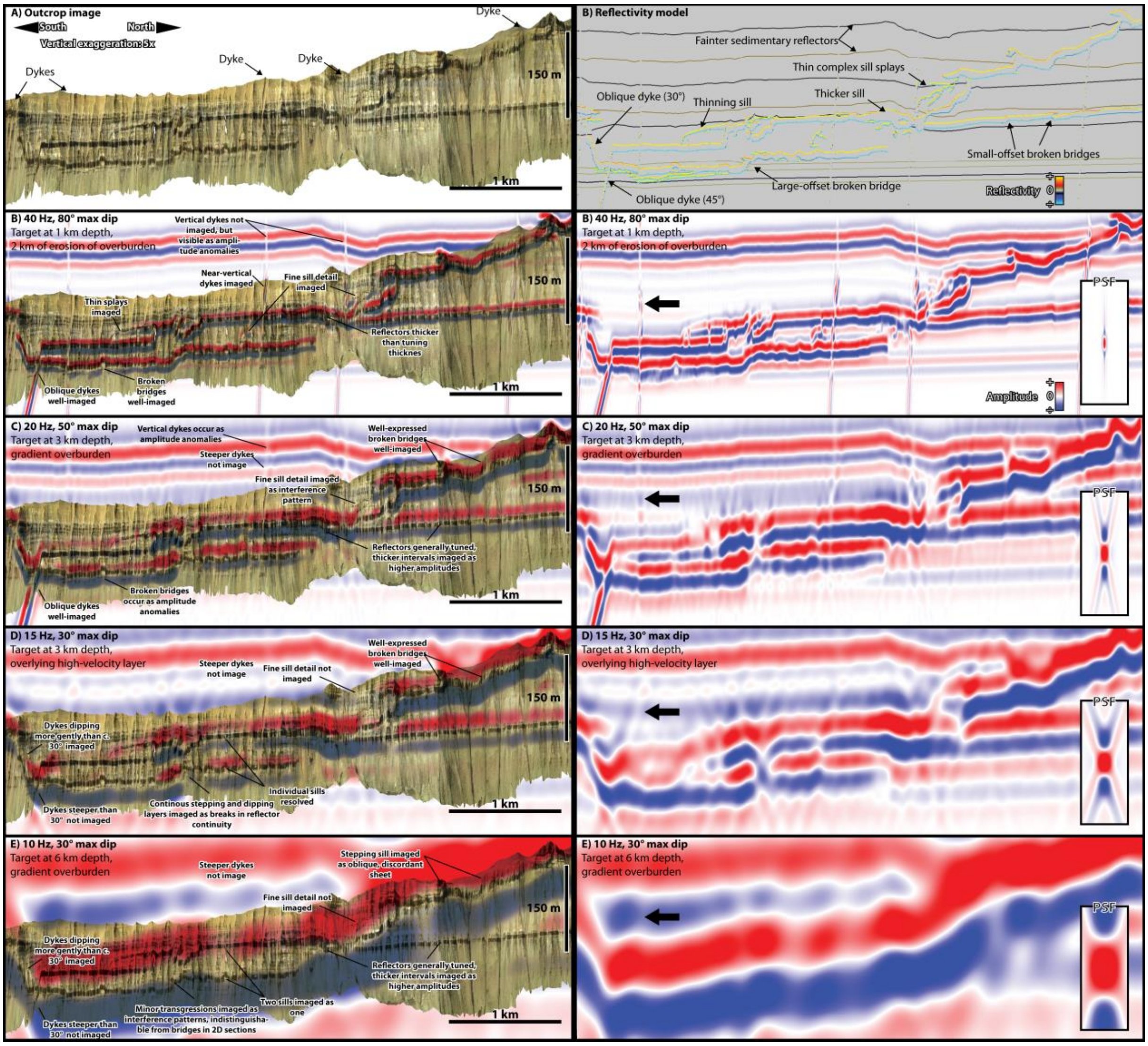


**Fig. 9:** Comparison of details of synthetic seismograms. See Fig. 5C for location. A) Outcrop image and reflectivy model. B) Synthetic seismogram corresponding to model D in Fig. 7. C) Synthetic seismogram corresponding to model A in Fig. 7. D) Synthetic seismogram corresponding to model D in Fig. 7. E) Synthetic seismogram corresponding to model C in Fig. 7. Note that this model is generated using velocity model B in Table 1.

## 5. Forward seismic modelling

### 5.1 Base case: target at 3 km depth, simple overburden

For an overburden model with a linear velocity gradient, a target at 3 km depth, and an average host-rock P-wave velocity of 3.1 km /s at the target, the modelling indicates that high-quality, depth-migrated, zero-offset seismic data will appear similar to Fig. 5E. Despite the modest thickness of main sills (7-16 m) and relatively low seismic frequency (20 Hz) the sills are still clearly imaged in the synthetic seismogram, with reflections from the intrusions showing much higher amplitudes than reflections from sedimentary interfaces. Tuning thickness (e.g. Brown, 2011) is commonly taken as a quarter wavelength, which in this case ($V_P$=3.1 km/s, $f$=20 Hz) is c. 40 m. The sills are still visible, despite being thinner than seismic resolution, due to the large impedance contrast between host-rock and intrusions (c.f. Table 1a). Even sills as thin as 1.5 m are resolvable as amplitude anomalies, but it is likely that these could not be resolved in actual seismic data due to presence of noise.

The architecture of the sill complex is generally well-constrained in the synthetic seismogram, and both sills and oblique dykes with dips shallower than 45° are imaged. Broken bridges (c.f. Fig. 6) with large offsets (> 15 m) are clearly imaged as large steps in the simulated reflections, while broken bridges with offsets smaller than 15 are imaged as local amplitude anomalies in otherwise continuous reflections (Fig. 9C, 10c). It is likely that these amplitude anomalies caused by small-offset (< 15 m) broken bridges could not be separated from seismic noise in real seismic data. The steps observed on the investigated sill margins are not detectable in the seismograms, and this is probably due to the small step heights in this dataset, which are 2 m on average. In this seismogram, the broken bridges would be indistinguishable from larger-offset steps. Each sill in this data is generally imaged separately, as they are spaced more widely than ½ seismic wavelength.

Steeply dipping to vertical igneous features (> 50°) i.e. dykes, are not imaged directly in this seismogram, but dykes are still resolvable in the seismogram (Fig. 9C). This is mainly due to two reasons: (1) dyke margin irregularities are imaged instead of the dykes themselves, and (2) dykes lead to disruptions in original layer continuity that leads to discontinuities in the seismic data that may be interpreted as dykes. In simulations with lower lateral resolution, this is not imaged (Figs 9D,E).

Pronounced jack-up of host-rock above sills is common in the dataset from Jameson Land Greenland (Eide et al., 2017). This is resolvable in the synthetic seismogram, particularly near the top at 17 and 21 km (Fig. 5E), where the reflections from the sedimentary bedding are clearly offset across dykes and transgressive sills.

In a close-up view (Fig. 9C), m-scale sill splays are imaged as complex interference patterns.

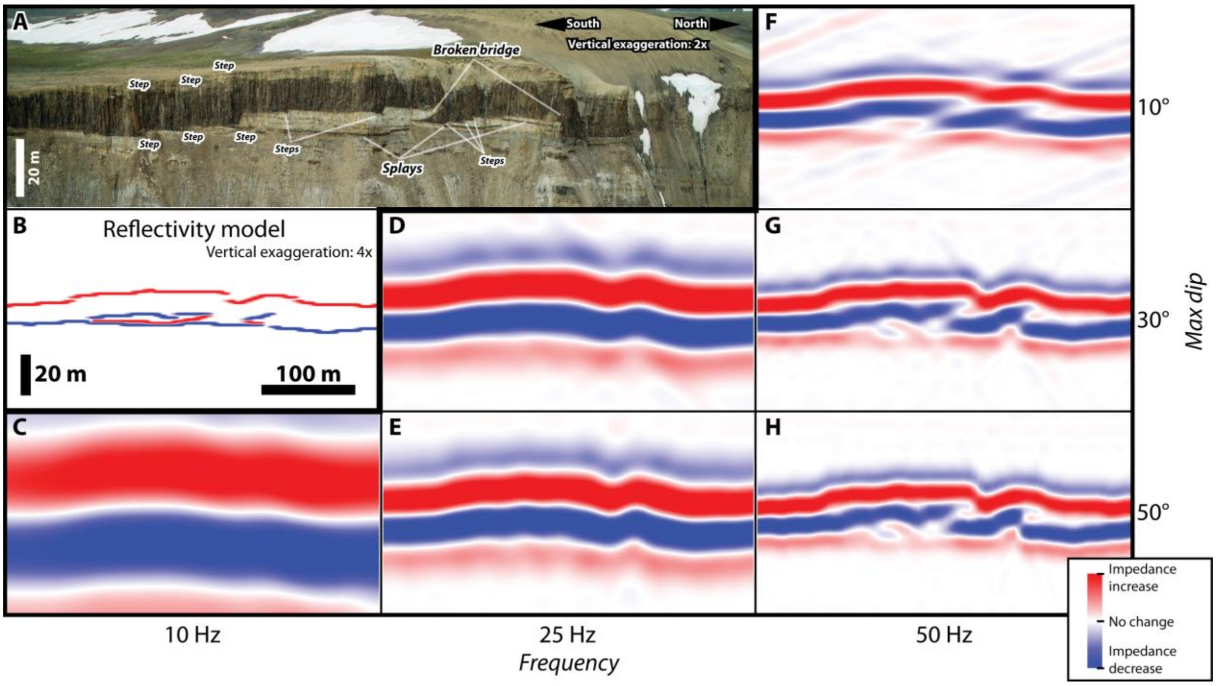


***Fig. 10:*** *Seismic imaging of broken bridges, steps and sill splays. See Fig. 5C for location. A) Outcrop image showing abundant splays, steps and broken bridges. B) reflectivity model generated from (A). C-H) Synthetic seismograms showing imaging with different lateral resolution (expressed as max imageable dip) and dominant frequency. Note especially the influence of a decrease in lateral resolution (F, G and H).*

## 5.2 Deep target: target at 6 km depth, simple overburden

For a target at 6 km depth, increased seismic velocity and density due to increasing compaction and diagenesis must be taken into account (Table 1b). For an overburden model with a linear velocity gradient, a target at 6 km depth, and an average host-rock P-wave velocity of 4.5 km /s, the modelling indicates that high-quality, depth-migrated, zero-offset seismic data will appear like in Fig. 5F. In this case, the dominant frequency would be 10Hz, yielding a tuning thickness of c. 100 m. Individual main sills (7-16 m thick) are not imaged, but instead the entire sill-complex appears as one single, complex reflector, and shows the highest amplitudes of all reflections in this seismogram due to the large amplitude contrast between intrusives and host-rock. However, at such depths, the amplitude contrast between host-rock and intrusives are smaller, and significant local amplitude decreases due to destructive interference occur. This happens particularly between dolerite intrusions and the overlying reflection generated at the interface between the Astartekløft Member and Nathorst Fjeld Member, for example at 12 km (Fig. 5F).

Prominent stepping sills, such as at 2 and 9 km along the profile (Fig. 5F), are imaged as oblique sheets which are discordant to bedding. Broken bridges are in some instances imaged as amplitude anomalies, but these are generally indistinguishable from amplitude anomalies generated by interference with reflections from host-rock stratigraphy. Due to loss of lateral resolution at depth, oblique dykes dipping more than c. 30° are not imaged due to destructive interference (Figs. 5F at 7 km; 9E).

In several cases, such as at 7 km along Fig. 5F, closely spaced sills only generate one compound reflection, and individual sills are therefore not resolvable. Jack-up of host-rock above sills at the scale observed in this outcrop (c. 8 m) is not resolved (Fig. 5F).

## 5.3 Imaging below high-velocity layers

Reduction in both the seismic frequency and lateral resolution below shallow sills is evident from analysis of the studied seismic data and overburden modelling (Figs 3, 7). This leads to a prominent decrease in seismic quality (c.f. Figs 9C,D), which is particularly evident in the broadening and thickening of the point-spread-functions (c.f. insets in Figs 9C,D). In the generated seismograms, the effect of decreasing the lateral resolution, and thereby the ability to image more steeply dipping interfaces, is striking. Broken bridges are not well imaged anymore, and stepping sills rather occur as oblique sheets. Steeply dipping oblique dykes (>35°) are not imaged, and amplitude anomalies related to presence of vertical dykes are no longer modelled to occur because of the low lateral resolution.  Furthermore, decreased frequency leads to less well-defined reflectors. Imaging of fine-scale features with varying frequency and lateral resolution is covered comprehensively in Fig. 10.

## 5.4 Shallow imaging

Provided that seismic acquisition-problems related to hard seafloor can be overcome, the model with 2 km of erosion (Fig. 7D) resulted in the best quality seismic (Figs. 9B,). In particular, compare the point-spread-function of this seismogram to the others (inset in Fig. 9B). In this case, the synthetic seismogram shows a remarkable similarity to the input model, due to the high frequency and the high max dip, yielding high lateral and vertical resolution. Tuning thickness (19 m) is close to the sill thickness, and a clear relationship between thinning of sills and decrease in amplitude occurs. Steeply dipping interfaces, broken bridges and steps are clearly imaged, and even some meter-scale sill-splays are resolved.

# 6. Discussion

## 6.1 Seismic resolution and seismic detectability of igneous sills

### 6.1.1. Seismic Resolution and Amplitude of Igneous Intrusions

In order to understand issues with imaging igneous sill complexes in reflection seismic data, seismic resolution and seismic detectability must be discussed. The limited range of frequencies available in seismic surveys leads to limitations in the lower limit of bed thicknesses that may be uniquely resolved (e.g. Simm and Bacon, 2014). The limit of thickness of beds that may be resolved is termed vertical resolution, and the seismic resolution is commonly approximated to be ¼ of the dominant wavelength ($\lambda = v_P / f$) of the signal, however, the exact value is dependent on the wavelet shape (e.g. Simm and Bacon, 2014). This is illustrated in Fig. 11, where imaging of a 100 m thinning wedge is simulated using a 2D point-spread function (PSF). Using a PSF based on a Ricker wavelet, onset of tuning occurs at λ/3 (c. 45 m). At wedge thicknesses greater than 45 m both the top and the bottom of the wedge is imaged as separate reflectors. The bed is therefore not resolved at thicknesses smaller than c. λ/3.

As the bed thickness approaches λ/4 (35 m), the side-lobes of the lower reflection start to interfere constructively with the main lobe of the upper reflection, leading to increased amplitude in this thickness range. At wedge thicknesses lower than c. λ/3, the thickness of the wedge may not be determined from the reflected signal, and his is known as the *tuning effect, tuning thickness* or the *Rayleigh Criterion* (e.g. Kallweit and Wood, 1982). The distance between the maximum amplitudes of the upper and lower reflector does thus not change as the wedge thins past the tuning thickness in Fig. 11D. As the main lobe of the upper reflection interferes constructively with the side-lobes of the underlying reflection, the reflection amplitude increases to a peak of c. 1.4x the normal amplitude. Peak tuning occurs at c. λ/5 (28 m).

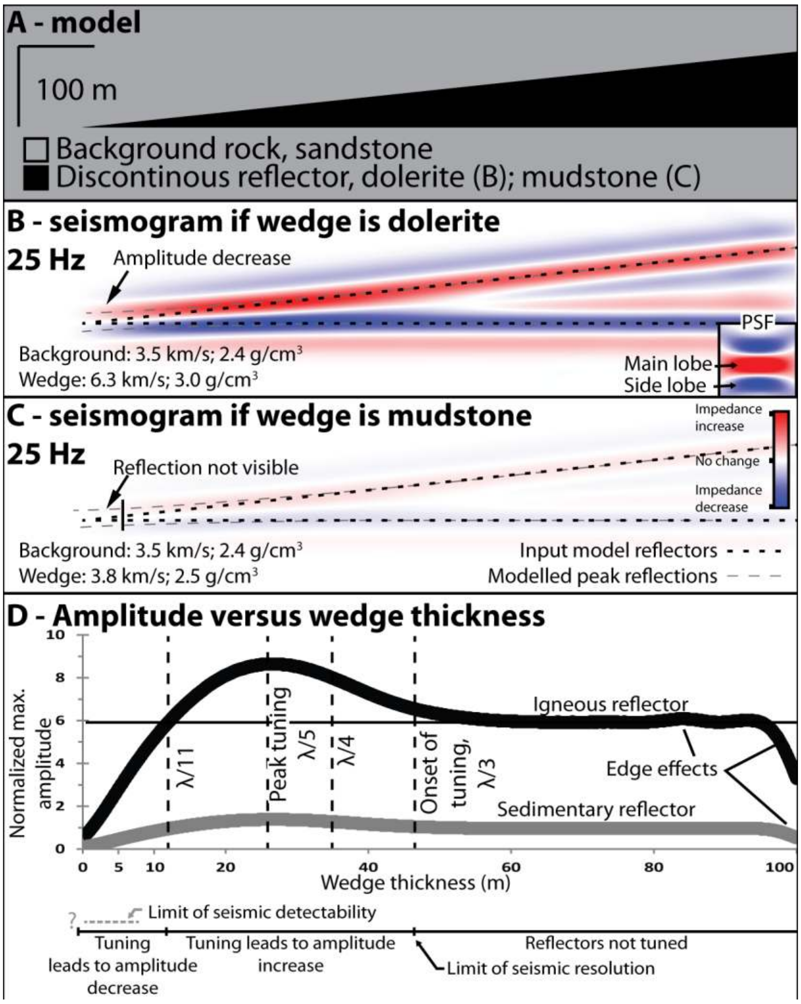


**Fig. 11:** Wedge-model illustrating the concepts of seismic tuning, seismic resolution and seismic detectability. Wavelength ( λ ) = 140, λ /4 = 37.5 m. A) Input model geometry. The model is 1 km long, and contains a 900 m long wedge thinning towards the left from 100 to 0 m. B) Resulting seismogram at 25 Hz using the inset PSF for a wedge consisting of dolerite within sandstone. Note the clearly resolved top and basal reflectors for wedge thicknesses from 100-45 m, the increased amplitude and overestimated thickness due to seismic tuning from 45-12 m wedge thickness, and the rapid amplitude decrease for wedge thicknesses of 12-0 m. C) B) Resulting seismogram at same conditions and amplitude scale for a mudstone wedge within a sandstone. Note the decreased amplitude compared to (B), and the lack of visible reflections at wedge thicknesses smaller than c. 10 m. D) Graphs showing normalized maximum amplitude per trace for (B) and (C). Note the similarity in the seismic signal for intrusions which are c 40 m thick and c. 10m thick, indicating that relatively thin intrusions such as the ones in this study would be nearly indistinguishable from thicker intrusions in seismic data as long as the intrusions are thinner than seismic resolution.

As the wedge thins further than the peak tuning thickness of λ/5, there is an overall decrease in the amplitude strength, first because the side-lobe and main lobes interfere in a non-optimal way, and then as the oppositely-directed upper and lower main lobes starts to interfere with each other. In the modelling run presented in Fig. 11, there is a gradual decrease of amplitude strength as the wedge thins further. It is important to note that the reflection amplitude at λ/11 (11 m) is equal to the reflection amplitude at the top of a dolerite reflector which is not tuned and therefore, theoretically, relatively thin intrusions should be imaged.

### 6.1.2 Seismic detectability of sill intrusions

The seismic detectability of an interface relates to whether a reflection from an interface can be recognized in seismic data or not. This is a much more difficult issue to address than seismic resolution, as it is a function of many more variables than the seismic resolution, some of which are poorly constrained and difficult to model. In order to detect a reflection, it needs to have a significantly higher amplitude than the baseline set by seismic noise in a survey. As seismic data is subject to significant amounts of processing to remove noise, it is hard to provide definitive answers without knowing each particular case, but we can offer some general considerations.

The amplitude of a reflection depends on the contrast in acoustic impedance (P-wave velocity × density) across the layer interface. It follows from the input values used in this study (Table 1), that doleritic intrusions have much higher P-wave velocity and density than their sedimentary host rocks. Thus, the amplitudes from the top of igneous intrusions are much higher than amplitudes from sedimentary reflectors. This is shown in Fig 11, where the amplitude of a reflector from a sandstone-dolerite interface is about 6 times higher than the amplitude of a sedimentary reflector, all other things being equal.

Assuming that the data quality of a seismic survey is good enough to image a siliciclastic sedimentary reflector at depth, it follows from the wedge model in Fig. 11D that it should be theoretically possible to image and detect reflections even from very thin dolerites (i.e. thicknesses down to a few meters). Furthermore, it also follows that sills with a thickness in the order of λ/11 (11m) would be imaged with the same reflection strength as an un-tuned reflection from a sediment-dolerite interface (Fig 11D). This implies that igneous intrusions with thicknesses as the ones seen in studied cliffs in East Greenland should be easy to detect in seismic. This is supported by Sheriff and Geldart (1995) who suggested that low-amplitude hydrocarbon-bearing sandstone within mudstones could be detected in "reasonable data quality" at thicknesses of λ/20 – λ/30. Because these *rules-of-thumb* are determined with reflections from siliciclastic reflections in mind, and because of the much greater acoustic impedance contrast between dolerite and siliciclastic rocks, we suggest that a limit of detectability of λ/30 is probably too pessimistic *in principle* for what can be detected in good-quality seismic data. However, in reality, combined seismic and well-log datasets from areas with igneous intrusions commonly reveal large amounts of un-imaged sill intrusions (Fig. 12). In the following sections, we investigate this.

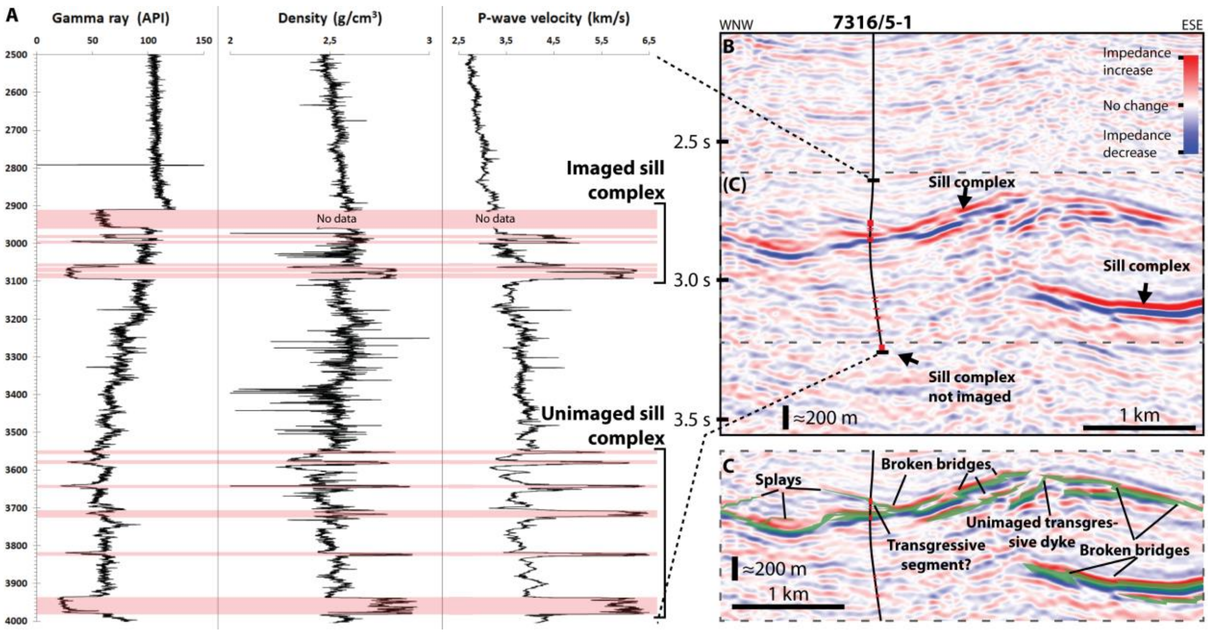


***Fig 12****: Comparison of well and seismic data of sill complexes. A) Data from well 7316/5-1 from the Vestbakken Volcanic Province on the W Barents Sea Margin. Red fields show depths of interpreted igneous intrusions. B) Uninterpreted seismic section around the well in (A). Note that only the upper sill-complex is imaged, although the lower sill complex is of similar thickness . Red fields on well indicate intrusions in (A). C) Interpreted section from B. Note the similarity of the actual seismic data to the synthetic seismograms generated from outcrop data from E Greenland, and that geometries on seismic reflectors corresponds to geometries observed in outcrop.*

### 6.2 Modelling limitations and seismic noise - When can we not see deep intrusions at all?

In all types of modelling, there are limitations and simplifications. The method applied here has significant advantages compared to methods used for modelling igneous intrusions in previous studies (e.g. Magee et al 2015), particularly because this model takes limitations in horizontal resolution into account, because it includes the effect of overburden on seismic imaging at the target, and because it more accurately represents steeply dipping features such as dykes. The method applied in this study also has three particular limitations (which would also be present in 1D convolution) that should be pointed out before inferences made from analysis of the synthetic seismograms are applied to actual seismic data.

### 6.2.1 Dykes and near-vertical interfaces

Near-vertical (> 70°) interfaces such as steeply dipping dykes are modelled as a series of vertically offset points rather than continuous reflectors (Fig 4B). This is because reflection coefficients are calculated vertically in a grid. Furthermore, the theoretical background of ray-based seismic imaging and processing assumes relatively flat layers. However, the dykes modelled in this study (e.g. Fig 5E) resemble interpreted dykes in seismic data from the NE Barents Sea (Minakov et al., 2017, their fig. 5), showing that the representation of dykes in the model achieves acceptable results.

Interestingly, in seismic data with low lateral resolution (e.g. Fig. 9D, E), stepping sills are imaged as oblique sheets. Also, because the impedance contrast between intrusives and host-rocks is so high, very little interference-effects between the layered host-rock and the oblique intrusions are observed. This is in contrast to modelling results reported previously lead to apparent steps (‘*pseudosteps*’) in the sill geometry in the seismic data, which were hypothesised to not be present in the subsurface (c.f. Magee et al 2015). This is almost opposite to the results we see in similar models. This is likely because the model employed in previous study employed 1D convolution, which dramatically overestimates horizontal resolution (c.f. Fig. 4, e.g. Magee et al 2015,). We therefore expect ‘pseudosteps’ to be of little importance, especially in seismic data below c. 1 km where horizontal resolution is low.

### 6.2.2 Noise

The simulated seismograms are generated without simulating acquisition noise. However, the simulated seismic images are not simply smoothed versions of the input models, but do contain several examples of discontinuous reflectors, dipping noise and amplitude variations which are not part of the input models, but that rather the results of simulation of diffractions occurring at discontinuities of reflectors. Fig 9E contains abundant examples of this.

In all models, the modelled reflections from the sill intrusions are coherent and show considerably stronger amplitudes than sedimentary reflectors. In all cases, even in Model D (Fig. 9E) which simulates imaging at 6 km depth, the reflections from the intrusives are detectable. In this case, the background seismic velocity is c. 4.6 km/s, the signal frequency is 10 Hz, and consequently, the wavelength of the signal is 465 m. This would mean that onset of tuning would occur at around 150 m, peak tuning would occur at 90 m, and onset of amplitude fall below the amplitude at no tuning would occur at 37 m. The sills in the study area are c. 10 m thick (λ/47), and would in this case give stronger reflections than an un-tuned sedimentary reflector. However, if acquisition noise was also modelled, would the intrusives be detectable? This is a difficult question to answer, because it is difficult to model acquisition noise in a rigorous manner (Scales and Snieder, 1998). The common way of adding noise to seismic sections could of course be performed, but the detectability of the reflections would depend on the amplitude of the noise, which would be entirely arbitrary because no good way of estimating the amplitude of noise in processed seismic data exists. Still, the modelling shows that as long as reflections from sedimentary interfaces are imaged, reflections from even relatively thin sills (more than a few meters) should be imaged as well.

### 6.3 Comparison to real data

A number of important issues are highlighted as the results from this study are compared to actual reflection seismic data.

#### 6.3.1 Thickness and tuning

The vast majority of published examples of sill intrusions imaged in seismic data occur as tuned reflectors (e.g. Cukur et al., 2010; Schofield et al., 2012b; Haafez et al., 2017; Schmiedel et al., 2017), with a few exceptions where also the basal reflection is imaged separately (Hansen and Cartwright, 2006b; Jackson et al., 2013; Schofield et al., 2015). Commonly, these tuned reflections are interpreted to originate from sills that have thicknesses that lie close to the assumed tuning thickness of c. $\lambda/4$. However, the wedge model presented in Fig. 11, which models the seismic response to a reflection from a sediment-dolerite interface, indicates that a sill with a thickness of $\lambda/11$ should be capable of generating a reflection with the same amplitude as a sill with a thickness of $\lambda/3$. These results imply that the majority of mafic sill intrusions are relatively thin compared to seismic resolution, and is in accordance with datasets of thicknesses of drilled intrusions (Schofield et al., 2015). Also, tuned reflections should not simply be assumed to be close to the tuning thickness.

In deep and low-frequency seismic data, there is also the possibility that what is imaged as one reflector is in fact a reflection of several stacked sills, as in the example of Fig. 5F. This illustrates that seismic reflection data have the potential to significantly overestimate the thickness of discrete sills in a basin.

#### 6.3.2 Un-imaged sills, and applications to seismic interpretation

The modelling and discussion above indicates that, theoretically, even thin sills should be possible to image in seismic reflection data, even at relatively deep (> 3 km's) basinal levels. However, many real world datasets combining well- and seismic data shows that large amounts of sills are in fact not imaged at all. For example, Schofield et al (2015) demonstrated that only a small number of the sills identified in well data from the Faroe-Shetland Basin are actually imaged in seismic data. A similar relationship can be seen in seismic data from the Vestbakken Volcanic Province at the western Barents Sea Margin (Figs. 1, 12). Exploration well 7317/5-1 encountered 12 sills which are clearly identifiable in wireline well logs (Fig. 12A, (Omosanya et al., 2016). These igneous intrusions make up an upper and a lower sill complex. While the upper complex is relatively well imaged in seismic data, the lower sill complex, consisting of intrusions ranging in thickness from 5 to 43 m, is not imaged at all (Fig. 12B). There may be several reasons for this:

Although single continuous bodies of dolerite have a great potential for imaging, real sill complexes do not consist of single continuous bodies, but rather several complex, interconnected sills and thinner sill splays (e.g. Schofield et al., 2015, Eide et al., 2017; Fig. 13). In the case of thick sill complexes with large amounts of thin intrusions (c.f. Fig. 13E, Schofield et al., 2015, their fig. 6B), seismic imaging will result in a strong reflection at the top of the sill complex, as the average velocity and density increases. Below the top, reflections from several thin intrusions interfere with each other, and yield no coherent reflections (Fig. 13E). Furthermore, although there is a strong acoustic impedance contrast at each sediment-dolerite interface, these occur more closely spaced than seismic resolution, leading to smaller effective impedance contrasts and therefore low seismic amplitudes.

Additionally, the lateral velocity structure can be highly variable through a sill complex (see Fig. 13E). This means particular velocities used in migration are highly variable laterally, which is not ideal when working with time migrated data, as imaging is often made worse by a poor choice of velocity models, leading to migration artefacts which may fully mask underlying, weaker reflection from sills.

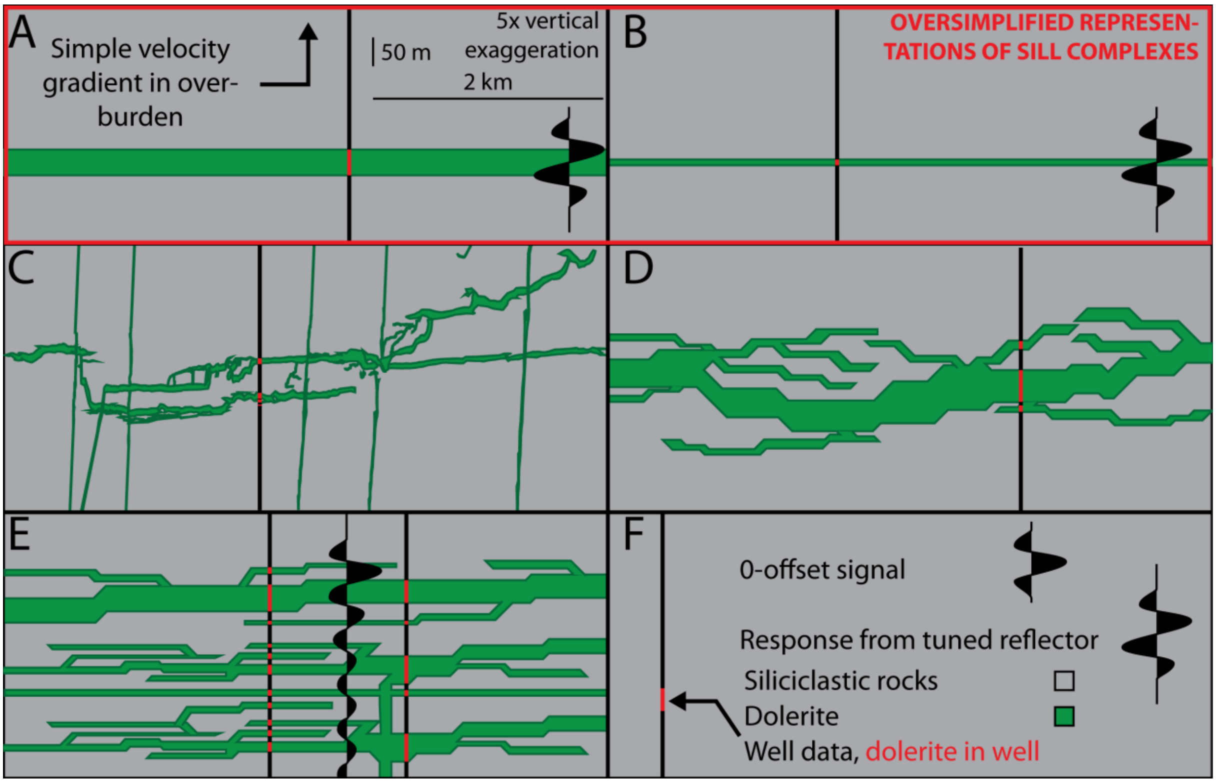


***Fig 13:*** *Comparison of conceptual geometries of sill complexes, where individual sills are below seismic tuning thickness. A) Simple and unrealistic tabular sill, just below tuning thickness. B) Simple and unrealistic tabular sill, with a thickness of ¼ of the tuning thickness. Note that these are imaged at comparable amplitudes. C) Real-world example of a small- volume sill complex. D) A larger-volume sill-complex. Note large lateral variability in thickness and amount of dolerite. E) High-volume sill complex consisting of abundant thinner sills and splays. This example would likely give a strong reflection at the top, and weak and incoherent reflections below the top due to interference below seismic resolution. F) Legend.*

## 6.4 Future Perspectives

Considerable amounts of work have been done on the so-called 'sub-basalt imaging problem', which has yielded good results (c.f. Gallagher and Dromgoole, 2007). However, significantly more sedimentary basins contain extensive igneous sills, than large basaltic piles (e.g. South Australian Margin; Holford et al. 2012). Despite this, there is little awareness of issues relating to imaging of sill intrusions, and in particular imaging below sill complexes or the 'Sub-Sill Imaging Problem'. As accurate basin modelling relies on a good appreciation of source rock extent and thickness within a sedimentary basin, and given that in many sedimentary basins, potential source rock intervals (and occasional reservoirs) are inferred to occur below sill complexes, it may be time to properly address the imaging issue and the 'Sub-Sill Imaging Problem'.

Importantly, synthetic seismic modelling presented in this paper indicates that relatively thin igneous sills could potentially be imaged in reflection seismic data. However, inspection of available datasets shows considerable migration artefacts below shallow intrusions, which points towards poor representation of sills and geometries in velocity models. This is due to the low thickness, high velocity, complex structure, and overall velocity changes that are difficult to represent in velocity models. Depth migrated data is potentially better at improving imaging, such data is only as good as the initial velocity model used, and that model is only as good as the understanding of the sub-surface geology. Therefore, improvement of this requires a greater understanding of where sills will appear in sedimentary basins, better methods to predict architecture of sill complexes, , in order to allow better informed velocity models to be attained. Until now, conceptual understanding of sill complexes drawn from analysis of field data has only been applied for the interpretation of seismic data. Such insights could also lay the foundations for progress in processing and imaging and addressing of the 'sub-sill imaging problem'.

Imaging of igneous dykes is probably hampered by a lack of awareness of the importance of these. Such discontinuous features, when viewed in a processing phase, may be simply regarded as seismic artefacts and subsequently removed. This is a problem, as these can act as significant barriers or conduits to hydrocarbons and other basinal fluids, depending on post-emplacement evolution (e.g. Rateau et al. 2013; Senger et al., 2015; Eide et al., 2017). Seismic modelling performed in this contribution indicates that these could potentially be imaged in seismic data, and tentative interpretations of dykes have been made in regional seismic lines (Minakov et al., 2017).

## 7. Conclusions

This study has investigated how mafic sill-complexes emplaced within sedimentary basins are imaged in reflection-seismic data at a variety of depths and under a variety of overburden conditions. This is important for seismic interpretation of sill complexes, which presents significant challenges for seismic interpretation, geologic forecasting and drilling. This has been done by generating synthetic seismograms using a 2D convolution method on a seismic-scale, sill-bearing outcrop-section from east Greenland, performing ray-based modelling to investigate lateral resolution at target depth for a variety of overburdens and target depths, and by comparing the synthetic seismograms to actual seismic. These are the main findings:

1) It is important to be aware of the illumination at the target depth, as low illumination leads to low maximum imaged dips and low lateral resolution will make it impossible to image steeply dipping features such as oblique dykes which might have important reservoir implications. 1D seismic convolution does not take horizontal resolution into account, and consistently overestimates horizontal resolution.
2) Igneous rocks, such as basalt flows or shallow intrusions, occurring between the imaging target and the seismic source and receivers will decrease vertical and particularly horizontal resolution at the target depth by absorbing high frequencies, seismic energy and diverting low-incidence seismic rays.
3) Commonly used rules-of-thumb for seismic detectability likely underestimate the minimum thickness of beds of mafic intrusions may be imaged in reflection seismic data because of the high acoustic impedance contrast between mafic intrusions and siliciclastic rocks.
4) The thickness of sills, either thick or thin, is no guarantee that they might be imaged. Migrated seismic data has the potential to image thin sills if the velocity models are good and sills are not so close that they interfere with each other
5) In reflection-seismic data of a reasonable quality, subseismic sill emplacement features such as splays, broken bridges and transgressive dykes may be inferred from sill geometries and amplitude variations.
6) Apparent steps in seismic data arising from interference between a layered host-rock and oblique dykes is likely not to occur in actual seismic data below c. 1 km depth because of limited horizontal resolution.

7) Improved imaging of sill complexes is important and may be achieved by integrating conceptual understanding of sill complexes and improved velocity models.

## 8. Acknowledgements

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# 11. Tables

Table 1: Input data for modeling. NCS, Norwegian Continental Shelf; MD, measured depth.

***a) Values for target at 1 and 3 km depth***

| Name | Facies association | $V_P$ | $V_P/V_S$ | Density | Source | Depth |
|---|---|---|---|---|---|---|
| | *#* | *km/s* | *fraction* | *g/cm3* | *NCS well/article* | *MD (m)* |
| Igneous intrusions | 1 | 6,3 | 1,86 | 3,0 | Smallwood and Maresh, 2002 | |
| Homogeneous sandstone | 2 | 3,7 | 1,80 | 2,5 | Add 5% to FA3 | - |
| Sandstone | 3 | 3,5 | 1,80 | 2,4 | 6407/2-1 | 3220-3235 |
| Poorly cemented sandstone | 4 | 3,4 | 1,80 | 2,3 | Subtract 5% FA3 | - |
| Heterolithic | 5 | 3,3 | 1,80 | 2,3 | 6407/2-1 | 3030-3050 |
| Mudstone | 6 | 3,8 | 1,80 | 2,5 | 6407/2-1 | 2920-2940 |
| Organic rich mudstone | 7 | 2,6 | 1,80 | 2,3 | 6407/2-1 | 2885-2900 |

***b) Values for target at 6 km depth***

| | | | | | | |
|---|---|---|---|---|---|---|
| Igneous intrusions | 1 | 6,3 | 1,86 | 3,0 | Smallwood and Maresh, 2002 | |
| Homogeneous sandstone | 2 | 4,8 | 1,80 | 2,7 | Add 5% to FA3 | - |
| Sandstone | 3 | 4,6 | 1,80 | 2,6 | 6506/12-10A | 5525-5545m |
| Poorly cemented sandstone | 4 | 4,4 | 1,80 | 2,4 | Subtract 5% FA3 | - |
| Heterolithic | 5 | 4,7 | 1,80 | 2,5 | 6506/12-10A | 5565-5582m |
| Mudstone | 6 | 4,7 | 1,80 | 2,4 | 6506/12-10A | 5618-5635m |
| Organic rich mudstone | 7 | 3,9 | 1,80 | 2,5 | 6506/12-10A | 5330-5335m |