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1 **Laterally Confined Volcanic Successions (LCVS); recording rift-jumps**
2 **during the formation of magma-rich margins**

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10 **Abstract**

11 Seaward Dipping Reflectors (SDRs) are a characteristic feature of magma-rich
12 margins, and represent the generation of large volumes of flood basalts at the point of
13 continental break-up. A number of recent studies provide new insights into the
14 emplacement and tilting of SDRs and conclude that the majority of SDRs are
15 contained within new magmatic crust that has a close affinity to oceanic crust.
16 However, the process by which these initial magmatic systems evolve into a fully
17 established spreading centre remains poorly understood. Several characteristic
18 features of magma-rich margins may be explained by the occurrence of rift-jumps
19 during SDR emplacement, yet the cause and prevalence of such rift-jumps remain
20 unknown.

21 Here we constrain the 3D geometry of the continent-ocean transition in the
22 Orange Basin, offshore South Africa. This allows us to test if, where and why such rift

23 jumps occur. Our results demonstrate an order of along-strike segmentation previously
24 unobserved in these settings. We demonstrate that the SDR belt is disrupted by the
25 occurrence of a volcanic-stratigraphic package, defined as the Laterally Confined
26 Volcanic Succession (LCVS), not previously identified on a rifted margin. We interpret
27 this as a magmatic spreading centre that was abandoned by a subsequent rift-jump.
28 Identification of LCVSs is important for two reasons. First, we argue that the LCVS
29 formed via the same process as SDRs, and hence provides a unique example of SDR
30 geometry prior to their separation onto conjugate plates. Second, as we can map out
31 the 3D geometry of the LCVS and SDRs, we propose that rift-jumps during magma-
32 rich margin formation may be fundamental to the establishment of a laterally
33 continuous incipient spreading centre.

34

35 **Keywords:**

36 Volcanic margin, seaward dipping reflectors, rifting, South Atlantic

37

38 **1. Introduction**

39 Magma-rich margins form when continental breakup is accompanied by the
40 emplacement of large volumes of igneous rock (White and McKenzie, 1989). The
41 addition of magmatic material to the thinned continental crust has made these margins
42 more difficult to study than their magma-poor counterparts, which are defined by
43 detachment faulting, hyperextension and mantle exhumation (Manatschal, 2004;
44 Péron-Pinvidic and Manatschal, 2008). Hence despite the widespread occurrence of

45 magma-rich margins globally, many elements of their magmatic and tectonic evolution
46 remain poorly understood.

47 In seismic reflection profiles, the diagnostic feature of magma-rich margins is
48 the occurrence of thick sequences of seaward dipping reflectors (SDRs) in the upper
49 crust (Hinz, 1981; Mutter et al., 1982; White and McKenzie, 1989). Drilling and
50 geophysical studies (Eldholm and Grue, 1994; Jackson et al., 2000; Larsen and
51 Saunders, 1998; Mutter et al., 1982; Planke and Eldholm, 1994) have consistently
52 shown that SDRs consist predominantly of flood basalts. It has been demonstrated
53 that the innermost SDRs were emplaced on top of attenuated and intruded continental
54 crust (Larsen and Saunders, 1998) whilst the outermost SDRs were emplaced as the
55 uppermost layer of thickened oceanic crust (Larsen and Saunders, 1998; Mutter et al.,
56 1982; Paton et al., 2017). Therefore, the emplacement and tilting of SDRs record how
57 tectonic and magmatic processes interact during the transition from continental
58 thinning to seafloor spreading, and understanding their structural evolution is critical
59 to understanding continental break-up. Several studies have used long-offset long-
60 recording time 2D seismic reflection profiles to investigate SDR formation (e.g. Pindell
61 et al., 2014; Quirk et al., 2014; McDermott et al., 2015; Paton et al., 2017), and these
62 have been supported by numerical modelling (Buck, 2017; Corti et al., 2015) and field
63 studies (Abdelmalak et al., 2015). These studies suggest that SDRs were originally
64 emplaced as sub-horizontal lava flows in a magmatic zone located along the rift axis
65 (e.g. Paton et al., 2017). Given the continued accretion of magmatic material in this
66 zone, the older lava flows are transported away from the rift axis and develop axial
67 dips. These dips form via a combination of magmatic loading and faulting (Buck, 2017;
68 Clerc et al., 2015; Corti et al., 2015; Paton et al., 2017; Quirk et al., 2014).

69 Rift-jumps on fully established mid-ocean ridges occur across a range of scales,
70 from 10's (Hey, 1977) to 100's (Mittelstaedt et al., 2011) km, and in response to local
71 and regional processes. It has recently been suggested that repeated small rift-jumps
72 (10's km) also occur during SDR emplacement (Buck, 2017). Where rift-jumps, in this
73 context, are defined as offsets in the location of the axial magmatic-zone. The
74 occurrence of these events may play an important role in controlling the geometry and
75 asymmetry of conjugate margins (Buck, 2017). However our understanding of the
76 prevalence and driving mechanisms of such rift-jumps is currently limited, as is their
77 impact on subsequent oceanic basin formation. Compounding this uncertainty is that
78 current models only address SDR emplacement in two dimensions.

79 In this study we utilise a closely spaced 2D seismic reflection grid from the
80 Orange Basin, offshore South Africa, to map the continent-ocean boundary in three
81 dimensions. Specifically, we identify a new volcanic-stratigraphic package that
82 resulted from a rift-jump during SDR emplacement. As this new packages records the
83 abandonment of a magmatic spreading centre, it allow us to study the geometry of
84 SDRs prior to their separation onto conjugate plates. The local scale of our study area
85 allows us isolate this abandoned magmatic system and investigate the processes
86 controlling its formation. We demonstrate that during SDR emplacement, rift-jumps of
87 this scale result from interactions between separate magmatic systems that are
88 segmented along-strike. The resulting models have a significant impact on our
89 understanding of magmatic processes at the critical point of continental break-up.

90

91 2. Geological Setting

92 The West African magma-rich margin extends between the Walvis Ridge in the
93 north and the Cape Segment Boundary in the south (Austin and Uchupi, 1982; Franke,
94 2013; Gladczenko et al., 1997; Koopmann et al., 2013; Light et al., 1993). It contains
95 several sedimentary basins, the southernmost and largest being the Orange Basin
96 (Séranne and Anka, 2005). The depocentre (Fig. 1) is located offshore South Africa
97 and Namibia and trends NNW-SSE between 34° and 29°S, having an approximate
98 length of 600 km. To the north it is bounded by the Luderitz Basin, and to the south by
99 the Agulhas-Falkland Fracture Zone.

100 The rift geometry of the Orange Basin is considered to be an interplay between
101 the regional extension associated with Gondwana break-up in the Late Jurassic, which
102 commonly reactivated pre-existing crustal lineaments (Clemson et al., 1997;
103 Gladczenko et al., 1997; Mohammed et al., 2016; Paton et al., 2016), and Atlantic
104 rifting that formed a series of NNW-SSE trending half-grabens (Fig. 2, Light et al.,
105 1993; Clemson et al., 1997; Paton et al., 2016; Mohammed et al., 2016). These half-
106 grabens are controlled by landward-dipping normal faults that may be similar to those
107 observed of rifted margins worldwide (Clerc et al., 2016). The syn-rift stratigraphy is
108 largely undrilled, however, wells within the South African Orange Basin reveal that the
109 half-grabens contain continental sandstones (alluvial and fluvial), lacustrine shales
110 and basic volcanics (Gerrard and Smith, 1982; Jungslager, 1999). Above the syn-rift
111 succession, a wide belt of SDRs is observed (Fig. 2), which is indicative of the
112 transition from continental rifting to seafloor spreading (Fig. 2, Paton et al., 2016).
113 Several wells within the Namibian Orange Basin sampled the landward edge of the
114 SDRs (Wickens and McLachlan, 1990). These showed that the SDRs consist of
115 basalts interbedded with aeolian and fluvial sandstones (Mohammed et al., 2016),

116 indicating subaerial emplacement. The SDRs in the Orange Basin form part of the
117 continuous c. 1600 km long SDR-belt extending northward to the Walvis Ridge
118 (Gladczenko et al., 1997; Koopmann et al., 2013; McDermott et al., 2015). Regionally,
119 a dominantly volcanic lithology of the SDRs is indicated by numerous seismic
120 refraction profiles and potential field studies (Collier et al., 2017; Corner et al., 2002;
121 Hirsch et al., 2009; Koopmann et al., 2014).

122 In the Orange Basin, continental breakup was diachronous and became
123 younger towards the north, as is constrained by the occurrence of the M9-M4 seafloor
124 spreading lineations, all of which occur within the Hauterivian (Collier et al., 2017).
125 These seafloor spreading lineations occur within the unequivocal oceanic crust located
126 seawards of the SDRs (Collier et al., 2017), and hence are not subject to the
127 uncertainties associated with interpreting magnetic lineations in areas of ambiguous
128 crustal type.

129 Following breakup the margin began to subside thermally, and open marine
130 conditions were established during the Barremian-Aptian (Paton et al., 2008).
131 Deposition continues to the present day, with the marine post-rift sediments reaching
132 a current maximum thickness of 5.6 km (Dalton et al., 2016).

133

134 **3. Study area and data**

135 The study area is located in the South African Orange Basin (Fig. 1) and
136 extends for approximately 100 km along strike. The primary 2D seismic reflection
137 profile survey used in this study, which was acquired by Spectrum, was SPOB12 and
138 images the entire continent-ocean transition across the margin. Dip lines trend NE-
139 SW and are typically spaced at 30 km (Fig. 1a). The seismic data were recorded on a

140 10,050 m streamer containing 804 channels with receiver groups spaced at 12.5 m.
141 The source was towed at 8 m depth and a shot interval of 25 m was used. Given the
142 shot-receiver geometry, the maximum common midpoint gather (CMP) fold is 201 with
143 CMPs being located every 6.25 m. Recording time was 10 s with a sample interval of
144 2 ms. The seismic data were pre-stack-time-migrated and pre-stack-depth-migrated,
145 although additionally the raw gathers have been used to estimate seismic velocity.

146 A second seismic survey, K2002, was used to correlate horizons between the
147 widely spaced SPOB12 lines (Fig. 1b). This survey had a recording time of 7 s and
148 does not extend across the entire continent-ocean transition.

149 Seismic profiles are displayed with a blue reflection (positive) marking a
150 downward increase in acoustic impedance, whereas red reflections (negative) mark a
151 downward impedance decrease.

152

153 **4 Seismic Interpretation**

154 *4.1 Methods*

155 The seismic units mapped in the study area have neither been penetrated by
156 boreholes nor extensively mapped in seismic data. Hence crucial to this study was the
157 differentiation of normal oceanic crust and SDRs, and then the identification of several
158 packages within the continent-ocean transition.

159 Across the study area the oceanic crust, SDRs and associated volcanic
160 packages are separated from the post-rift succession by a high-amplitude positive
161 reflector (Fig. 3). Previous studies have referred to this reflector as the break-up
162 unconformity (Franke, 2013; Gerrard and Smith, 1982). We find this term unsuitable
163 as many of the SDRs were likely emplaced as the uppermost part of thickened oceanic

164 crust (Hinz, 1981; McDermott et al., 2015; Mutter et al., 1982; Paton et al., 2017), and
165 hence break-up of the continental lithosphere pre-dates the occurrence of this
166 unconformity. Instead of the 'break-up unconformity' we refer to this reflector as 6At1,
167 which is in accordance with regional sequence stratigraphic models (Muntingh and
168 Brown, 1993).

169 The age of the volcanics underlying 6At1 are uncertain. In comparison to other
170 magmatic margins, e.g. the Vøring Plateau (Mutter et al., 1982) or the SE Greenland
171 margin (Larsen and Saunders, 1998), the volcanics mapped here are considered to
172 have been emplaced across a relatively short period of time (e.g. 2-4 Myrs; Koopmann
173 et al., 2013) prior to the onset of the first normal oceanic crust in the Late Hauterivian
174 (Collier et al., 2017).

175 Within this tectono-stratigraphic framework we apply a seismic characterisation
176 technique that uses changes in seismic facies (Planke et al., 2000) to differentiate
177 between the SDRs and normal oceanic crust. Packages within the Hauterivian
178 volcanics were then defined through mapping and correlating changes in reflector
179 configuration (Mitchum et al., 1977).

180 *4.1.1 Identification of normal oceanic crust*

181 Across the SW of the study area, the crust underlying the post-rift has a
182 characteristic seismic character (Fig. 3, Fig. 4). The base of the post-rift is located at
183 depths of 6.5-6.9 s TWTT and is defined by a high amplitude positive reflector which
184 is underlain by 0.2-0.3 s TWTT of chaotic reflectors with moderate to low amplitudes.
185 Beneath this layer is a zone of seismic transparency with a TWTT thickness of 0.6-0.8
186 s TWTT (Fig. 3a). The lower crust contains high amplitude reflectors with very variable
187 dip directions, varying from landward-dipping, to near-horizontal, to seaward-dipping

188 (Fig. 3a). This zone of reflectivity is underlain by a band of near-horizontal high-
189 amplitude reflectors (Fig. 3a).

190 This seismic character is consistent with examples of high-magma-supply
191 oceanic crust worldwide (Bécel et al., 2015; Ranero et al., 1997; Reston et al., 1999)
192 and is interpreted as such. In these settings oceanic crust typically conforms to the
193 Penrose model of basalts, dolerite dykes and lower crustal gabbros (Penrose field
194 conference on ophiolites [Geotimes, v. 17, p. 24–25]).

195 The high amplitude top-basement reflection in Fig. 3 and Fig. 4 represents the
196 basalt-sediment interface and the 0.2-0.3 s TWTT thick zone of chaotic reflections
197 underlying it are likely to correspond to the reflectivity within pillow basalts and sheet
198 flows. The zone of seismic transparency lying directly beneath this represents sheeted
199 dykes and the uppermost gabbros (Fig. 3b). In the lower crust, the high amplitude
200 reflectors observed here are similar to the intra-gabbroic reflectors (Fig. 3b) imaged in
201 oceanic crust worldwide (Bécel et al., 2015; Ranero et al., 1997; Reston et al., 1999).
202 In seismic reflection data the oceanic Moho is often imaged as a diffuse zone of high
203 amplitude reflectivity (Mutter and Carton, 2013), which is how we interpret the band of
204 near horizontal reflectors located at c. 9.3 s TWTT (Fig. 3b).

205 *4.1.2 Identification and mapping of SDRs*

206 SDRs were identified by their characteristic geometry (Mutter et al., 1982),
207 whereby, all reflectors dip seawards, have a convex-up geometry, and diverge
208 downdip into a zone of chaotic seismic imaging (e.g. Fig. 3). Mapping the width of the
209 SDR belt is commonly used to constrain the variations in the amount of volcanics
210 emplaced along-strike in this region (Koopmann et al., 2014, 2013). In this study, the
211 width of the SDR belt has been calculated by measuring the horizontal distance from

212 the downdip termination of the first (oldest, and landward-most) to the last (youngest,
213 and seaward-most) SDR (Fig. 3).

214 *4.2 Basin architecture*

215 Using the classifications outlined above, SDRs and oceanic crust were mapped
216 across the study area, and variations from these two seismic units were identified and
217 described.

218 Oceanic crust is located across the SW of the study area (Fig. 3, Fig. 4),
219 landwards, this oceanic crust transitions into a belt of SDRs, within which all reflectors
220 dip seawards with a downdip divergence (Fig. 3, Fig. 4). Pre-stack depth migrated
221 profiles, which show the depth-converted geometry of the SDRs, indicate that
222 individual reflectors have convex-up with dips ranging from 5-20° (Fig. 3c). In addition
223 to their diagnostic geometry, the SDRs have a distinctive seismic character comprising
224 strong amplitude variations along individual SDRs (Fig. 3a). Occasional high amplitude
225 saucer shaped reflectors are also observed (Fig. 3b). These are broadly concordant
226 with the SDRs with discordance occurring near the saucer tips and correspond to
227 localised sill intrusions (e.g. Planke et al., 2005).

228 Underlying the SDRs are densely spaced landward-dipping reflectors which
229 have variable amplitudes (Figs 3a and 4b). Where these two sets of reflectors
230 intersect, the landward dipping reflectors cross-cut the SDRs, although no intra-SDR
231 offsets are observed (Fig. 4b). We interpret these landward-dipping reflectors as dykes
232 which were emplaced at the same time as the SDRs. This interpretation is supported
233 by the lack of offsets observed where the landward-dipping reflectors cross-cut the
234 SDRs, which indicates that they are not faults. The seismic character of this set of
235 reflectors is also similar to that of dyke swarms imaged in the North Sea basin (Phillips
236 et al., 2018) and the Vøring margin (Abdelmalak et al., 2015).

237 In the south of the study area the SDR belt is very wide (e.g. 88 km in Fig. 1b,
238 Fig. 1b), and consists of a continuous succession in which no major unconformities
239 are present. 30 km landward of the SDR/oceanic crust interface the quality of the
240 SDRs imaging is reduced by an overlying volcanic complex consisting of mounded,
241 rugose high amplitude reflectors (these are labelled as 'post-SDR volcanics' in Fig.
242 3b). Disrupted SDRs can be identified beneath this volcanic complex, hence it was
243 either emplaced during the emplacement of the latest SDRs (i.e. during the
244 emplacement of those SDRs located seawards of the mound) or during the earliest
245 post-rift. The occurrence of this structure is limited to this seismic line.

246 The >80 km SDR belt width in the south (Fig. 3) is in noticeable contrast to a
247 section 30 km towards the north-west (Fig. 4) in which the width is 26 km. This,
248 therefore, corresponds to a significant narrowing of the SDR belt by ~ 54 km over 30
249 km along-strike (Fig. 1b).

250 This substantial narrowing of the SDR belt towards the north-west coincides
251 with the presence of a previously unrecognised seismic package (orange package in
252 Fig. 4b) that only occurs in the north of the study area and underlies the SDRs. This
253 additional package is thickest in a central axial zone (labelled Axis 1.1), reaching a
254 thickness of c. 0.85 s TWTT (c. 2400 m thick based on the interval velocities shown in
255 Supp. Fig. 2b), and thins in a near symmetrical geometry both landwards and
256 seawards (Fig. 4b). A fanning of dips is observed within the package and stratal
257 thinning is accommodated by onlap towards the mini-basin flanks. The base of the
258 package is a high amplitude onlap surface, whilst its top is defined as the reflector
259 above which there is no prominent thickening into the axial trough. Slightly east of the
260 central axis, the package is cross-cut by a near vertical zone of disruption; reflectors
261 can be correlated across this zone without any offset and hence it cannot be

262 interpreted as a major fault. The width of the package has been plotted onto the margin
263 domain map in Fig. 1b, and its occurrence directly correlates with the northward
264 narrowing of the SDR wedge. Given the geometry of this package we name it the
265 Laterally Confined Volcanic Succession, or LCVS. Its volcanic lithology has been
266 validated through seismic velocity analysis (see Supplementary Material).

267 Directly underlying the LCVS are two wedges of reflectors that both thicken
268 downdip into an axial zone of chaotic imaging (Wedges 1 and 2 in Fig. 4b). Wedge 1
269 contains dominantly landward dipping reflectors, whilst Wedge 2 consists of reflectors
270 which all dip seawards and are identical to SDRs. Wedge 1 is less well imaged than
271 its seaward dipping counterpart, which is likely to be a result of the locally thicker
272 overburden.

273 Our observations demonstrate that the SDR belt narrows northwards across
274 the study area, and that this narrowing is accompanied by the occurrence of a new
275 package, the LCVS. In order to understand the lithology, evolution, and significance
276 of the LCVS we now consider the temporal and spatial relationship between the SDRs
277 and the LCVS.

278 *4.3 Relationship between the LCVS and the SDRs*

279 As noted previously, the SDR belt narrows from 88 km to 26 km across the
280 relatively short distance of 30 km (Fig. 1b), and this narrowing is accompanied by the
281 occurrence of the LCVS. In order to investigate the temporal and spatial relationship
282 between the LCVS and the SDRs, three reflectors (R1, R2 and R3), have been
283 correlated across the dataset (these reflectors are shown in figures 3b and 4b). This
284 allows us to identify three distinct stage of margin evolution: *stage a* which is bounded
285 by R1 and R2; *stage b* which is bounded by R2 and R3; and *stage c* which consists of
286 the SDRs overlying R3. In this section we summarise the observations for each of

287 these stages, and in a subsequent section (Section 5) we propose a model for their
288 evolution.

289 *4.3.1 Stage a (R1-R2)*

290 The package bounded by R1 and R2 allow us to investigate the along-strike
291 relationship between the LCVS and SDRs (Fig. 5). In the north of the study area, R1
292 and R2 correspond to the top and the base of the LCVS respectively (seismic line A,
293 Fig. 5). The R1-R2 isochron (Fig. 6a) demonstrates that the LCVS trends NNW-SSE,
294 is c. 50 km long and up to 30 km wide. The thickness of the LCVS is greatest within a
295 NNW-SSE trending axial zone (Axis 1.1 in Fig. 6b), although this also shows along-
296 strike thickness variations. Seismic line A (Fig. 5) is located through the thickest part
297 of the LCVS (Fig. 6a), with vertical thickness reaching 0.85 s TWTT in the axial zone.
298 Away from this axial zone the landward and seaward stratal thinning is near
299 symmetrical.

300 Seismic line B (Fig. 5) is located some 15 km SE of seismic line A, and is
301 positioned over the southernmost flank of the LCVS (Fig. 6a). Here the thickness of
302 the axial zone reaches 0.32 s TWTT (c. 0.9 km; Supp. Fig. 2b) and, while stratal
303 thinning does occur both landwards and seawards of this zone, it is not as pronounced
304 as in line A. Located in the south of the study area, seismic line C (Fig. 5) shows that
305 reflections R1 and R2 bound a package of SDRs and there is no LCVS present. Close
306 to their down-dip termination this package reaches a maximum thickness of 0.54 s
307 TWTT (c. 1.5 km, Fig. 5). The R1-R2 isochron map (Fig. 6a) illustrates that this lateral
308 transition, from LCVS to SDRs, occurs through the progressive diminishment of the
309 former and the development of the latter. No seismically resolvable faults are present
310 in this transition zone. Therefore, it is through this gradual lateral transition that the
311 SDR belt shows southward widening. Also evident from the isochron (Fig. 6a) is that

312 the downdip termination of R1-R2 bounded SDRs, named Axis 1.2 (Fig. 6a, and Fig.
313 5), is offset ≥ 10 km SW from the LCVS axial zone (Axis 1.1 in Fig. 6a).

314 Mapping of the package bounded by R1-R2 has demonstrated that during *stage*
315 *a*, the LCVS was developing in the north of the study area, whilst SDRs were
316 developing in the south.

317 4.3.2 Stage b (R2-R3)

318 It has been demonstrated that the nature of R2 varies across the study area –
319 in the north it defines the top of the LCVS (lines a and b in Fig. 5), whilst in the south
320 it is an intra-SDR belt reflector (line c in Fig. 5). The package bounded by reflectors
321 R2 and R3 also varies structurally across the area. In the south (Fig. 3b), this package
322 consists of a wedge of SDRs which conformably overlie the *stage a* SDRs. Meanwhile,
323 in the north of the study area this package is reduced in thickness and differs
324 geometrically from SDRs (Fig. 4b). Whereas SDRs show continuous downdip
325 divergence, here thickness variations are limited to the updip pinchout of the package
326 and downdip from this area stratigraphic thickness appears relatively constant (Fig.
327 4b).

328 4.3.3 Stage c (post-R3 SDRs)

329 *Stage c* is characterised by the occurrence of SDRs across the study
330 area. These SDRs form a continuous succession from R3 to the transitional boundary
331 between the SDRs and normal oceanic crust (Fig 3b, Fig. 4b). An isochron of the *stage*
332 *c* SDRs shows thickness variations across the study area (Fig. 6b). The top of the
333 isochron is defined as 6At1 (Fig. 3a, Fig. 4a) and the base is defined as composite
334 horizon consisting of R3 and the SDR downdip terminations (this composite horizon
335 is shown in figures 3a and 4a). The isochron (Fig. 6b) demonstrates that continuous

336 SW thickening occurs from the updip pinchout of the succession to the first downdip
337 SDR termination, which is labelled as Axis 2 on the isochron (Fig. 6b) and the
338 corresponding seismic sections (Fig.4b, Fig. 3b). Axis 2 trends NNW-SSE across the
339 study area (Fig. 6b), and although several kinks are observed we interpret this as a
340 continuous axis (as opposed to the two axes being present in Stage 1). This
341 interpretation is based on *stage c* being defined by the development of SDRs across
342 the study area, whilst *stage a* was defined by the development of SDRs and the LCVS.
343 Immediately SW of Axis 2, thickness variations are less pronounced until the SDR belt
344 begins thinning SW towards the location of the first 'normal' oceanic crust (labelled in
345 Fig. 6b).

346 *4.3.4 Comparison between features from stages a to c*

347 Axes 1.1 and 1.2 were defined in relation to *stage a*, whilst Axis 2 was defined
348 in relation to the *stage c*. As will be shown in the discussion, these axes can be used
349 to infer the 3D evolution of the SDRs and the LCVS, as such the spatial relationships
350 between them is important. All of the axes trend approximately NNW-SSE and are
351 margin-parallel, however the offsets between the *stage a* and *stage c* axes vary. In the
352 north, where the LCVS is present, there is an offset of ≤ 25 km between Axis 1.1 and
353 Axis 2 (Fig 6b, Fig 4b). Meanwhile in the south of the study area, where axes 1.2 and
354 2 are separated by a continuous succession of SDRs, there is an offset of c. 12 km
355 (Fig. 3b, Fig. 6b).

356 **5. Discussion**

357 *5.1 Formation of the Laterally Confined Volcanic Sequence*

358 During emplacement, the lava flows that became SDRs were emplaced above
359 an axial magma-source (Larsen and Saunders, 1998; Mutter et al., 1982; Paton et al.,

360 2017; Quirk et al., 2014). Continued plate separation resulted in the accretion of more
361 magmatic material in this axial zone which forced older lava flows to be transported
362 away from the axis and to be progressively rotated to their present dips. Above the
363 axial magma-chamber the SDRs were fed by intrusions, and this SDR/intrusive
364 contact is preserved today at their down-dip terminations (Paton et al., 2017). In
365 seismic reflection data, these down-dip terminations are defined by rapid dimming into
366 a zone of chaotic seismic imaging where occasional landward dipping reflectors are
367 present. In terms of seismic character, this is similar to the axial zone that separates
368 the landward and seaward diverging wedges that are observed beneath the LCVS
369 (wedges 1 and 2 in Fig. 4b). Based upon the similarities between geometry, seismic
370 character (Fig. 4b) and bulk lithology (Supp. Fig. 1; Supp. Fig. 2), we interpret the
371 package underlying the LCVS as being formed by the same processes as SDRs. In
372 other words, these two wedges of volcanics were fed from an axial magmatic
373 spreading centre (Fig. 7a, i), which today is preserved as the axial zone of chaotic
374 imaging (Fig. 4b).

375 The LCVS is unfaulted and onlaps the underlying volcanic package (Fig. 4B),
376 and the seismic velocities indicate that it also comprises mainly volcanics (Supp. Fig.
377 1; Supp. Fig. 2). The axial zone, in which this package is thickest, overlies the magma
378 source that fed the underlying volcanics (Axis 1.1 in Fig. 4b). This observation, in
379 addition to axial disruptions within the LCVS, strongly suggests that it was also
380 sourced from this underlying magma-source (Fig. 7a, i). As mentioned previously, the
381 volcanics directly underlying the LCVS were emplaced via the same mechanism as
382 SDRs. Meanwhile, the reflectors *within* the LCVS onlap against these underlying
383 volcanics, and are often confined to a mini-basin. The structural confinement indicates
384 that extrusion rate was now less than the subsidence rate, with this subsidence being

385 driven by either waning magma-supply (Mutter et al., 1982) or flexure driven by
386 underlying intrusives (Buck, 2017; Corti et al., 2015). The unfaulted nature of the LCVS
387 is important as it provides insights into the processes controlling subsidence on
388 magma-rich margins. The relative roles of faulting (Clerc et al., 2015; Geoffroy et al.,
389 2015; Quirk et al., 2014) and magmatic loading (Buck, 2017; Corti et al., 2015; Paton
390 et al., 2017) in generating the dips recorded by SDRs are disputed. The LCVS provides
391 an unambiguous example of subsidence being generated without normal faulting.
392 Hence the architecture of this package supports models of SDR formation in which
393 subsidence is generated via the magmatic loading of thin and weak crust (Buck, 2017;
394 Corti et al., 2015; Paton et al., 2017).

395 The LCVS is overlain by SDRs (Fig. 4b), the emplacement of which indicates
396 the eventual abandonment of the magmatic zone feeding the LCVS and the
397 establishment of a new magmatic zone ≤ 25 km to the SW (Fig. 7b, i). This new
398 magmatic zone remained active throughout the remainder of the breakup process,
399 resulting in an SDR belt with a width of 26 km. Continental breakup results in a wedge
400 of SDRs being preserved on each conjugate margin. Hence the wedge of SDRs
401 conjugate to that shown in Fig. 4B is located offshore Argentina (Franke et al., 2007;
402 Koopmann et al., 2014; Paton et al., 2017). The width of the Argentinian SDR belt
403 varies along-strike and there are examples where the SDRs have similar dimensions
404 to those located seawards of the LCVS (Franke et al., 2007), however further work
405 would be needed to validate this.

406 Our model implies that the LCVS and the underlying package formed via the
407 same mechanism as SDRs, however, the subsequent jump in the location of
408 magmatism resulted in this early magmatic spreading system being preserved in its
409 entirety (Fig. 7a, i). The majority of previous studies have implied that flood basalts

410 erupted during the rift-drift transition are sourced from a single axial magma-system,
411 which results in wide belts of SDRs (Mutter et al., 1982; Paton et al., 2017; Quirk et
412 al., 2014). However here we demonstrate that this magma-source can jump in space
413 and time. Whilst the possibility of 'rift-jumps' disrupting the SDR belt has been
414 suggested by previous authors (Buck, 2017; Pindell et al., 2014), this is the first
415 unambiguous example of one such jump in seismic reflection data.

416 *5.2 What drives rift-jumps during SDR formation?*

417 It has recently been suggested that a series of rift-jumps can occur during this
418 SDR emplacement (Buck, 2017). This model is supported by the presence of
419 unconformities within SDR belts, indicating episodic emplacement (Franke et al.,
420 2007; Koopmann et al., 2013). This is in contrast to previous studies, which suggest
421 that SDR emplacement is, to a large extent, a continuous process (Mutter et al., 1982;
422 Paton et al., 2017; Quirk et al., 2014).

423 We observe that both models are applicable to different parts of the study area
424 - in the south of the area the SDR belt is 88 km wide (Fig. 3) and its structural uniformity
425 indicates that the SDRs were emplaced above, and transported away from, an axial
426 magma source that remained spatially fixed (Fig. 7a, ii, Fig. 7b, ii). Meanwhile, in the
427 north of the study area (Fig. 4) the magma-source did not remain fixed but jumped,
428 leading to the preservation of the LCVS (Fig. 7a, i; Fig. 7b, i).

429 From these results it is evident that rift-jumps of the scale recorded here are
430 local, not regional, features. Through assessing the 3D evolution of the study area we
431 can assess what is driving a rift-jump in the north, and continuous SDR emplacement
432 in the south. It is suggested that the rift-jump observed has fundamental implications
433 for SDR emplacement.

434 A three-stage model, stage a–c, is presented to demonstrate the 3D tectono-
435 magmatic evolution of the study area (Fig. 7). During *stage a*, the LCVS and underlying
436 package were emplaced contemporaneously with SDRs in the south (Fig. 5). The axial
437 zone, Axis 1.1 (Fig. 6a), of the LCVS directly overlies the magma-chamber that fed
438 this and the underlying package. The contemporary SDRs being emplaced to the
439 south were fed from a magma-chamber located beneath their present-day down-dip
440 terminations (Axis 1.2 in Fig. 6a). Therefore, at this stage of the breakup process,
441 magmatism was partitioned along-strike into two segments, which were laterally offset
442 by c. 15 km (Fig. 7a, iii). Although one such feature is not imaged in these seismic
443 reflection data, it is possible that two these segments were separated by a transfer
444 zone with dextral offset (Fig. 7a, iii).

445 During *stage b*, SDR emplacement continued in the south of the study area, as
446 is evident from a thick wedge of volcanics developing during this time (Fig. 3b). These
447 SDRs conformably overlie those emplaced during *stage a*, indicating that locally, SDR
448 emplacement was continuous from one stage to the next (Fig. 7b, ii). The north of the
449 study area, meanwhile, is characterised by relatively little active magmatism. The
450 magmatic zone feeding the LCVS was abandoned, and a package with relatively
451 constant thickness was emplaced (Fig. 7b, i) conformably on top of the LCVS. This
452 package is interpreted as volcanics that were passively infilling remnant topography
453 (Fig. 7b, i).

454 *Stage c* is defined by SDR emplacement across the study area, as is shown by
455 SDRs being present in seismic lines both in the north (Fig. 4b) and the south (Fig. 3b).
456 These SDRs are relatively uniform in terms of along-strike thickness and width (Fig.
457 6b). In section 4, Axis 2 was defined as the downdip termination of the oldest SDR
458 within this package (i.e. the base of the package), this axis represents the trend of the

459 magmatic zone which was established during this stage. Axis 2 trends NNW-SSE and
460 only small kinks in this trend are present. This, in addition to the along-strike uniformity
461 of this package, indicates that the *stage c* SDRs were emplaced from a laterally
462 continuous axial zone that extended across the study area(Fig. 7c, iii).

463 Our model suggests that from *stage a* to *stage c* there is a transition from a
464 magmatic system that was segmented along-strike (Fig. 7a, iii), to one that was
465 relatively continuous along-strike (Fig. 7c, iii). By comparing the magmatic axes
466 defined for each of these stages, we can assess how this transition occurs. In the
467 south of the study area, there is an offset of 12 km (Fig. 3b) between the *stage a* (Axis
468 1.2) and *stage c* (Axis 2) axes (Fig. 6b). As was discussed above, in this part of the
469 study area, SDR emplacement appears to have been relatively continuous through
470 time. Therefore the offset between axes 1.2 and 2 is explained by 12 km of SDR
471 emplacement occurring between *stages a* and *c* (Fig. 7b, ii). However, in the north of
472 the study area, the offset between the *stage a* (Axis 1.1) and *stage c* (Axis 2) axes
473 cannot be explained by this mechanism. Rather, this offset results from a rift-jump of
474 25 km (Fig. 7c, i), resulting in the present-day preservation of the LCVS (Section 5.1).
475 Therefore it is clear that this rift-jump records this transition from a laterally segmented,
476 to an axially aligned magmatic system. The specific cause of the rift-jump can now be
477 investigated.

478 It is notable that the laterally continuous magmatic system of *stage c* occurs
479 along the same trend as that present in the south of the study area during both *stages*
480 *a* and *b* (see maps in Fig. 7). This may imply that the southern magmatic system of
481 *stage a* (axis 1.2) was propagating northwards (maps in Fig. 7). If this was the case
482 then the localisation of strain on the propagating magmatic zone may have resulted in
483 the abandonment of the LCVS. In two-dimensions, this process would be recorded by

484 the rift-jump evident in seismic sections through the north of the study area. This
485 mechanism for LCVS formation is consistent with observations from overlapping
486 seafloor spreading centres – where one spreading segment can be ‘decapitated’ by
487 the propagation of the other (Macdonald et al., 1998). The northward propagation of
488 individual magmatic segments is also consistent with the regional opening direction of
489 the South Atlantic (Jackson et al., 2000), although such northward opening has been
490 previously unrecognised at this scale. We suggest that *stage b*, where active
491 magmatism was occurring in the south but not in the north, marks the stage where
492 northward propagation of the southward segment coincided with the abandonment of
493 that feeding the LCVS.

494 This process provides a mechanism for the rift-jumps of similar magnitude that
495 have recently been modelled as occurring during SDR emplacement (Buck, 2017).
496 These models use rift-jumps to explain the 2D structure of magma-rich margins,
497 however our results show that these events may result from 3D interactions between
498 separate magmatic segments. Whilst our results suggest that these processes act on
499 a local scale, they will necessarily affect the interpretation of isolated seismic reflection
500 lines.

501 *5.3 Rift-jumps and margin segmentation*

502 The Orange Basin is divided along strike into a series of segments separated
503 by transfer zones with a spacing of >100 km (Clemson et al., 1997; Koopmann et al.,
504 2013). These transfer zones can be either strike-slip faults or basement highs
505 (Clemson et al., 1997). The data shown in this study are located between two such
506 segment boundaries, the Namaqua Segment Boundary to the north, and the Cape
507 Segment Boundary to the south (Fig. 1, Clemson et al., 1997; Koopmann et al., 2013).
508 We have demonstrated that during the emplacement of the LCVS the basin was

509 segmented along strike into a series of laterally offset magmatic centres (*stage a*, Fig.
510 7). These magmatic centres were laterally offset 15 km from one another, and it is
511 possible that a transfer zone (Fig. 7, *stage c*) separated them. This segmentation
512 occurs between the regionally spaced segment boundaries described in previous
513 studies (Clemson et al., 1997; Koopmann et al., 2013), and hence represents an order
514 of segmentation that has not been previously recognised on magma-rich margins.
515 Furthermore, in contrast to the regional segmentation previously described, this
516 smaller-scale segmentation was not present throughout the entire breakup process: it
517 was present during the emplacement of the LCVS (*stage a*, Fig. 7), but was erased by
518 a subsequent rift jump (*stage c*, Fig. 7). We have suggested that the mechanism
519 driving this rift-jump was the northward propagation (section 5.2) of an adjacent
520 magmatic segment. However, regardless of the mechanism, it is notable that the
521 system evolved from being segmented along-strike (*stage a*) to relatively continuous
522 along-strike (*stage c*). We suggest that rift-jumps such as that interpreted here, may
523 occur to accommodate the formation of a laterally continuous magmatic spreading
524 zone. As such, this model may be applicable to the generic process of SDR
525 emplacement. Specifically, it offers a mechanism for the occurrence of the discrete
526 wedges of SDRs observed on 2D seismic reflection profiles across the South Atlantic
527 magma-rich margins (Franke et al., 2010; Koopmann et al., 2013). These wedges are
528 separated by unconformities, and may record a series of small (<10 km) rift-jumps as
529 the system becomes progressively less structurally segmented along-strike.

530

531 **6. Conclusions**

532 SDRs have become diagnostic features of magma-rich margins globally. In this
533 study we identify a new volcanic package, the Laterally Confined Volcanic Succession
534 (LCVS), which is also present in these settings. The LCVS was produced by a rift-
535 jump during SDR emplacement, and provides an example of SDR-geometry prior to
536 plate separation.

537 Previous models have suggested that rift-jumps may be common during the
538 formation of SDRs, however the driving mechanism remained uncertain. The 3D
539 evolution of the magmatic system in the Orange Basin indicates that such rift-jumps
540 occur to allow strain localisation on a continuous axial magmatic zone. This results in
541 the abandonment of the structural segmentation present earlier in the breakup
542 process. We suggest that this process may be fundamental in the development of an
543 incipient spreading centre, and that such magmatic-driven rift-jumps should be
544 observable on other magma-rich margins.

545

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552

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726

727 **Figure Captions**

728 **Fig. 1.** a) Regional map showing the location of the Orange Basin, outlined in a red
729 dashed line, in relation to Namibia and the Republic of South Africa (RSA). Seismic
730 lines are also shown, with SPOB12 in black and K2002 in orange. The segment
731 boundaries of Koopmann et al., 2013 are shown in as brown dashed lines. In a) the
732 study area is shown a red box. b) Sketch map of the study area, showing the location
733 of seismic lines and the width of the structural remains described in the text. Black
734 dashed line show bathymetry (m below sea level).

735 **Fig. 2.** Representative cross-section through the Orange Basin, line location is shown
736 in Fig. 1a. Figure demonstrates the structure of the syn-rift, SDRs and post-rift.
737 Modified from Mohammed et al., 2016.

738 **Fig. 3.** a) PSTM seismic reflection line with a vertical exaggeration of 5:1, the line
739 location is shown in *Fig. 1b*. The black dashed line demonstrated the base of the
740 isochore shown in Fig. 6b. b) Seismic line shown in a) with interpretation, four
741 packages of SDRs are distinguished (shown in red, dark red, orange and light brown)
742 and are separated by reflectors R1, R2, and R3. The oceanic crust is shown light blue,
743 and the syn-rift in dark blue. c) Line drawing of structure from the pre-stack depth
744 migrated (PSDM) seismic line, this section has no vertical exaggeration.

745 **Fig. 4.** a) PSTM seismic reflection line with a vertical exaggeration of 5:1, the line
746 location is shown in Fig. 1b. The black dashed line demonstrated the base of the
747 isochore shown in Fig. 6b. b) Seismic line shown in a) with interpretation: the oceanic
748 crust is shown in light blue, and the syn-rift in dark blue. In the area containing the
749 SDRs and the LCVS four packages are distinguished and these are separated by
750 reflectors R1, R2 and R3. c) Line drawing of structure from the pre-stack depth
751 migrated (PSDM) seismic line, this section has no vertical exaggeration.

752 **Fig. 5.** Three seismic lines, a-c, showing the along-strike transition from the Laterally
753 Confined Volcanic Succession to SDRs. The line locations are shown on the isochrons
754 in Fig. 6. Reflectors R1 and R2 are annotated in each line, as are the stratal geometries
755 in the package that they bound.

756 **Fig. 6.** a) vertical TWTT thickness map (an isochron) of the package shown in Fig. 5,
757 which is bounded by reflectors R1 and R2 (*Stage a*). The three seismic lines (a-c)
758 shown in Fig. 5 are also annotated and labelled here. b) isochron of the *Stage c*
759 package which is defined as the SDRs overlying reflector R3. The base of this isochron
760 is illustrated in both figures 3a and 4a.

761 **Fig. 7.** Three stage conceptual model for the 3D development of the LCVS and SDRs
762 across the study area. a) shows basin evolution during *stage a*, as described in the
763 text. b) shows basin evolution during *stage b*. c) shows basin evolution during *stage*
764 *c*. For each stage the colours relate to those shown in earlier figures (Fig. 3, Fig. 4,
765 Fig. 5). Each stage of margin evolution is represented by two representative cross-
766 section reconstructions (i and ii), and one sketch map (iii). Cross section i)
767 demonstrates the evolution of the seismic reflection line shown in Fig. 4 which is
768 characterised by the occurrence of a rift-jump and the preservation of the LCVS.

769 Cross-section ii) shows the evolution of the seismic line shown in Fig. 3, which was
770 characterised by continuous SDR emplacement through time.

771

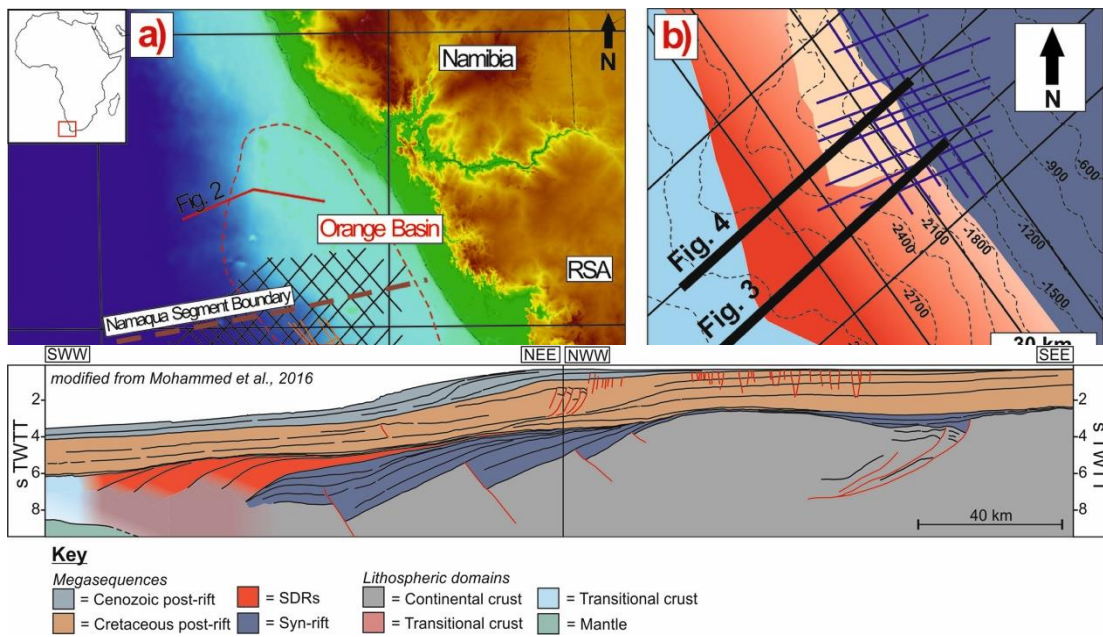


Figure 1 (colour)

Figure 2 (colour)

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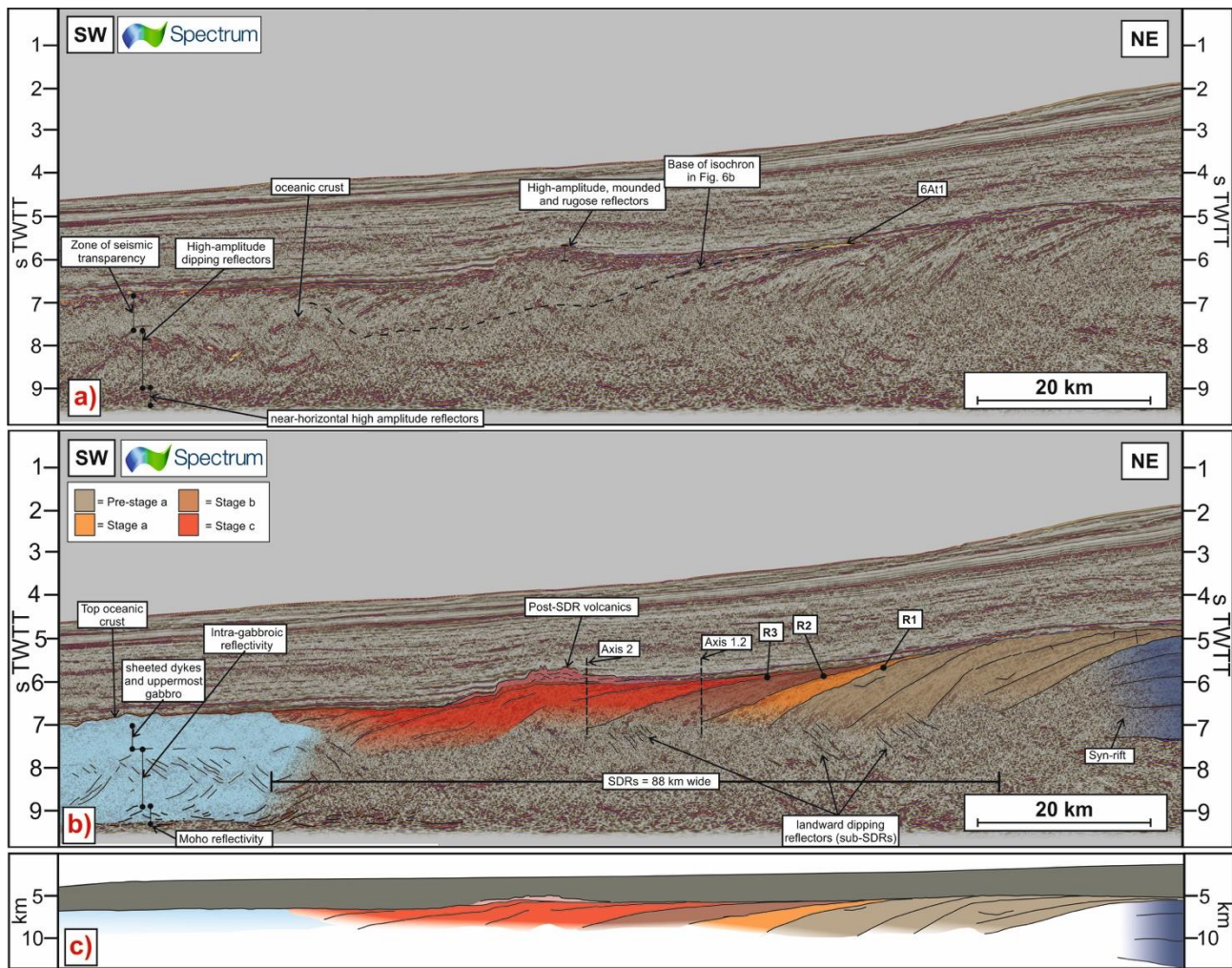


Figure 3 (colour)

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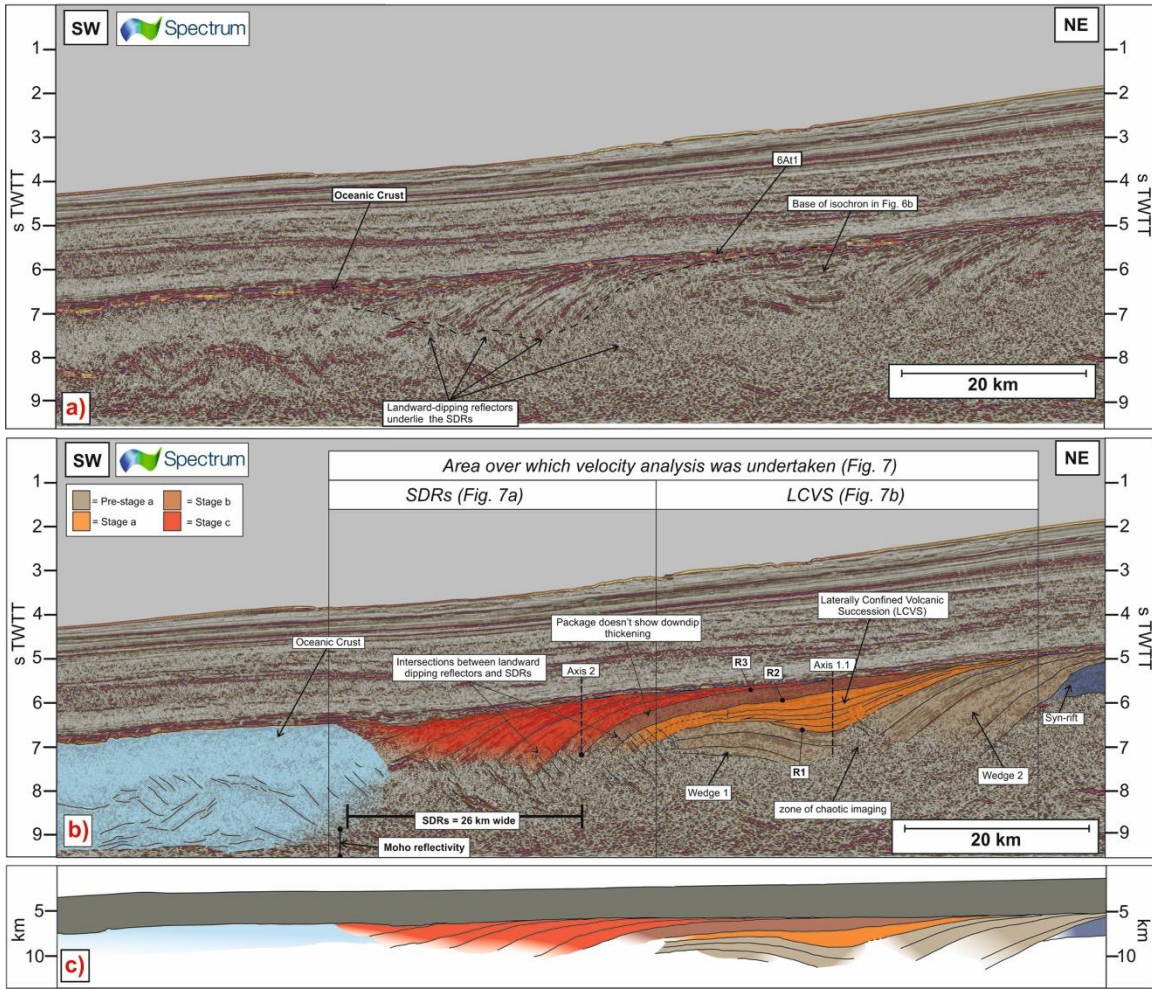


Figure 4 (colour)

794

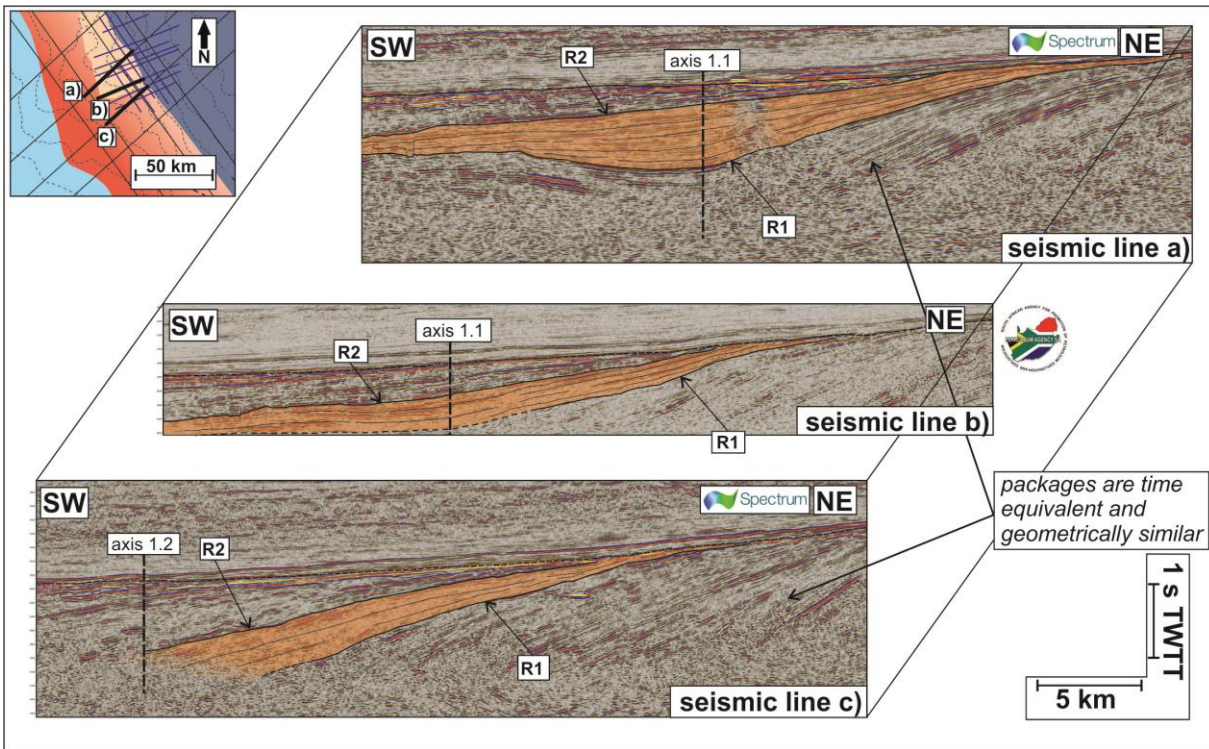


Figure 5 (colour)
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796

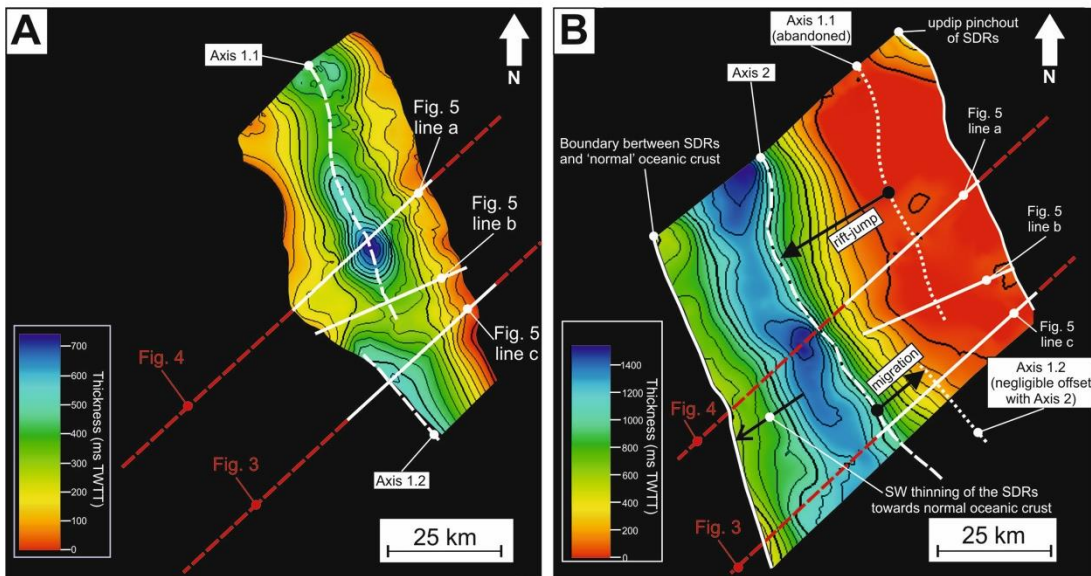


Figure 6 (colour)

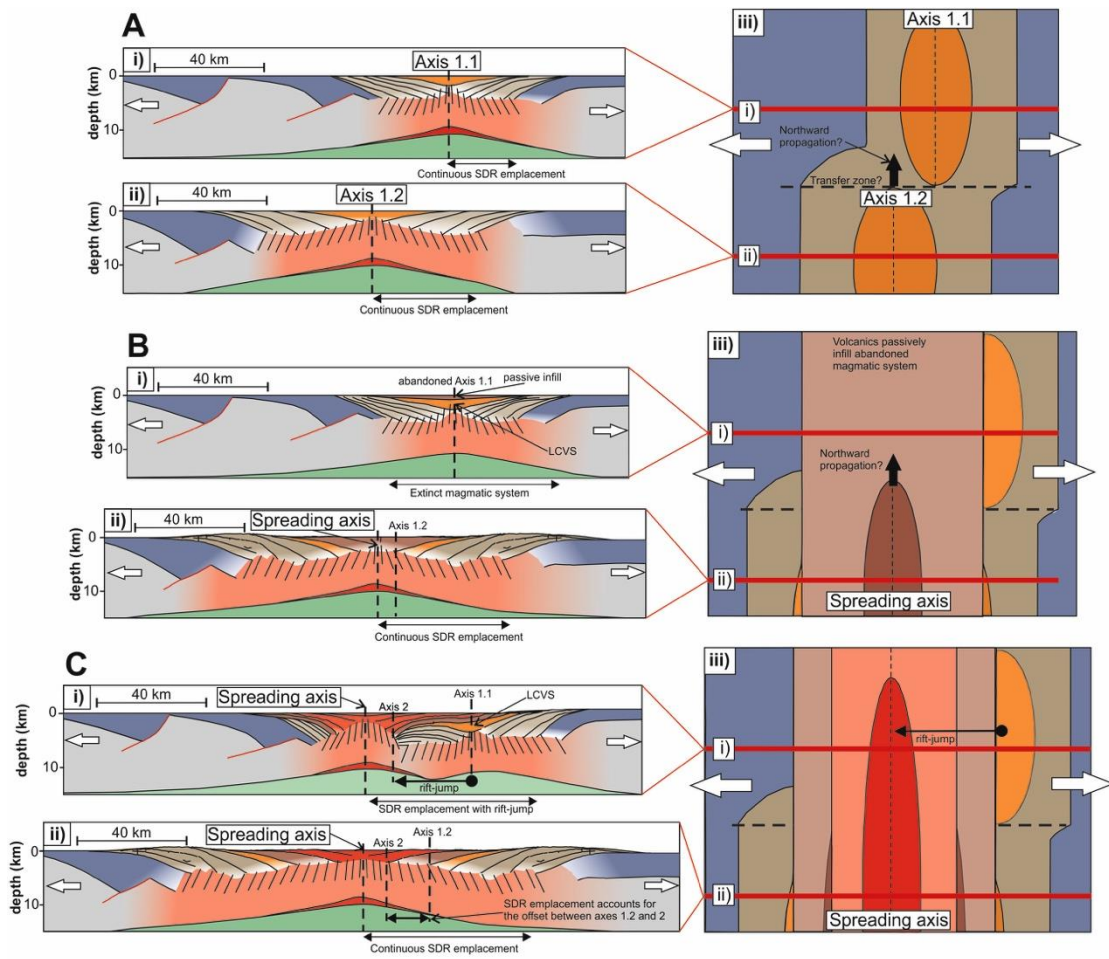


Figure 7 (colour)