Seismic velocity structure of seaward-dipping reflectors on the South American continental margin

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1 Abstract

Seaward dipping reflectors (SDRs) are a key feature within the continent to ocean transition zone of volcanic passive margins. Here we conduct an automated pre-stack depth-migration imaging analysis of commercial seismic data from the volcanic margins of South America. The method used an isotropic, ray-based approach of iterative velocity model building based on the travel time inversion of residual pre-stack depth migration move-out. We find two distinct seismic velocity patterns within the SDRs. While both types show a general increase in velocity with depth consistent with expected compaction and alteration/metamorphic trends, those SDRs that lie within faulted half grabens also have high velocity zones at their down-dip ends. The velocity anomalies are generally concordant with the reflectivity and so we attribute them to the presence of dolerite sills that were injected into the lava pile. The sills therefore result from late-stage melt delivery along the large landward-dipping faults that bound them. In contrast the more outboard SDRs show no velocity anomalies, are more uniform spatially and have unfaulted basal contacts. Our observations imply that the SDRs document a major change in rift architecture, with
magmatism linked with early extension and faulting of the upper brittle crust transitioning into more organised, dike-fed eruptions similar to seafloor spreading.

**Keywords:**

Continental breakup; Volcanic passive margins; Seaward-dipping reflectors; tomographic inversion

## 2 Introduction

Seaward-dipping reflectors (SDRs), with their characteristic wedges of oceanward-diverging reflectors, were first recognised on seismic profiles in the 1980s (Hinz, 1981; Mutter et al., 1982). They have since been shown to be a common feature within the continent to ocean transition zone of volcanic passive margins worldwide (e.g. Franke, 2013). Drilling showed them to consist of sub-aerially erupted basaltic lava flows interbedded with thinner layers of volcanic tuff and terrestrial sediment (Roberts et al., 1984; Eldholm et al., 1987; Larsen et al., 1994). Several studies have shown how this layering causes seismic scattering, complex multiples and mode conversions (Planke et al., 1999; Ziolkowski et al., 2003). On mature margins the SDRs are generally buried beneath kilometres of post-rift sediments which often results in interference from the sea-bed multiple and further limited their early interpretation. These issues combined to restrict early attempts to image their lowermost parts and their important contacts with the basement. The geological process responsible for the characteristic SDR geometry and the nature of the underlying rift architecture is therefore an outstanding question.

Due to the seismic imaging issues summarised above and the restriction of most boreholes to the upper parts of the most landward sequences, much of our current understanding of
SDRs draw on observations from rare onshore exposures and modern-day analogues. For example, exhumed flanks along southeast Baffin, west Greenland and east Greenland display arrays of landward-dipping normal faults with kilometre-scale offsets that are intimately associated with syn-rift volcanism (Geoffroy et al., 2001; Geoffroy, 2005). Along southeast Greenland, where the depth of erosion is deepest, gabbroic bodies which are interpreted to represent the solidified and depleted magmatic source of the overlying lavas are found within the continental basement (Myers, 1980; Brooks and Nielsen, 1982; Klausen and Larsen, 2002). These observations suggest that syn-rift volcanism and footwall rotation could be responsible for SDR formation in the early break-up stages at least. This idea has been supported by the limited offshore drilling in the North Atlantic (Gibson and Love, 1989; Larsen et al., 1994). However, critics of this model for the origin of SDR geometries cite the lack of evidence for large landward dipping faults within active rift areas such as Ethiopia (Corti et al., 2015).

The earliest published SDR formation models used comparisons with the Tertiary lava sequence in eastern Iceland to emphasise the role of volcanic loading onto a weak lithosphere in the generation of accommodation space and reflector dips (Hinz, 1981; Mutter et al., 1982). This flexural mode of SDR formation by steady-state, sub-aerial seafloor spreading has more recently been developed by Buck (2017) and Morgan and Watts (2018). In these models the flexural load consists of solidified feeder dykes added to the edge of a thin elastic plate. Distinct SDR wedges are formed by jumps in the location of dyke injection or hiatuses during the continental extension process. In the current scientific literature these two mechanisms (footwall rotation and magmatic loading) have been regarded as
competing. Establishing the correct mechanism clearly has important implications for the rheology of the continental lithosphere during the transition to seafloor spreading.

A recent impetus from deep-water hydrocarbon exploration has resulted in the next generation of 2D seismic imagery being collected which is pushing forward our knowledge of the process of continental breakup at volcanic margins (e.g. Quirk et al., 2014; Stica et al., 2014; Abdelmalak et al., 2015). By using long hydrophone streamers and deep-towed sources rich in low frequencies it is now possible to resolve the internal structure and even penetrate thick SDR sequences. For the first time conclusive evidence for fault-bounded SDRs has been documented offshore, with several authors proposing hybrid models containing a combination of faulting and flexure as controls on SDR geometry (Quirk et al., 2014; Paton et al., 2017).

A promising aspect of the new generation data is the ability to extract robust seismic velocities from the extrusive volcanic sequences to aid their interpretation (e.g. Ogilvie et al., 2001; Spitzer et al., 2005; Goncharov and Nelson, 2012). An advantage of this method is that it produces results that are typically an order of magnitude finer than conventional wide angle imaging using seabed receivers and surface shots (e.g. White et al., 2008). Within the continent-ocean transition zone seismic velocity is a particularly sensitive parameter to rock lithology (igneous and sedimentary) together with porosity, pore aspect ratio and alteration.

The aim of this study is to apply an automated pre-stack travel-time tomographic inversion to two long-offset, surface-towed seismic reflection profiles from the volcanic margins of South America. To our knowledge an analysis of this type has not previously targeted SDR sequences across an entire volcanic continental margin. By determining the velocity
structure we will gain insights into the origins of SDRs and hence place new constraints on
the process of continental breakup.

3 Previous work

We previously presented an interpretation of a set of industry-standard, 2D seismic
reflection profiles and magnetic anomaly data from the volcanic margins of the South
Atlantic (McDermott et al., 2018; Collier et al., 2017). The seismic dataset was acquired by
ION Geophysical between 2009 and 2012 with a 10.2 km long-offset 408-channel streamer,
a 50 m shot interval and a tuned air gun array with a volume of 6420 in³. The complete
seismic survey comprises 25 dip- and 15 strike- pre-stack, depth-migrated (PSDM) sections,
with a total line length of almost 22,000 km that covers offshore Argentina, Uruguay and
southern Brazil (the Pelotas Basin). The Paraná continental flood basalt province lies
onshore the northern part of the study area (Figure 1).

The SDRs extend near continuously from southern Argentina to the Pelotas Basin (Figure 1).
They have an across margin width of 70-120 km in the south which increases dramatically to
> 300 km in the north. Several boreholes have penetrated the most-landward parts of these
sequences and have confirmed the presence of basaltic flows with weathered tops,
interbedded with terrestrial sediments (Gordon and Mohriak, 2015). Individual flows are
typically between 0.5 and 18 m thick, which is less than the 20-40 m vertical resolution of
the seismic reflection data. This implies the reflectors themselves do not represent
boundaries between individual “hard” lava flows and “soft” interbedded sediments but
from groups of these layers constructively interfering (Planke and Eldholm, 1994).
In our recent study we chose not to apply the “inner” and “outer” SDR terms established by Planke et al. (2000). Instead, we introduced new terminology to describe the characteristics of the SDRs: Type I and Type II. An important distinction is that offshore South America there are not two spatially distinct sets of SDRs, but rather they onlap each other. This distinction has been previously noted by Franke et al. (2010) and Soto et al. (2011). Due to the association of the SDRs with other volcanic features such as lava deltas, we interpret them all as subaerial. A summary of the key characteristics of the two SDR types is given in Figure 2. The seismic imagery is now of sufficient quality to interpret the presence of rotated fault blocks beneath the landward flows. We termed these SDRs as Type I to distinguish them from the more seaward Type II SDRs that are unfaulted and either stack on top of the Type Is or have a relatively flat contact with the underlying basement. Type I are both shorter and straighter reflection packages that form fault-bounded wedges. The Type II reflectors form concave downward, off-lapping reflection packages, with dips increasing at their distal ends (McDermott et al., 2018).

In terms of distribution, the Type I SDRs have uniform widths and thicknesses along the entire margin studied, whereas the Type II SDRs become significantly wider and thicker in the north, co-incident with the onshore Paraná flood basalt province (Figure 1). In this region, individual reflectors also become significantly longer and so we subdivide the Type II SDRs into IIa in the south and IIb in the north. The most likely explanation for the longer flow lengths in the north is higher topographic elevation due to anomalously hot asthenosphere at the time of breakup. This is consistent with the rapid increase in the volume of SDRs and also an increase in the thickness of the oceanic crust in this area (Taposeea et al., 2016). Independent interpretations of subsets of the commercial seismic
data used by us have been presented in the Pelotas (Stica et al., 2014) and Uruguay regions (Paton et al., 2017; Conti et al., 2017). These authors also document the presence of both faulted and non-faulted SDRs.

For our seismic velocity inversion analysis we selected two multichannel seismic reflection profiles that represent the key characteristics of the SDRs along the margin (Figure 3). For both lines the post-rift does not present any specific imaging challenges, such as salt, free gas, channels etc., so we could focus on model building within the SDRs. The quality of the depth imaging on both lines is demonstrated by the presence of significant reflectivity below the SDRs, either in the form of lower-crustal reflectivity and/or what is interpreted as Moho. This allows a confident interpretation of the base of the SDRs where the reflectivity abruptly terminates. Line A is from offshore Uruguay and displays three packages of classic Type I SDRs, the most oceanward of which is partially overlain by a volumetrically modest Type IIa sequence with characteristically short reflectors. The SDRs reach a maximum thickness of around 4 km, which is in marked contrast to Line B, which lies immediately offshore the Paraná of southern Brazil, where they exceed 12 km. Two packages of Type I SDRs are also visible on Line B (x=35-80 km, Figure 3), but they are both overlain by thick Type IIb SDR units with characteristically long reflectors. This profile crosses a tectonic transfer zone that developed into the Chui Fracture Zone once seafloor spreading got underway (Figure 1). This zone is clearly seen in the seismic by uplifted basement blocks and results in a set of landward dipping SDRs (x=185-210 km).

4 Method

The seismic velocity models were generated with a PSDM tomographic method (Jones, 2010). This solves for the travel times between the source and receivers by finding the
isotropic velocity model that best flattens primary reflection events across offsets within common reflection point (CRP) gathers. The starting velocity model for each profile consisted of a combination of a mature tomographic inversion model for the post-rift sediments with one developed from manual semblance picks across the SDRs every 750 – 1000 m. The manual part of the model was smoothed laterally and vertically to remove any local artefacts and the maximum velocity was capped at 6 km/s. This removed any potential biases towards high velocities along the base of the SDRs.

The raw time gathers were pre-stack depth migrated with a Kirchhoff algorithm in shortest ray-path mode. Tests showed that a migration aperture of 6000 m and a maximum dip limit of 65° ensured that most of the primary SDR signal was included. The CRP gathers were then conditioned to reduce multiples and high-frequency noise within the SDRs. A Tau-P Radon multiple suppression scheme was used with a P-range of 3000/-100 ms and a P-cut value of 150 ms at 10.2 km offset (Figure 4a-c). This was followed by a 3-6-30-40 Hz Ormsby bandpass filter. Finally, an angle-mute of 45°, combined with an x-z mute to reduce NMO-stretch at deeper levels was applied (Figure 4d-e). The coherency (or semblance) was then calculated at each depth point across all offsets, with a value of 1 indicating perfect horizontal alignment.

The residual move-out for each horizon across all offsets was then calculated using a wavelet tracking technique (Jones, 2010). Horizon seed points were made on the zero-offset trace on every fifth CRP (75 m) if its amplitude exceeded a given threshold and the coherency exceeded 0.5. In general, the vertical spacing between seed points within the SDRs was on the order of 50 - 100 m. The residual move-out picks were only used if the autotracker could follow an event for least 70% of the total (muted) offset.
The selected move-out residuals were then input into a tomographic inversion scheme that uses a conjugate-gradient (damped least squares) solver to update the velocity field (Fruehn et al., 2014). The method incorporated “hard constraint layers”, above which the velocity was fixed during the inversion. This allowed a “layer stripping” approach to be applied, with successively deeper levels within the post-rift sediment overburden being fixed as the inversion progressed. A 500 m (20 CRP) horizontal by 200 m depth cell size was found to give the most stable inversion results for the target (SDR) region. The velocity solver accounts for both the dip and depth of a reflection by scanning through an appropriate stacked image (made with either the previous tomographic model or a model update derived from 2nd order residual move-out picks) within a user defined window. The algorithm assumes that structures within this window can be approximated by a plane. If the assumption is broken, it will result in incorrect dips and poor signal coherency which will, in turn, reduce the residual pick density within each inversion cell. It is therefore better to use a smaller window in areas of high dip. A relatively small window size (7 CRPs x 35 m) was used to account for the higher dips present towards the base of the SDR sequence. Events with dips greater than 40° were ignored during updates.

After each iteration the quality of the output model was visually checked for the presence of anomalous velocity contrasts such as localised horizontal or vertical striping. Most irregularities were small and could be corrected for by smoothing the model or by increasing the cell size during inversion. The change in the flatness of the CRP gathers was also visually reviewed (Figure 5). A better model produced flatter gathers to longer offsets whilst maintaining geologically reasonable velocities. If the quality of the CRP gathers deteriorated, the residual depth picks would be assessed and the inversion re-run. Once a
velocity model update was deemed acceptable it was used as input for the next
tomographic run. In practice, the process of residual move-out tracking, tomography and
remigration was stopped after three iterations as this was judged to result in optimal
flatness of the gathers across the profiles. In order to highlight where the seismic velocity
field was well-constrained by the data the final coherence values were averaged onto a
10x0.6 km grid and a bulk coherency threshold of 0.3 used in our interpretation.

5 Results

The final interval velocity models are presented as underlays to the pre-stack depth-
migrated seismic sections to show their relationship with the reflectivity in Figures 6 and 7.
Enlarged versions, together with interpretations, are given in Figures 8 and 9. In these
figures the bulk coherence threshold is marked with the white dashed line. Velocities below
this line are not constrained by the tomographic inversion and should be ignored. As can be
seen the entire SDR sequence for Line A is above the coherence threshold, and most of the
SDR sequence for Line B. To compare velocity-depth trends we extracted vertical profiles
across the SDRs at 1 km along-profile intervals and present them individually (Profiles i-iv,
Figures 8d, 9d) or stacked (Figures 8e, 9e).

The post-rift sedimentary sequence displays velocities increasing from 1.8 to 4.5 km/s with
depth, consistent with expected compaction trends for terrigenous sediment (Hamilton,
1979). Overall, the velocity gradient is similar throughout but because sediment thickness
ranges between 3 and 6.5 km the velocity contrast at the top of the SDRs varies. In places
there is a small positive velocity step as the SDRs are entered (e.g. Profiles i and ii, Figure 8),
but elsewhere there is a smoother transition and the picking of the Top SDR horizon relied
on a change of reflectivity character. The seismic velocity within the uppermost SDRs ranges
between 3.0 and 4.8 km/s, with higher velocities where the SDRs are more deeply buried (Figures 8e, 9e). The maximum seismic velocity recorded in the SDRs is 7 km/s.

Within the upper 1-2 km of SDRs, there is a steep velocity gradient (0.5-1 /s) everywhere. When the velocity reaches ~5.5 km/s the gradient reduces to 0.25-0.5 /s. However, whilst the upper parts of the SDRs show similar velocity-depth characteristics, the lower parts do not. Instead we observe two contrasting velocity patterns which directly correlate with the two SDR types we identified previously. The Type I SDRs on Line A are laterally variable with three distinct velocity highs at their down-dip ends (Figures 6 and 8). Individual velocity highs are typically 4-6 km wide, 2-3 km thick and spaced 10 km apart, with velocities reaching 6.5-6.8 km/s. We are confident about the presence of these high velocity features as they were also identified during the manual semblance analysis. They were deliberately removed from the starting model by capping at 6 km/s but were consistently reintroduced by the tomographic inversion. The locations of these high velocity zones correlate with the basement-bounding faults interpreted from the reflectivity alone. Note that this pattern is not seen within the Type I SDRs imaged on Line B (Figures 7 and 9), but this is entirely consistent with their much deeper burial and subsequent rotation by the huge volumes of the later Type IIb SDRs, which may have resulted in the velocity contrasts at the down-dip portions being undetected.

In contrast to the Type I SDR pattern both the Type II SDRs continue to show a moderate positive velocity gradient with depth (Figures 8d and 9d), with sub-horizontal velocity contours that crosscut the dipping reflectors (Figures 8b and 9b). Distinct velocity highs are absent. This uniformity of the Type II velocity-depth profiles can be readily seen from the
narrow range within the stacked profiles of Figures 8e and 9e, which are in marked contrast
to the broad spectrum of velocities obtained across the Type I SDRs.

The velocity inversion has also provided new constraints on the nature of the top ~5 km of
the basement below the SDRs for the first 50 km or so of the continental (western) ends of
both profiles. Overall the continental velocities are about 0.5 km/s higher on Line B than
Line A. This may be a result of the thicker sedimentary sequence closing micro-fractures and
pores within the basement or higher volumes of igneous intrusion within the continental
crust. The latter interpretation is supported on Line A by the presence of two high-velocity
regions where interval velocities reach 6.8 km/s (x=25-50 km, z=5-12 km). Similarly, on Line
A a ~50 km portion of oceanic crust (x=145-170 km), with seismic velocities reaching 7.2
km/s within the semblance threshold on the eastern end of the line. Line B terminates
before full oceanic crust is developed.

6 Discussion

6.1 Result validation

In our study we have used software routinely applied by us for the commercial imaging of
hydrocarbon targets. In many cases, the validity of the method has therefore been
independently confirmed (e.g. Fruehn et al., 2014). Many of the previously published
studies that extract velocities from surface-towed hydrophone data have focussed on the
so-called “sub-basalt imaging” problem on the NW European shelf (notably in the Faeroe-
Shetland area). Here thick, flat-lying basaltic layers overlie potential hydrocarbon reservoir
rocks (e.g. Fliedner and White, 2001). Various approaches to velocity extraction have been
used and in some cases the results could be directly compared to downhole sonic logs that
had been calibrated with VSP/check-shot surveys. These comparisons are typically within
0.1 km/s (3-5%), and we expect our accuracies to be similar (Ogilvie et al., 2001, Fruehn et al., 2008).

For our study, independent validation of the seismic velocity results across the SDRs comes from wireline logs in the region of Line B (Gordon and Mohriak, 2015) and BGR04-REFR01 wide-angle seismic profile that is co-incident with Line A (Becker et al., 2014). In the boreholes (see Figure 1 for locations) the wireline logs show lava flow interior velocities range between 4.6-6.2 km/s, with flow tops and interbedded sediments falling to 3.0 km/s. Given the vertical resolution in our new models is 200m and individual lava flows are 0.5-18 m thick, the velocities extracted from the seismic data will depend on the relative proportions of flows and sediment interbeds together with the thickness of the flows. The seismic velocities we have obtained at the landward-top of our SDR sequences on Line B (e.g. profile iv 4.7 km/s, Figure 9d) are therefore entirely consistent with the available wireline-logs.

In the BGR04-REFR01 wide-angle seismic model velocities range between 4.5 km/s at the top to 6.5 km/s at the base of the SDRs (Becker et al., 2014). These values compare favourably with our Line A, where they range between 3.5 and 6.8 km/s. To compare the results in more detail we show a velocity-depth profile extracted from the wide-angle model at the location of our profile ii (black line, Figure 8d). As can be seen there is an overall good match between the results, both in the sediments and SDRs, with the two profiles differing less than 0.5 km/s. Receiver spacing for the wide-angle experiment was 20 km, and so its velocity model did not detect any lateral velocity variation across the SDRs and specifically any velocity highs. We note however that our seismic model shows slightly faster SDR velocities throughout than the wide-angle model. Possibly this is a result of anisotropy i.e.
the variation of seismic velocity with direction of travel between our (near vertical) and wide-angle (near horizontal) acquisition. SDR sequences have been shown to exhibit transverse anisotropy on the order of 10-15% (Planke et al., 1999). This possibility however would need to be confirmed by comparison with a higher quality wide-angle profile with closer instrument spacing.

6.2 Composition, compaction and alteration trends

In order to compare the velocities of the SDR sequences across the margins and to try to separate the effects of composition changes from compaction and alteration trends in the data we followed the approach taken by Goncharov and Nelson (2012) and converted the extracted profiles from depth to lithospheric pressure. We use a simple 3-layer model (water, sediment and SDR with densities of 1000, 2200 and 2800 kg/m³ respectively) to correct for the variation in the overburden thicknesses (Figure 10). This approach shows that although the velocities are lowest at the top of the Type IIb SDRs, this unit has the lowest overburden pressure. When compared at equivalent pressures, the Type IIa SDRs actually display the lowest velocities i.e. at the top of Type IIa (150 MPa) velocities are between 4.3-5 km/s compared to 5.5-6.3 km/s for Type IIb at the same pressure. The simplest explanation for this difference is an increase in the relative volcanic component and/or an increase in average flow thickness towards the Paraná volcanic province. The velocities of the Type I SDRs are the most variable, suggesting that there is much greater heterogeneity within these SDRs. This is consistent with their landward location and association with large normal faults in a sub-aerial rift environment that would be expected to deliver higher levels of spatially variable terrigenous sediment compared to the Type II
SDRs which according to our interpretation are equivalent to subaerial seafloor spreading in the later stage of breakup.

In order to further investigate the seismic-velocity depth trends we made a comparison with seismic velocity profiles obtained from a field experiment in Iceland (Darbyshire et al., 1998). Here, because of the mid-ocean setting continental sediments can be assumed to be minimal and so the trends can be interpreted as a response of extruded basalt to compaction and alteration. On Iceland, the overall pattern of the velocity-depth trend is very similar to that seen in our Type IIb SDRs. This further supports the idea that the difference between the seismic velocity of the Type IIa and IIb SDRs results from an increase in lava content northwards along the margin. The very different Type I SDR trend reinforces the idea that these early SDRs were formed by a different mechanism than the subaerial seafloor-spreading that generated both the Type II SDRs and the upper Icelandic crust.

The Type II SDRs document a change in velocity gradient at about 200 MPa. The Icelandic profiles have a similar pattern, but with an inflection point slightly shallower (160-180 MPa). The frequency bandwidth of the seismic reflection data we used is 5-50 Hz (at the 20 dB level) within the SDR sequence, giving a limiting vertical resolution of 25 m. Within this level of resolution the reflectivity images give no indication of any change in the nature of the basaltic flows and interbeds, so we interpret the velocity-depth patterns in terms of compaction and alteration trends. In oceanic crust there is a similar seismic velocity gradient change at the Layer 2/3 boundary which has also been interpreted as an alteration front within the sheeted dykes (Detrick et al., 1994). We interpret the rapid velocity gradients at the top of the Type II SDRs as due to compaction (i.e. reduction in porosity) of the volcanic-sedimentary sequence in response to increasing burial depth. Borehole observations have
shown that mechanical compaction within lava flows is generally minimal, so much of the
change is likely to be due to the sedimentary interbeds (Planke and Eldholm, 1994). In the
mid-to-lower part of the Type II SDRs, the smaller gradients are likely to be due to
metamorphism under zeolite to granulite facies. The maximum velocity (~7 km/s) within the
deepest Type IIb SDRs is consistent with the values measured by Christensen (1996) for
greenschist facies basalt at 400 MPa (6.88 +/-0.2 km/s). It seems likely therefore that the
velocity structure observed in the Type II SDRs results from (i) diagenetic compaction of the
volcanic/sedimentary sequence (upper SDRs) followed by (ii) gradual alteration of the
basalts to low-grade zeolite facies basalts (middle SDRs) and greenschist facies basalts
(lower SDRs). Similar conclusions were made by Mjelde et al. (2007) from wide-angle
seismic models of the Norwegian margin.

6.3 Type I high velocity bodies
To our knowledge velocity highs have not been previously detected at the seaward end of
any SDR packages worldwide. Given the velocity highs exceed the compaction and alteration
related velocity trends established by the Type II SDRs, the most plausible explanation for
them is either a lateral change in the ratio of volcanic to sedimentary rock, the presence of a
different volcanic rock type and/or mode of emplacement (i.e. not extrusive). Velocity in
these regions reach 6.5 - 6.8 km/s which is consistent with the measured velocity of dolerite
(6.71 +/- 0.27 km/s) at similar pressures (Christensen, 1996). We favour an interpretation of
the presence of a different rock type because of the geometrical relationship with the SDR
packages. The velocity highs are associated with continuous reflectors and so our preferred
interpretation is that they are late-stage sills injected deep within the lava flow pile. Similar
bodies have been imaged within sedimentary sequences in the Voring Basin of Norway.
Berndt et al., 2000). Here up to 100 m thick olivine-gabbroic sills have been sampled and shown to give local seismic velocity anomalies of up to 7.4 km/s. They tend to occur in groups, within a vertical interval of 2-3 km and individual sill lengths of up to 5 km. These dimensions are similar to the high velocity zones seen in our images (Figure 8). In our case, late-stage injection into the lava pile would provide the conditions for slow cooling and so growth of large crystals that increases the seismic velocity relative to the surrounding extrusive basalts. The magma would have been sourced from a plumbing system that exploited the higher permeability along the fault systems that now bound the Type I SDR packages and explains their restriction to the down-dip portion of the SDRs. This interpretation is consistent with our analysis of the magnetic properties of this region, where Type II SDRS were found to correlate with linear magnetic anomalies but Type I SDRs did not (Collier et al., 2017). The eruption of Type II SDRs marks a period of more-organised volcanism, where linear fissures can be maintained to feed significant volumes of magma. The two SDR types undoubtedly mark a transition from fault-dominated extension in the continental crust to magmatic-dominated extension in the oceanic domain.

7 Conclusions

We have generated velocity models across a set of SDRs from two commercial seismic reflection profiles in the South Atlantic using commercial in-house software and expertise. Our analysis has provided constraints on the physical properties of the material to supplement the reflector geometries provided by the stacked sections. The method used an isotropic, ray-based approach of iterative velocity model building based on the travel time inversion of residual pre-stack depth migration move-out. This is far superior to conventional semblance analysis in the time-domain, and provides velocity estimates of
Velocity models with a vertical and lateral resolution of 200 and 500 m respectively were generated together with formal uncertainty bounds. Our work shows:

- The two classes of SDR identified from their reflectivity and magnetic properties (McDermott et al., 2018; Collier et al., 2017) correlate with distinct seismic velocity patterns. The velocities of the Type I SDRs are the most variable, suggesting that there is much greater heterogeneity within these SDRs.

- The Type II (non-fault bounded) SDRs display seismic-velocity depth trends consistent with compaction and alteration. We interpret the trend as due to (i) the diagenetic compaction of the volcanic sequence (upper SDRs) followed by (ii) the gradual alteration of the compacted basalts to low-grade zeolite (middle SDRs) and then greenschist facies basalts (lower SDRs). Those closer to the onshore Paraná flood basalt province (Type IIb) have higher pressure-corrected seismic velocities suggesting they contain a higher proportion of extrusive volcanic rock. Our interpretation of the Type II SDRs bears a strong similarity to the Tertiary lava flow sequence exposed in Iceland.

- The Type I (fault bounded) SDRs show seismic velocity highs at their down dip ends which we interpret as doleritic sills that were injected into the lava pile. This magma was supplied from feeder systems that exploited the large landward dipping faults that bound the packages and control their geometries.

Our work implies that both the tectonic footwall rotation and flexural-magmatic loading model for SDR generation operate, but at different stages of breakup. The fault-controlled (Type I) SDRs are less voluminous and may be deeply buried by the magmatic-
loading controlled (Type II), which explains why they have been missed in many previous studies of volcanic continental margins.

8 Figure captions

Figure 1. Topographic map of the study area showing seismic line coverage, margin segmentation and spatial extents of the two types of SDR (McDermott et al., 2018). The locations of the two seismic profiles presented here are marked with the bold lines.

Figure 2. A summary of features of SDR types identified from our previous interpretation of reflectivity and magnetic anomaly data. Note the different length scales of the sketches. See McDermott et al., 2018 and Collier et al., 2017 for further details.

Figure 3. PSDM images of the two seismic reflection profiles used in this study. Note that the sections are plotted with the same vertical exaggeration to allow a comparison of dips. The lateral extent of SDR types is shown with the horizontal bars. See Figure 1 for line locations and Figure 2 for a fuller description of SDR type characteristics.

Figure 4. Gather conditioning in the migrated depth domain. The location of the panels is given in Figure 7. The red line shows the picked Top SDR horizon and all panels have the same amplitude scaling applied. Panels (a) – (c) show the effect of varying the P-cut value during Tau-P multiple suppression. A wide P-cut value (e.g. 250 ms) leaves multiples in the record which are much more effectively removed using a P-cut of 150 ms. Panels (d) – (e) show migrated gathers before and after the angle and outer trace mute application.
Figure 5. Comparison of a single common-reflection point (CRP) gather depth-migrated during three sequential tomographic updates to the velocity model. The red and pink lines show the picked Top and Base SDR horizon respectively and all panels have the same amplitude scaling applied. Note how the final tomographic update (d) produced the flattest, most continuous reflections, particularly below a depth of 10 km.

Figure 6. Tomographic inversion results for Line A. In all sections the white dashed line is the computed coherence threshold (i.e. velocity field deeper than this boundary is not constrained). The red and pink lines mark Top and Base SDR as picked from the interpretation of the reflection image alone (Figure 3). (a) Seismic reflection profile underlain by inverted interval velocity field. The depth and lateral resolution of the velocity grid is 200 m and 500 m respectively. (b) Interval velocity field with 0.5 km/s contours. (c) Coherency showing the level of confidence in tomography above the 0.3 threshold. Boxes mark locations of other figures as indicated.

Figure 7. Tomographic inversion results for Line B. Details as Figure 6.

Figure 8. Enlarged section of Line A. (a) Seismic reflection profile. (b) Seismic reflection profile underlain by inverted interval velocity field, the coherency threshold indicating the level of confidence in tomography. (c) Interpretation cartoon. (d) Individual velocity-depth profiles subdivided into Type I (red) and Type IIa (blue) SDRs as shown in the interpretation cartoon. For comparison, the black line is a velocity-depth profile extracted from wide-angle seismic profile BGR04-REFR01 (Becker et al., 2014) at the location of profile ii. (e) Stacked velocity-depth profiles across the Type IIa SDRs for the region shown.
Figure 9. Enlarged section of Line B. Details as Figure 8. Note that the location of profile iii (dashed line) is landward of the seismic section shown in (a-c) but can be seen in Fig.7.

Figure 10. South American SDR interval velocity-pressure profiles obtained in our study compared to those from the ICEMELT experiment in Iceland (grey, Darbyshire et al., 1998).

The conversion from depth to lithospheric pressure for our profiles used a single 3-layer model with densities of water, sediment and SDR-sequence of 1000, 2200 and 2800 kg/m$^3$ respectively (Brocher, 2005). The conversion from depth to lithospheric pressure for the Icelandic profiles used a single 2800 kg/m$^3$ layer, and applied a correction for an estimated 3 km of material removed by glacial erosion (84 MPa, Siler and Karson, 2017).

9 Acknowledgements

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10 References


Goncharov, A. & G. Nelson, 2012. From two way time to depth and pressure for interpretation of seismic velocities offshore: Methodology and examples from the Wallaby Plateau on the West Australian margin. Tectonophysics 572, 26-37. DOI: 10.1016/j.tecto.2012.06.037


Use of low frequencies for sub basalt imaging. Geophysical Prospecting 51, 169-182.
Graphical Abstract (for review)

This study:

Type I SDR
Late-stage sill injection
Continental crust
SDR dip due to footwall rotation

Type II SDR
Magmatic crust
SDR dip due to magmatic loading
Figure 1

Click here to download Figure: Figs_revision_small_all.pdf
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<th>Sub-aerial</th>
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<td>• Not fault-bound, curved towards distal end</td>
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<td>• True-dip; range = 0-14°; median = 4°</td>
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Figure 3
Figure 4

- **a)** Pre-Tau-P
- **b)** P-cut = 250 ms
- **c)** P-cut = 150 ms
- **d)** No mute
- **e)** Angle & outer mute

**Figure 4**

- **Tau-P transform test**
  - Multiples
  - Remnant multiples

- **Trace muting**
  - Stretch

**Legend**
- CRP
- Depth (m)
- Amplitude
Figure 5
Figure 6
Figure 7

Line B

Distance (km)

Depth (km)

Interval Velocity (km/s)

Coherence

Fig. 9

Fig. 4a-c

Fig. 4d-e

V-Z Profile iii

V-Z Profile iv

Coherence threshold

Top SDR

Base SDR
Figure 8
Figure 9
Figure 10