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1 Conjugate volcanic passive margins in the austral segment of the South Atlantic – architecture 2 and development

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13 Abstract

14 The thorough interpretation of the South Atlantic long offset seismic lines provides a new
15 view of the magmatic and crustal architecture and of the evolution of the conjugate volcanic passive
16 margins (VPM) along the austral segment of the South Atlantic Ocean. In this study, we make the
17 synthesis of previously published and unpublished seismic profiles that are reinterpreted
18 consistently, and, for the first time, displayed in a conjugate view. The seaward dipping reflector
19 sequences (SDR) expose here three main structural types (inner, outer and intermediate) that can be
20 distinguished based on both the curvature of the reflectors and the topography displayed at their
21 downdip terminations. We observe a specific correlation between the thinning profile of the crust
22 and the geometric attributes of SDR. This correlation suggests a genetic relationship that is
23 discussed in-term of rheological evolution of the crust during the magma-assisted lithospheric
24 break-up. The South Atlantic margins architecture also featured a moderate but significant
25 asymmetry, with a relative difference estimated to about 60% for the margin width, the total volume
26 of SDR and their spatial distribution. A change toward a more symmetric architecture is however
27 observed when approaching the highly magmatic area of conjugate Walvis Ridge and Rio Grande
28 Rise. The influence of the pre-rift inheritance, the mode of rifting and the coeval mantle dynamic
29 are put into balance to explain this global architecture. In the asymmetric domain, the extrusive
30 sequences show evidences for westward rift jumps that participate to develop a wider margin on the
31 African side. We propose that an asymmetric thermal structure controlled by the profile of the
32 lithosphere asthenosphere boundary may have promoted the formation of a wider African margin.
33 The focusing of the last rift axis is associated to the development of the thickest SDR wedges above
34 an ultimate necking zone of the crust. Conversely to the early stages of the rift which appears

segmented and diachronous, the final localization of the tectono-magmatic activity appears coarsely synchronous, at least, along the northern half of the austral segment. A further higher thermal gradient than elsewhere associated to a thicker and a weaker preexisting continental crust and a more intense magmatic input is inferred to have promote the development of symmetric and very wide conjugate margins close to the Walvis Ridge and Rio Grande Rise.

40

41 1. Introduction

42 The SE-Brazil/Uruguay and Namibia conjugate Volcanic Passive Margins (VPM) developed
43 during the opening of the South Atlantic Ocean austral segment (Fig. 1) and the emplacement of the
44 Early Cretaceous Paraná-Etendeka Large Igneous Province (LIP; Hinz, 1981; White and McKenzie,
45 1989; Courtillot et al., 1999; Buiter and Torsvik, 2014; Foulger, 2018). As elsewhere, the
46 development of these VPM is related to the formation of a set of spectacular wedges of seaward-
47 dipping reflectors sequences (SDR). Where outcropping (eastern and western Greenland) or drilled
48 (ODP 642 Norway, Kudu and Murombe wells in Namibia; Florianópolis High in SE Brazil), the
49 SDR consist of subaerial volcanic flows interbedded with minor volcano-detrital and continental
50 sediments (Planke et al., 2000; Larsen et al., 1998; Wickens and McLachlan, 1990; Bueno et al.,
51 2007).

52 Two main conceptual models are invoked in the formation of VPM. The SDR structure
53 would be controlled by a system of landward dipping detachment faults (Clerc et al., 2015;
54 Geoffroy, 2005; Geoffroy et al., 2015) or by flexural subsidence from each side of a magmatic
55 spreading axis (Mutter et al., 1982; Buck, 2017; Morgan and Watts, 2018). The first model implies
56 the volcanic wedges are contemporaneous with the syn-rift necking and thinning of the continental
57 basement. The second implies that most of the SDR wedges is the post-rift product of an in-excess
58 magmatism associated to spreading and continental drift.

59 Previous studies on Namibia and Uruguay/Brazil VPM have essentially identified the
60 successive SDR wedges and characterized their along-strike segmentation (Clemson et al., 1997;
61 Hinz et al., 1999; Franke et al., 2007, 2010; Koopmann et al., 2014a). Thanks to the recent high
62 quality deep multi-channel seismic profiles, the imaging of both the base of the crust, the crustal
63 fabric and the SDR internal organization can be approached (Stica et al., 2014; Clerc et al., 2015;
64 Geoffroy et al., 2015; McDermott et al., 2015, 2018; Reuber et al., 2019; Suppl. Mat. Table I). The
65 quality and density (50-100km of spacing, Fig. 1) of these datasets offer an opportunity to evaluate
66 the organization of the SDR structure both along dip and strike directions, allowing to perform a

precise mapping of their architecture and to discuss first order elements such as the along-dip crustal deformation and the along-strike propagation of the lithospheric break-up.

Similarly to other margins (Sapin et al., accepted), the structural evolution of VPM may be impacted by many different factors such as rheological heterogeneities (Callot et al., 2002; Gac and Geoffroy, 2009), the thermal architecture of the lithosphere (Simon et al., 2009; Armitage and Collier, 2018), the rift history (Armitage et al., 2010; Guan et al., 2018), the structural inheritance (Franke et al., 2007; Corner et al., 2002; Chenin et al., 2015; Petersen and Schiffer, 2016; Peace et al., 2018; Phillips et al., 2018; Reuber & Mann, 2019), the kinematic and obliquity of the rift system (Schiffer et al., 2019; Brune et al., 2018), the buoyancy forces (Bellhasen et al., 2013) and the mantle drag on the lithosphere (Clerc et al., 2018; Jolivet et al., 2018).

The development of a significant asymmetry in conjugate VPM architecture has been pointed out in several places (Eldholm et al., 2000; Mohriak et al., 2002; Mohriak & Fainstein, 2012; Becker et al., 2014). It has generally been related to similar mechanisms observed in magma-poor margins, such as crustal-scale detachment fault in a simple shear mode of deformation (Blaich et al., 2009; Becker et al., 2014, 2016), rift jumps (Buck, 2017; Norcliffe et al., 2018), or to asymmetric magmatic accretion (Reuber et al., 2019).

In this paper, we focus on a rigorous description of: the SDR wedges, the crustal thinning profile and the tectonic structures observed within the crust, on both conjugate margins. The internal structure of SDR is described and used to re-qualify the « Inner » vs. « Outer » usual nomenclature introduced by Planke et al. (2000). The evolution of the shape and of the thickness of SDR are analyzed and mapped out together with the shape and architecture and the deep crustal reflections (lower crustal package, Moho reflection(s), etc.).

The integration of the crustal structure and the SDR packages organization together within a comprehensive synthesis on the pre-rift inheritance of the austral South Atlantic domain puts the basis of a re-evaluation of the primary factors controlling its opening, the rheological behavior of the crust and the apparent asymmetry of the finite architecture of these margins.

93

94 **2. Geological setting**

The studied conjugate VPM developed between the Agulhas-Falkland (AFFZ) and Florianópolis (FFZ) Fracture Zones, as the result of the SW Gondwana break-up and opening of the austral segment of the South Atlantic Ocean (Moulin et al., 2010; Fig. 1). This ocean opened between the Rio de la Plata (RdIP), Kalahari and Congo cratons and throughout two generations of orogenic belts: the Late Precambrian Pan-African/Brasiliano belts to the north and the Late

100 Paleozoic Gondwanides to the south, comprising the Ventana and Cape Foldbelts (Figs. 2 & 3a;
101 Pangaro & Ramos, 2012; Paton et al., 2016). The break-up axis is strongly oblique to the E-W
102 compressive structures of the Gondwanides and to the NNE-SSW trending structures of the Pan-
103 African Damara Belt. But elsewhere, the break-up axis appears as coarsely parallel to the N-S
104 structural grain of the Pan-African/Brasiliano orogenic belts (de Wit et al., 2008; Fig. 3a).

105 The austral South Atlantic opening is contemporaneous with Early Cretaceous voluminous
106 magmatism found onshore, in conjugate dykes swarms (Fig. 1; Almeida et al., 2013; Will et al.,
107 2016; McMaster et al., 2019), a set of intrusive centers identified onshore Namibia (Corner et al.,
108 2000, 2002) and in the Paraná and Etendeka continental flood basalts (Peate et al., 1992; Turner et
109 al., 1994). Offshore, this magmatism was drilled at the top of the high-frequency seismic reflectors
110 packages forming the so-called sequences of SDR (Hinz, 1981; Wickens and McLachlan, 1990;
111 Gladczenko et al., 1997, 1998; Hinz et al., 1999; Franke et al., 2007; Koopmann et al., 2014a; Soto
112 et al., 2011; Stica et al., 2014; Gordon et al., 2017; Fig. 2).

113 Assuming the geological time scale of Gradstein et al. (2012), the magnetic anomalies
114 identification ascribed to the oceanic crust domain by previous works (Fig. 2) suggests that drift of
115 conjugate plates initiated around 135 Ma (Late Valanginian/Early Hauterivian times) north of the
116 AFFZ (M11 anomaly for Hall et al., 2018; M10r anomaly for Collier et al., 2017) and around 129
117 Ma (M2 anomaly) south of the Rio Grande Rise/Walvis Ridges (Franke, 2013; Collier et al., 2017;
118 Fig. 2), meaning a 6 Myr northward propagation along a nearly 2500 km long segment.

119 The Rio Grande Rise (RGR) and Walvis Ridge (WR) form conjugate volcanic plateaus that
120 lies on an anomalously thick crust (20-25 km in Fromm et al., 2017; Graça et al., 2019) where some
121 relicts of continental basement were recovered (e.g. Foulger, 2018; Graça et al., 2019). They are
122 classically considered as the remnants of the post-breakup continuation of the excess of magmatism
123 related to the trail of the Tristan da Cunha plume (Fig. 1, O'Connor and Duncan, 1990; Peate et al.,
124 1990; Ryberg et al., 2015).

125 North of the RGR/WR, the oceanic realm developed entirely within the Cretaceous
126 Magnetic Quiet Zone, posterior to 126 Ma M0 anomaly (Moulin et al., 2010; Collier et al., 2017).
127 In this central segment of the South Atlantic, the onset of the seafloor spreading is considered as
128 Late Aptian, around 112-113 Ma, following the deposition of the salt province (Karner & Gambôa,
129 2007; Torsvik et al., 2009; Moulin et al., 2010; Perez-Diaz and Eagles, 2014). This indicates over
130 16 Myr delay for the connection between the tip of the austral and the central segments of the South
131 Atlantic. This delay is accommodated both in the buffer zone of the Santos Basin (Fig. 2, Moulin et
132 al., 2013) and in the highly magmatic conjugated segments of the WR/RGR (e.g. Stica et al., 2014).

133

134 2.1. Pre-rift basement

135 2.1.1. Proterozoic orogeny

136 The Pan-African/Brasiliano orogenic belts cover the northern two third of the austral South
137 Atlantic continental margins. They outcrop in the Kaoko, Damara, Gariep and Saldania belts in
138 Africa, and along the Dom Feliciano, Ribeira and Araçuaí belts in South America (Fig. 3a). The
139 Pan-African/Brasiliano orogenic axes form a « triple junction » characterized by the NNW-trending
140 belts along the Cretaceous margins (Brasiliano, Kaoko and Gariep belts) and a NE-trending arm
141 across the Africa (Damara, Zambezi, Mozambique and Lufilian arc belts; Gray et al., 2008, Miller,
142 2013).

143 These mobile belts developed during the Neoproterozoic-Cambrian amalgamation of the
144 Gondwana and the eastward subduction of the Adamastor Ocean (Heilbron et al., 2008). The
145 Precambrian tectono-stratigraphic units involved in these orogenic belts can be correlated on either
146 side of the South Atlantic (Fig. 3a). They include a long-lived and huge calc-alkaline continental
147 magmatic arc, a wide back-arc basin and the inverted sequences of the passive margins developed
148 on the fringes of the Congo, Kalahari and Rio de la Plata cratons (Fig. 3a; Gray et al., 2008;
149 Frimmel et al., 2013; Heilbron et al., 2008, 2017; Oriolo et al., 2018; Basei et al., 2018). The
150 widespread post-collisional magmatism and the presence of some metamorphic core complexes
151 testify of a possible NS extensional collapse of the belt between the Cambrian and the Ordovician
152 (Heilbron et al., 2008; Goscombe et al., 2018; Hueck et al., 2018; Lehmann et al., 2016).

153 The main oceanic suture would be located on the American side (Fig. 3a) and, according to
154 the works of Frimmel et al. (2011 & 2013), Will et al. (2014) and Will & Fimmel (2018), the South
155 Atlantic Ocean would opened preferentially through the relics of the closed back-arc basin located
156 south of the Kaoko belt. But, northward, the break-up axis cut through the huge magmatic arc (Fig.
157 3a).

158 159 2.1.2. Paleozoic to Jurassic

160 Following the Precambrian orogeny, Cambrian to Jurassic sedimentary successions were
161 deposited in subsiding areas (Congo, Paraná & Chaco-Paraná, Karoo basins; Milani et al., 2007;
162 Linol et al., 2015). Along the southwestern fringe of the Gondwana, the coeval sequences deposited
163 in a back-arc setting during the subduction of the Panthalassa Ocean beneath the SW Gondwana
164 (Claromeco, Ventana, Cape basins; du Toit, 1937; Milani & de Wit, 2008; Pankhurst et al., 2006;
165 Uriz et al., 2011, 2016; Tankard et al., 2012; Lovecchio et al., 2020). These basins were then
166 inverted in the Permian forming the Cape and Ventana Fold Belts which represent the northern front
167 of the Gondwanides orogeny (Fig. 3a; Hansma et al., 2016).

168 During the Early-Middle Triassic, a set of localized rifted basins develop further North,
169 along a coarsely E-W trending transtensional corridor which reactivated pre-Cambrian orogenic
170 structures including those of the Damara Belt (Daly et al., 1989; de Wit et al., 1988; Zerkass et al.,
171 2005). In the Waterberg Basin in Namibia, the coeval syn-rift successions do not exceed 700m,
172 demonstrating a very limited stretching (Fig. 3b; Wanke et al., 2000; Smith and Swart, 2002).

173 From the Triassic up, a global aridification led to development of a giant desert through SW
174 Gondwana. The Botucatu and Etjo fluviatile and aeolian sands are found beneath and intercalated at
175 the base of the Paraná and Etendeka flood basalts (Jerram et al., 1999, Waichel et al., 2006). These
176 sands are inferred to be Jurassic to Early Cretaceous in age (e.g. Wanke et al., 2000; Milani et al.,
177 2007).

178

179 2.2. Mesozoic rifting

180 Several rifted basins are found along the southern half of the austral South Atlantic margins:
181 the Outeniqua, Malvinas, North Falkland, San Jorge, Rawson, Valdes, Colorado, Salado and Punta
182 del Este basins (Fig. 3b). These early basins trend obliquely to the South Atlantic axis and are
183 related to polyphase Jurassic extensional events that predate the South Atlantic margins
184 development (e.g. Franke, 2013; Frizon de Lamotte et al., 2015; Lovecchio et al., 2018 & 2020).
185 The trend of these early rifts is strongly controlled by pre-existing orogenic structures (Fig. 3b;
186 Jacques, 2003; Pangaro and Ramos, 2012; Chernicoff et al., 2014; Franke, 2013; Autin et al., 2013;
187 Salomon et al., 2017; Lovecchio et al., 2018; 2020). For instance, the Salado and Punta del Este
188 basins are localized along a 2.1 Ga suture zone embedded within the Rio de la Plata Craton (Reuber
189 and Mann, 2019; Fig. 3b). The undated syn-rift volcano-sedimentary units infill WSW-ENE-
190 trending half-grabens which may exceed 6km in thickness (Stoakes et al., 1991; Tavella and Wright,
191 1996; Ucha et al., 2004; Tugend et al., 2018).

192 The South Atlantic rift is classically presented as ranging from Late Jurassic/Early
193 Cretaceous to Hauterivian and the post-rift deposits as starting in the Barremian (Jungslager, 1999;
194 McMillan, 2003; Cartwright et al., 2012; Broad et al., 2012). This timing was mainly defined in the
195 poorly developed grabens and hemi-grabens, trending parallel to the margin along the South Africa
196 and southern Namibian shelves (Fig. 3b). The associated syn-rift infill is less than 3 km thick
197 (McMillan, 2003). The top of the syn-rift successions is dated Valanginian-Hauterivian (140-130
198 Ma; McMillan, 2003, Broad et al., 2012) and the base was never drilled. Off South Africa, these
199 successions consist in continental sediments (fluvial claystones, sandstones, and pebble beds),
200 volcanoclastics, and volcanics in places (Broad et al., 2006; McMillan, 2003).

201 On the Argentina conjugate margin, Franke et al. (2007) note the lack of major rift structures
202 developed in the upper crust. Northward in Uruguay-Brazil, syn-rift structures are also poorly
203 developed and form small basins of about 1 km thick (Stica et al., 2014; Morales et al., 2017;
204 McDermott et al., 2018). The major part of the Pelotas basin (Fig. 2) would thus be floored by the
205 Dom Feliciano basement and the Pelotas batholith. The Permo-Triassic pre-rift sequences of the
206 Paraná Basin were however drilled at the base of the Punta Del Este rift and in the northern part of
207 the Pelotas Basin (Verovlasky et al., 2004; Bueno et al., 2007; Stica et al., 2014).

208

209 2.3. South Atlantic VPM development

210 The most conspicuous phase for South Atlantic magmatism corresponds to the emplacement
211 onshore and in a very short time of the Paraná-Etendeka traps: 131-133 Ma according to Renne et
212 al. (1992, 1996a, b) recently reevaluated to 134-135 Ma by Baksi (2018). Onshore Namibia, the
213 Etendeka traps lie either on Karoo sediments or directly on the Damara basement, indicating that
214 the Karoo cover was largely eroded before the Cretaceous events (Dauteuil et al., 2013). The
215 widespread Paraná Traps reach about 2 km in the central part of the Paraná Basin. The volcanic
216 sequences pinch-out toward the South Atlantic margins where Precambrian rocks crop out near the
217 coastline (Bueno et al., 2007; Gordon et al., 2017).

218 A younger and voluminous magmatic activity is evidenced in the same area. The different
219 generations of dyke swarms that outcrops in Namibia and SE Brazil (Fig. 2) span from 135 to 113
220 Ma (e.g., Almeida et al., 2013; Will et al., 2016). The recent works of Will et al. (2016) and
221 McMaster et al. (2019) document the petrogenetic evolution between three main generations of
222 tholeiitic dykes. The two older generations exhibit very similar compositions with the Parana-
223 Etendeka basalts and would be contaminated by an enriched component (lithospheric mantle or
224 continental crust). By contrast, the younger generation of dyke shows a more ‘primitive’
225 composition deriving from a depleted asthenosphere source. These results suggest the progressive
226 thinning of the lithosphere between 135 and 113 Ma (Will et al., 2016; McMaster et al., 2019).
227 Offshore, the seaward dipping volcanic sequences drilled in the northern part of the Pelotas Basin
228 range from 125 to 118 Ma (Fig. 2, 1-RSS-003 well in Bueno et al., 2007) and up to 113 Ma on the
229 Florianópolis High (Curumim Fm., Dias et al., 1994; Bueno et al., 2007; Gordon and Mohriak,
230 2015; Stica et al., 2014). A Late Aptian age was also obtained on basaltic flows erupted on the
231 northern flank of Walvis Ridge (Fig. 2, 114 Ma, Rohde et al., 2013).

232 SDR wedges are found almost all along the conjugate margins of the austral South Atlantic
233 (Fig. 2). Subaerial basalts were drilled off Namibia (Kudu wells in Fig. 2, Wickens and McLachlan,
234 1990; Clemson et al., 1997), at the top of inner SDR. Only the southernmost segment of South

Atlantic margins appears free of SDR (Fig. 2; Koopmann et al., 2014a, b). The pioneer works on seismic data covering the Atlantic VPM (Gladczenko et al., 1997, 1998; Clemson et al., 1997; Abreu, 1998; Hinz et al., 1999) suggest that SDR sequences form a coarsely continuous band of about 100 km wide all along the austral South Atlantic margins. Gladczenko et al. (1998) propose a wide (>100 km) Late Jurassic–Early Cretaceous volcano-sedimentary rift zone covered by the SDR. In Namibia, Bauer et al. (2000), using reflection and refraction seismic data, identify two main wedges of seaward-dipping reflectors separated by flat lying flows sequences. The available refraction data and gravity modelling show the existence of high-velocity and dense lower crust beneath the SDR sequences (Bauer et al., 2000; Blaich et al., 2009, 2011, 2013, Schnabel et al., 2008; Hirsch et al., 2009; Becker et al., 2014; Fromm et al., 2017; Planert et al., 2017).

Franke et al. (2007, 2010) in Argentina, Soto et al. (2011) in Uruguay, Stica et al. (2014) in SE Brazil and Koopmann et al. (2014a) in South Africa and Namibia emphasize a significant along-strike segmentation of the margin and of the earliest volcanic successions. Franke et al. (2007) divide the Argentina margin in three main segments that would be bounded by the Falkland, Colorado, Ventana and Salado transfer zones (Fig. 2). Soto et al. (2011) introduce the Rio de la Plata transfer zone (RPTZ) that lies between the Punta del Este Basin and the Polonio High (Fig. 2). Northward, Stica et al. (2014) divide the Pelotas Basin into three segments whom the boundaries locate coarsely in the continuation of the oceanic fracture zones: Meteor FZ (for the RPTZ), Chuí FZ, Porto Alegre FZ and the Florianópolis FZ (Fig. 2).

In Argentina, Franke et al. (2007) observed significant lateral offset in SDR distribution from each side of the previously quoted transfer zones. The authors observed that the width and volume of SDR tend to decrease from S to N, within each individual margin segment. They further proposed that South Atlantic break-up propagated by successive jumps, each transfer zones acting as rift propagation barriers. In their view, the decreasing in SDR width from S to N in each segment would sign a passive decompression of the mantle rather than the direct influence of a mantle plume. Franke (2013) hence proposes three successive breakup events at ~137 Ma between the Falkland TZ and Colorado TZ, 133 Ma between the Colorado and Salado TZ and 128 Ma for the segment north of the Salado TZ.

Conversely, in SE Brazil, Stica et al. (2014) observed a net increase in thickness and width of SDR toward the north. They also argue for a strongly diachronous and segmented opening between the Salado TZ and the FFZ. They note that the M2 and M0 anomalies rest on oceanic crust south of the Chui FZ, that only the M0 rest on the oceanic crust south of the Porto Alegre FZ (PAFZ in Fig. 2). They finally propose that the rifting of the northernmost segment between the PAFZ and the FFZ lasted much longer, at least till the end of Aptian. The Ar-Ar age of 113 Ma obtained on the

269 last SDR sequence of this northernmost segment indeed suggests a slower evolution of the rifting
270 (Stica et al., 2014).

271 On the conjugated margin, Clemson et al. (1997), Blaich et al. (2011 & 2013) and
272 Koopmann et al. (2014a) observe that gravity and magnetic data and the thickness and distribution
273 of SDR are also influenced by transverse-to-the-margin fractured zones, called, from S to N: Hope,
274 Namaqua, Orange, Lüderitz, Oystercliffs, Cape Cross and Walvis (Fig. 2). These transfer zones
275 correspond to inherited structures which are reactivated during rifting and SDR emplacement
276 (Clemson et al., 1997; Koopmann et al., 2014a; Corner et al., 2002; Corner and Durrheim, 2018). In
277 line with the Autseib and Omaruru major Pan-African structures, the Cape Cross TZ offsets the
278 landward boundary of SDR of about 160km (Fig. 2). The line of onshore intrusive complexes
279 Brandberg, Messum, and Cape Cross are also emplaced along these reactivated structures (Figs. 2
280 & 3b). The Orange TZ also significantly offsets the proximal structures of the Namibian margin.
281 This transverse structure was tentatively correlated to an inherited structure, the Kuboos-Bremen
282 lineament, by Clemson et al. (1997).

283

284 3. Data & methods

285 In order to constrain the 3D architecture of the conjugate South Atlantic VPM, we
286 reinterpreted 2D ION GXT long-offset seismic profiles (Fig. 1 and Suppl. Mat. Table I) and
287 additional 2D seismic surveys (Fig. 1) provided by TOTAL S.A. to pick the key horizons of the
288 Moho, the base- and the top-SDR.

289 The primary picking/mapping is done in the time domain of the seismic datasets. Regional
290 time-depth conversion laws were used to build isopach maps on each conjugate margin. In
291 Uruguay-Brazil this conversion law was built using a 3D-seismic survey from TOTAL S.A. It
292 consists in : $Z = 0.843.t + 1.46^{-4}.t^2$ (with Z equal to the depth below seabed and t equal the tow-way
293 travel time below the seabed). The extrapolation of this law to the entire Uruguay-Brazil margin
294 generates some deviations of less than 3 km in depth for the Moho and the base SDR relatively to
295 the same horizons determined on pre-stack depth migrated (PSDM) dataset from ION Geophysical.
296 We consider these discrepancies as suitable to analyze the first-order architecture of the margin. In
297 Namibia, the time-depth conversion law corresponds to a 5-layers model that was compiled to fit
298 perfectly both the Moho and the top-SDR horizons from the PSDM sections from ION
299 Geophysical.

300 The thickness maps of the crustal basement and of the SDR define the first-order
301 architecture of conjugate margins. Due to the variable qualities of the seismic data, the Moho and

base-SDR horizons have only been determined in Namibia and Uruguay-SE Brazil. But the facies and internal structure of SDR were also mapped in the Argentina and South Africa conjugate margins.

The Moho interface was coherently picked as lying below the deepest powerful reflections that lie at the crust base of the studied margins. The available refraction data were used to control our interpretation, especially for the Moho picking. A maximum of 4 km of difference is met between our interpretation of the Moho depth and those inferred from the available refraction data in Namibia (Bauer et al., 2000; Fromm et al., 2017). This difference was checked on five crossovers between studied lines and the published refraction lines. No refraction data is currently available along Uruguay/SE Brazil margin.

The GPlates (2.1.0) software (Müller et al., 2018) was used with the paleogeographic model of Moulin et al. (2010) in order to locate the conjugate profiles (Fig. 2), and to better assess the tectono-stratigraphic variability along strike before break-up. The Moulin et al.'s (2010) rotation poles were preferred mainly because they honor the intra-oceanic fractures zones. Moulin et al.'s (2010) reconstitution use microplates which boundaries perfectly match with the main transfer zones recognized such as the RPTZ (Fig. 3a & 2). The Moulin et al.'s model generates also the tightest fit between the oceanic fracture zones and magnetic marginal anomalies. We further observe that the Gondwanides northern orogenic front fit perfectly using this reconstitution (see Fig. 3a). It is convenient to note that the use of other palinspastic reconstructions can lead till up to 150 km of latitudinal offsets between these different features.

Based on our GPlates reconstruction, three conjugate seismic profiles are selected to represent the conjugate architecture of the southern, central and northern segments (# IV, V & IV in Fig. 2) of the Uruguay/SE Brazil and Namibian margins (Figs. 4 to 6, see also the reconstruction in the inset map in Fig. 7). The Moho, Base-SDR and Top-SDR are reported together with the internal fabric of both the crustal basement and of the intra-SDR reflections. Interpretations obtained on depth-converted seismic profiles are reported in the Fig. 7 for the entire 2D survey as plotted on the inset maps in Fig. 7.

The quality of the seismic data varies between the Uruguay-SE Brazil and Namibian lines. In general, intra-basement reflectivity is poorer in Uruguay-Brazil. Moreover, the thick post-rift sedimentary wedge of the Rio Grande Cone affects the seismic resolution of the deepest level of the margin (e.g. Line 040 in Fig. 5a). Such sedimentary features decrease strongly the accuracy of the base-SDR determination and the analysis of their internal reflectivity.

In both margins, the base-SDR is the most difficult interface to interpret. This is particularly true for the base of the innermost SDR which is often diffuse and badly defined, even more when it

rests on a roughly concordant seismic unit (e.g., line 6000, Fig. 6b) or in an intensively injected basement (e.g., line 6050, Fig. 7). Considering the difficulties to distinguish SDR from the underlying strongly intruded pre-rift basement, our interpretation was primarily guided by the geometric relationships. All seismic units mapped as SDR exhibit clear lateral variations of thickness (blue sequences in Figs. 4 to 7). When they occur, the underlying sequences of which the lateral variations of thickness are really equivocal are reported in dotted blue lines in interpretations (Figs. 4 to 7). Seismic units that do not show lateral variations of thickness are considered as pre-rift and reported in purple. The syn-rift reflector sequences that lie on the proximal continental shelf and which are disconnected from the SDR domain are reported in green. The reflections crosscutting the surrounding basement fabric, a layered sequence or the Moho are interpreted as magmatic injections (red lines in Figs. 4 to 6) or major faults. It is however highly likely that the SDR sequences comprise a large amount of intrusions.

4. Thinning profile of the crust

The crustal architecture of the South Atlantic conjugate VPM was identified on the basis of first order features that represent analogous elements on the conjugate profiles (Fig. 8). One of the most prominent conjugate structure is the seaward flexural hinge displayed by the top basement (Figs. 4 to 7). This key structure, called the Necking Line (NL, Fig. 8), corresponds to the onset of the significant thinning of the continental crust (Fig. 7 & 9a) and coincides almost everywhere with the landward edge for the successive SDR packages (Fig. 7). Morphologically, the NL is marked by a wavecut surface in the continent side (Fig. 4 to 7). Seaward, the thinned continental basement is almost entirely capped by SDR (Figs. 7). In that sense, the SDR domain marks the transition between the thick proximal continental crust and the thin distal crust (Fig. 8).

Moreover, when integrating the topography of Moho, the resulting crustal thickness profiles appear divided in different domains. Each of these crustal domains is delimited by more or less conspicuous inflexions of the slope of the Moho, the Proximal and Distal «Moho hinge lines » (PMHL and DMHL in Figs. 4 to 8). These Moho hinge lines are used to delimit the crustal domains namely Proximal Domain, Proximal Necking, Thinned Domain, Distal Necking and Distal Domain (Fig. 8). It is noteworthy this terminology is essentially descriptive here and only related to the main Moho hinge lines that are observed on the studied seismic lines. Therefore, it does not made references to the nomenclature developed for magma-poor passive margins, or to the associated specific tectonic processes (Jammes et al, 2009; Mohn et al., 2010; Sutra et al., 2013; Tugend et al., 2015). Also, in some locations, additional MHL lie between the distal and proximal necking areas.

369 These intermediate MHL are also mapped and described, to be as accurate as possible (Figs. 7 &
370 9b).

371

372 4.1. Proximal Domain

373 The seismic image of the Proximal Domain informs us on the pre-rift crustal basement
374 thickness and fabric. In Namibia, the thickness of the most proximal continental crust varies from
375 22 to 43 km (Fig. 9a). A domain of particularly thin crust is located between the Cape Cross and
376 Lüderitz segment boundaries. There the crust ranges between 22 and 26 km in thickness (from lines
377 5000 to 6050 in Fig. 7). In southern Namibia and off the Gariep Belt, the continental crust recovers
378 a more classical 30-36 km thickness.

379 In Uruguay-Brazil, the continental crust ranges from 32 to 35 km thick all along the
380 Proximal Domain. However, the proximal crust in the southern part of the Pelotas Basin is not
381 covered by the data (Fig. 9a). North of the CFZ, the continental crust thickened over 40km and its
382 base is not imaged by the seismic data (lines 90 in Fig. 4a & line 70 in Fig. 7).

383 The conjugate profiles indicate, thus, that the crust tends to thicken when approaching to the
384 northernmost segment of the studied margins. They also show that the Proximal Domain is never
385 associated to a significant thinning of the crust (Fig. 7).

386

387 4.2. Proximal Necking

388 In both conjugate margins, the Proximal Necking is marked by a conspicuous convergence
389 of the top-basement and the Moho (Figs. 4 to 7). The NL as previously defined is most often located
390 within the Proximal Necking domain and in general close to onset of Proximal Necking (Fig. 9b).
391 At the end of the Proximal Necking, the Moho's slope evolves from landward dipping to a flat or
392 gently landward dipping slope. This inflexion (PMHL 2) is more tenuous and not systematically
393 observed, especially on Namibian side (Fig. 7). But it is particularly noticeable on the southern
394 Namibian and Uruguayan margins (Figs. 6 & 7).

395 This inner necking most often coincides with the onset of the SDR (Fig. 7), however it tends
396 to be more landward south of the RPTZ-Lüderitz conjugated transfer zones. Here, the thinning of
397 the crust onsets prior to the SDR domain (Fig. 9b & lines 6500, 7000, 7500 in Fig. 7) and the
398 Proximal Necking is located beneath the proximal rift structures. The most obvious example lies on
399 the line 7500 where the continental crust thins from 40 to 30km before the NL and the onset of SDR
400 (Fig. 7). This thinning lies beneath a seaward dipping major fault which bounds a small half-graben
401 developed at the top of the crust (2-3 km thick). This extension is associated with a local crust

402 thinning of about 5 km (Fig. 7). A similar configuration is observed into the conjugate Punta del
403 Este basin (Fig. 9b and line drawing from Tugend et al., 2018, in Fig. 7). Here the syn-rift
404 succession reaches up to 6-7 km and its age is unknown but supposedly Upper Jurassic/Lower
405 Cretaceous (Tavella and Wright, 1996; Ucha et al., 2004; Morales et al., 2017; Lovecchio et al.,
406 2018; Tugend et al., 2018). This likely early rift offsets the Proximal Necking and the NL from
407 about 120 km relatively to the northernmost segments of the margin (Fig. 9b).

408

409 4.3. Thinned Domain

410 Seaward of the Proximal Necking, the crust never exceeds 22 km and is most generally
411 lower than 15 km. With the exception of the northern area, the South American margin presents a
412 narrow Thinned Domain (inexistent to 60 km between the RPTZ and PAFZ) and the Namibian side
413 possesses a wide Thinned Domain of 50 to 170 km within the segments III to V (Fig. 2). In the
414 northernmost segment VI, it is much more extended, forming a domain from 200 to 340 km wide
415 (Fig. 9b) on both sides.

416 In Namibian northern segments (V & VI), noticeable changes of the slope of the Moho lie
417 beneath the Thinned Domain. These slight inflexions of the Moho are distinguished as
418 “intermediate hinge lines”. They were tentatively distinguished on northern Pelotas Basin profiles
419 but are less noticeable there (Lines 70 – 90 in Fig. 7 and Fig. 4a). In map (Fig. 9b) this intermediate
420 ramp of the Moho merges with the DMHL 1 toward the south, and this, on both conjugate margins.
421 Intermediate Moho hinge lines are also observed on the conjugate southernmost segments (III, Figs.
422 7 & 9b). In Uruguay, they lie between the Punta del Este Basin and the SDR wedges. In this
423 segment the proximal rift structures testify that the crust undergone a significant stretching
424 landward of the SDR domain (Tugend et al., 2018; Fig. 7). In such configuration, the Proximal
425 Necking records an earlier rifting stage and the intermediate MHL record the syn-SDR thinning of
426 the crust.

427

428 4.4. Distal Necking

429 The DMHL 1&2 delimit a significant necking of the crust in the outer domain of the margin
430 (Figs. 7 & 8). The DMHL 1 corresponds to a significant increasing, sometimes abrupt, in the slope
431 of the Moho toward the Distal Domain (Fig. 6a-b, 5b). In Uruguay, the average slope of the Moho is
432 close to 3-4° beneath the Thinned Domain and increases brutally to reach between 10 and 15° in the
433 Distal Necking (lines 4500 & 4700 in Fig. 7). In Uruguay-SE Brazil, South of the PAFZ, the onset

434 of the Distal Necking is located between 60 and 140 km from the NL (Fig. 9b). In Namibia, it lies
435 further offshore, between 120 to 340 km from the NL.

436 The outermost Moho slope inflexion (DMHL 2) is the most conspicuous one and lies
437 between the Distal Necking and the Distal Domain (Fig. 8) which is characterized by either a sub-
438 horizontal Moho or a downward bulge-like topography (Figs. 5 to 7). The DMHL 2 is less
439 noticeable toward the north and the conjugate Florianópolis High-Walvis Ridge (Figs. 4 & 7). In
440 this highly magmatic area, the base of the crust stays at a greater depth than in the south precluding
441 a clear differentiation between various Moho slopes. South of the influence of the WR/RGR
442 magmatism, the end of the Distal Necking (DMHL 2) corresponds to a minimum of the basement
443 thickness (Fig. 4 to 8). This area is thinner than the crust located seaward and has been called the
444 crustal thin point (CTP) (Geoffroy et al., 2019).

445 South of the PAFZ, the distance between the NL and the crustal thin point ranges from 80 to
446 180 km in Uruguay-Brazil, and between 160 and 260 km in Namibia (Fig. 9b). The same distance
447 increases dramatically in northern Pelotas Basin where it reaches 340 km (line 090, Fig. 4a). On the
448 conjugate profile, south of Walvis Ridge, a distal Moho inflexion is not clearly observed (Fig. 4b).
449 Either this distance is even larger than in Brazil either the huge magmatism of Walvis Ridge would
450 preclude the formation of a similar distal necking.

451 452 **5. Seaward-Dipping Reflectors description, typing and distribution**

453 The top-SDR unconformity is a high-amplitude seismic marker. In the distal part of the
454 margins, it merges with the top of the flat-lying flows (Figs. 4 to 7). This unconformity represents a
455 very smooth depositional surface onlapped by the post-rift sediments (Fig. 10). The uppermost
456 reflections of SDR packages are generally flat (e.g. Fig. 10a-b-c) or present apparent toplaps up dip
457 terminations (e.g. Fig. 10d). But in detail, it is very hard to demonstrate the erosional character of
458 these terminations which most likely characterize the landward thinning of the flow sequences.

459 The base-SDR horizon is of unknown nature. It has been consistently picked as the base of
460 the seaward wedges independently of its possible nature (fault, detachment, level of intrusion,
461 facies/composition lateral variation, etc.).

462 From 4 to 7 different wedges of SDR can be individualized. Each sequence is bounded
463 either by a seismic unconformity or a noticeable contrast in the acoustic impedance or internal
464 architecture (Fig. 10). The detailed mapping of each succession was not possible due to the wide
465 spacing of available seismic lines. The volcanic sequences are thus regrouped into three types of
466 SDR which were distinguished on their internal structure (Fig. 10).

467

468 5.1. Seismic facies

469 5.1.1. SDR sequences

470 The seismic facies of SDR sequences show a great variability along-dip of the margin, as
471 well as within a single sequence (Fig. 10). The first emplaced SDR wedges are generally made of
472 packages of thick, long and high-amplitude reflections (Fig. 10a, b, c). This facies contrasts with the
473 SDR lying on the outer domain of the margin. Their internal reflectivity tends to be much lower
474 (Fig. 10d, e, f) with relatively continuous low amplitude reflections in between which high
475 amplitude and long reflectors can be intercalated (Fig. 10d). Theses outermost SDR show a higher
476 amount of intrusion figures such as saucer-shaped and transgressive sills (red arrows in Fig. 10d).
477 This higher frequency of intrusions must be taken with caution. It can only be due to a higher
478 contrast of impedance between intrusions and the volcanic host rocks or to the dip of the sequence
479 at the age of the intrusion. It is also possible that the concordant high-amplitude reflections could
480 also correspond to large sills intruded within a relatively homogeneous pile of lava flows.

481 This two-fold classification between the highly reflective and poorly reflective proximal
482 (inner) and distal (outer) SDR (also in Planke et al., 2000) is far from representative of the intrinsic
483 variability of internal facies. The majority of inner-most SDR wedges present alternating high and
484 low amplitude reflectors packages (Fig. 10a, b, c). Moreover, lateral facies variations are also
485 frequently observed with high-amplitude units passing laterally to a transparent facies without any
486 peculiar structural discontinuities (Fig. 10c). This evolution of seismic properties from high to low
487 amplitude reflections could either sign a change in the type of volcanic constructions (quantity of
488 sediments intercalated, sub-marine or continental lava emplacement, etc.), a change of rocks
489 petrology as it was evidenced in Norway (between the Vøring Lower and Upper SDR Series;
490 Abdelmalak et al., 2016b); or can be due to an acoustic blanking effect caused by the overlying
491 flows (like in Fig. 10c). On that point, the seismic reflection data only is not conclusive. In addition
492 to the flat-lying packages (Fig. 10d, e), some distinctive geometries may be observed within the
493 SDR wedges.

494

495 5.1.2. Lava deltas

496 Lava deltas are breccias of volcanic glass formed where flows reach the water and are thus
497 the most relevant volcanic construction to constraint the base-level elevation at the time of the
498 volcanic activity (Planke et al., 2000 & 2017; Abdelmalak et al., 2016a). Though rare in the south
499 Atlantic, such lava deltas are observed on the southward-dipping flank of the Walvis Ridge (Fig.

11a and line 3000 in Fig. 7) and Florianópolis High conjugate volcanic plateaus (Fig. 7 line 070, ~ km 150; line 3000, km 200 to 250, from 4 to 8 km in depth). In Namibia, the 3D seismic view (Fig. 11b) indicates the delta prograded toward both the SE (off the WR) and the NE (landward), in a maximum of 1000m deep basin. This observation shows that the margin segment localized south of Walvis Ridge (n°V in Fig. 2) was partially drown during the development of the northward-most subaerial SDR (Fig. 11a). On the Florianópolis High (Fig. 2), similar hyaloclastite deltas develop at the top of SDR in a basin of a maximum of 400m depth (line 070 in Fig. 7). The Walvis/Florianópolis Ridge marked thus an important barrier separating a subaerial realm in the north from a subaqueous domain in the south up to the Aptian/Albian transition (Stica et al., 2014; Bueno et al., 2007).

More reduced volcanoclastic deposits are also inferred at the landward edge of some sequences flows but also in large sag-type basins developed in central part of the Namibian thinned crust. These sag basins form between two topographic highs, the Necking Line and the seaward coeval SDR (Fig. 10g, h and in Fig. 7, lines 3000-3500-4500-5000-5500-6500). These depocenters (~100 km wide and maximum of 2-3 km thick) likely derive from a flexural deformation of the lithosphere during the subaerial development of the outer-more wedges. It is worth notice that shallow marine invasion could have begun in the proximal domain prior to the distal domain where continental SDR wedges were forming.

5.1.3. Volcanoes and eroded volcanic edifices

Volcanic edifices are locally preserved within SDR packages (Fig. 10g). They are always featured by an area (in 2D section) of chaotic internal seismic facies and by an inflexion of surrounding reflections (Fig. 10g). A high density of intrusions is often observed in their neighborhood. Such edifices are observed within and at the base of SDR, but also within flexural-shape basins (line 5500 and 4500 in Fig. 7). A seismically highly disrupted zone underlies these edifices. The associated chaotic facies may reach the lower crust suggesting the overlying edifices could be seated on long-lived magmatic injection area (Fig. 10g).

5.2. Geometric attributes

In addition to the seismic character, the geometry of the wedges must be considered in order to be more accurate in their typing. By doing so, three main types of SDR can be distinguished based on both the curvature of the reflectors and the topography displayed at their downdip terminations (Fig. 12). Two highly different patterns were found (inner-type vs. outer-type) and a third, which is the most widespread (intermediate-type) that shows intermediate characteristics. The

534 outer-type has been also divided into two sub-types (outer-types 1 and 2) to be as accurate as
535 possible.

536

537 5.2.1. *Inner-type: Planar-diverging wedges*

538 These sequences systematically correspond to the innermost and oldest emplaced SDR.
539 They show repetitive and coherent seismic pattern characterized by a seaward thickening wedge
540 with planar-diverging reflections (Fig. 10a, b, c and 12). Internal reflections pinch out landward
541 toward a single point (Fig. 12). The length of reflections corresponds roughly to the length of the
542 wedge. But this length may strongly vary, from 10 km (line 7500) up to 80 km (lines 7000 – 6500 in
543 Fig. 7). The typical thickness of these wedges ranges from 2 to 4 km and locally up to 6 km in
544 northern Pelotas half-grabens (Figs. 5a and 7).

545 Their geometric character, a fan-shaped package associated to a single rotation point,
546 suggest that they develop as tilted blocks bounded by continent-ward dipping normal fault. The
547 wavelength of the panels would be controlled by the detachment depth (in accordance with inner-
548 SDR development model of Geoffroy, 2005; Paton et al., 2017; McDermott et al., 2018). These
549 normal faults are not always identified on seismic profiles, but some independent features point
550 clearly to their presence. They are often underlined by powerful reflections cross cutting the
551 basement fabric and interpreted as magmatic injections or hydrothermal alteration within the
552 damage zone of the fault (e.g., Fig. 10b, d). When the fault plane is not clearly imaged, it is inferred
553 by significant inflexions and/or truncations in the deeper crustal reflections (Fig. 10a, b, c).

554

555 5.2.2. *Intermediate-type: Arcuate-diverging wedges*

556 This sequence is characterized by more arcuate internal reflections with downward
557 terminations dipping from 10 to 20° seaward (Fig. 10d). The landward tip of these intermediate
558 SDR is often very close to the pinch out of the inner wedges. This argues for a very smooth and low
559 relief of the margin during their development.

560 Intermediate SDR organizes in successive wedged sequences, with the up-dip tip of
561 individual wedges migrating seaward (Fig. 10d, 12). All together, they form systematically larger
562 and thicker volcanic wedges than the inner-type (Figs. 4 to 6). Their length ranges from 50 to 200
563 km (Fig. 13). Their thickness reaches up to 18 km in the northernmost segment (VI in Fig. 2, Figs. 4
564 and 7) but typically ranges from 6 to 10 km in the southern segments (Fig. 13). The related
565 sequences seal the underlying units which can be over-tilted and eroded (Fig. 5, 5, 9d). The base of
566 these SDR packages is unclear and the identification of single and large detachment fault even more

doubtful. The global curvature of the wedges does not seem consistent with the functioning of only one growth fault. We however recognize several fault-type offsets of seismic markers in the underlying and highly intruded basement. These packages are thus likely controlled by a set of low angle normal faults (Fig. 10d) strongly intruded. The progressive rotation and the apparently sequential deformation of the hangingwall likely forces the successive sequences of flows to offset seaward indicating important horizontal displacement.

Locally, these arcuate intermediate-SDR form «laterally confined volcanic successions» recently described by Norcliffe et al. (2018) in the northern Orange Basin. The lines 7500 (Fig. 7) and 6500 (Fig. 10h) in Namibia features this type of structures, which exhibits double-verging sequences (DVS) of volcanic wedges (Fig. 12). A conspicuous DVS is also imaged in the distal part of the northernmost seismic lines of Namibia (Fig. 4b and lines 2700 and 3000 in Fig. 7). This DVS can be mapped as a coarsely NS-trending axis which abuts against the FFZ and/or the Walvis Ridge (Fig. 13b). Similar but less well-developed DVS are recognized in lines 4500 & 5000 (Fig. 7).

These structures resemble to aborted rifts (Fig. 11). The presence of such structures shows that an important hesitation in rift localization must have accompanied the South Atlantic opening. Outside of these aborted volcanic rifts, the stratigraphically coeval volcanic sequences form « classical » inner-type seaward dipping volcanic wedges. It is noteworthy that sag-type volcanic or volcanoclastic basins tend to develop above these fossil rifts, after or during the transient stage of the rift jump (Fig. 10h and Fig. 7).

5.2.3. Outer-type 1: Strongly arcuate-diverging wedges

The outer-SDR are poorly to not really defined in the literature, mainly because of the poorly resolved and scarce available seismic images of these distal-most sequences (Hinz, 1981; Hinz et al., 1999; Planke et al., 2000; Berndt et al., 2001). Planke et al. (2000) define outer-SDR as being separated from the inner-SDR by an outer-high, which would correspond to a volcanic mound or a rotated fault block delimited by a continent-ward dipping fault against which the inner-SDR develops. According to the same authors, outer-SDR would correspond to sequences of deep marine sheet flow or pillow-basalt volcanics which would contrast with the subaerial flood basalts emplaced in the inner-SDR. They also point that outer-SDR are smaller and weaker with less prominent internal reflections. Franke et al. (2007) use terminology of outer-SDR as exclusively related to highly arcuate pattern of thin and discontinuous reflections located in the outermost part of the Argentina margin. The authors show that such shape of SDR are not exclusively found in relation with outer highs. In accordance with Planke et al's (2000) interpretations, Franke and

600 coworkers (2007) infer that these outer wedges have been emplaced under subaquatic conditions
601 due to the progressive subsidence of a volcanic-magmatic crust.

602 The definition of outer-SDR can be reevaluate in the light of the impressive quality of the
603 ION south Atlantic seismic dataset. Indeed, the geometric attributes of the outer-type 1 are very
604 peculiar and they must be distinguished from the inner-most and much less arcuate outer-type 2.

605 The outer-type I SDR present diverging reflections which are more arcuate than
606 intermediate SDR, with apparent maximum dip ranging from 20° to 30° seaward (Fig. 10e). The
607 downdip termination of reflectors dim on a seismically badly defined basal surface which
608 corresponds to a sub-horizontal to continent-ward low-dipping irregular discontinuity (Figs. 10e and
609 5b). Powerful and discontinuous reflections interpreted as intrusions occur beneath the basal contact
610 indicating this contact corresponds to a preferential injection zone of magma. These wedges not
611 only growth seaward but also subsides all along with time as suggested by backstepping landward
612 onlaps (mimicking in seismic aggrading foresets as observed in sedimentary basins; Fig. 10e). This
613 suggests a distinct isostatic or/and flexural behavior of the underlying crust.

614

615 5.2.4. Outer-type 2: Planar to lowly arcuate and poorly diverging wedges

616 The most distinctive attributes of the outer-type 2 are their location in the outermost part of
617 the margin, and the noticeable decrease of the coeval magmatic production (Fig. 10d, e, f). They
618 form sequences of generally lower than 2 km in thickness (Fig. 10d) outside of the influence of the
619 highly magmatic northern segment. When approaching this northern area, the outermost sequences
620 represent very wide (~100km) and slightly thicker unit (< 4 km) of planar or poorly arcuate
621 reflections (Figs. 10f, line 3000 in Fig. 7).

622 In the 2/3 south of the Namibian margin, they form thinner seismic units (<2km thick)
623 which are difficult to distinguish from flat-lying isopach sequences of flows (Fig. 10d, e). They
624 frequently onlap the previously emplaced outer-type 1 or intermediate-type wedges indicating that
625 the corresponding flows or volcano-detrital material tend to fill a residual low formed at the end of
626 the previously emplaced SDR (Fig. 10e). As they most often merge laterally with the flat-lying
627 flows (Fig. 10d, e) they also can be considered as passively flexured coeval flows (Fig. 10d and 12).
628 They thus bring an ambiguity regarding the identification of the seaward boundary of the SDR
629 domain. In this study, the criteria used to localize the seaward boundary of the SDR domain is the
630 inflexion displayed by the top of volcanics (Fig. 12). Landward of the SDR seaward boundary,
631 outer-type 2 sequences are slightly but significantly tilted seaward. Their base is generally poorly
632 tectonized (Fig. 10d, e, f). Seaward of the SDR end, the coeval sequences became flat-lying or

633 slightly tilted continent-ward. The SDR seaward boundary is the most often coinciding with a
634 paleo-scarp, not necessarily associated to a fault, that isolates a marginal basin passively infilled
635 with post-rift sediments (Fig. 10d, e).

636 In Namibia as in Uruguay-Brazil, the outermost volcanic successions are characterized, in
637 accordance with Planke et al.'s (2000) descriptions, by a chaotic, low amplitude facies but also by
638 either some short and thin wedges of lava (Fig. 10e, f), or by very long and relatively thin seaward
639 tilted sequences of planar-diverging flows (Fig. 10d). Everywhere, they override the footwall of the
640 intermediate-SDR and merge seaward with the flat-lying flows sequences.

641

642 5.3. SDR distribution in space and time

643 On maps, the distribution of the SDR types (Fig. 13b) corresponds to the horizontal
644 projection of their downdip termination, except for the first wedges of SDR for which all the area
645 covered by the corresponding lava flows is mapped (Figs. 4 to 6).

646

647 5.3.1. Along-dip distribution

648 In the South American plate, the inner-SDR form a band of 30 to 80 km wide where the
649 volcanic sequences reach 4 to 8 km thick (Fig. 13a, d). Conversely, where the crustal thinning is
650 more distributed and progressive as along the Namibian margin, the inner SDR developed along a
651 much wider domain, ranging from 80 to 150 km (Fig. 13b).

652 The distribution of the intermediate-type SDR forms a relatively symmetrical conjugate
653 domain ranging from 30 to 70 km wide, with a thickness ranging from 4 to 10 km south of the
654 PAFZ (Fig. 13a, d). When approaching the FFZ, they cover a much wider domain of about 200 km.
655 In this area, the intermediate SDR constitutes the thickest volcanic successions reaching up to 15-20
656 km thick (Figs. 13a & 7).

657 A failed magmatic rift tends to develop only on the Namibian side (Fig. 13b). The
658 corresponding sequences emplace generally after the first wedges of inner-SDR (Fig. 7) and before
659 the thick intermediate SDR wedges developed seaward of these aborted rift (lines 5000, 5500, 6500
660 & 7500 in Fig. 7). Therefore, each of them is sealed by a new sequence of intermediate SDR which
661 developed seaward and above of the basement block presently located in outer position. This last
662 block which is underlain by continental-type seismic fabric, should represent the uplifted shoulders
663 of the previously developed volcanic rift (6500 in Fig. 7).

664 The highly arcuate outer-type 1 are emplaced on the same conjugate segment of the margin,
665 confined north of the Cape Cross and Chui conjugate segment boundaries (Figs. 9d). The outer-type
666 1 structure evolves to outer-type 2 toward the south (Fig. 13b). This evolution of the internal

667 structure is correlated to a net decreasing of the SDR thickness, from close to 8 km thick (in line 70)
668 to less than 4 km thick (in line 40). In Namibia, outer-type 2 SDR are not always developed and the
669 flat lying flows tend to directly onlap the seaward flexure of the intermediate-SDR, between the
670 Cape Cross and Lüderitz TZ (Figs. 7 & 9d).

671

672 5.3.2. Along strike segmentation & timing

673 The period of activity of the transfer zones together with the SDR age strongly vary along
674 the margin.

675 The Rio de la Plata transfer zone (RPTZ) corresponds to a 100km along-strike area without
676 clearly developed SDR (Soto et al., 2011; McDermott et al., 2018; Fig. 13a). The line 4300 which
677 cross cuts this transfer zone shows SDR only at the foot of the present-day continental slope (Fig.
678 7). South of the RPTZ, the distalmost SDR of the segment #III (Fig. 2) are dissected by a set of
679 northward-dipping normal faults (Soto et al., 2011) which are not observed in the conjugate
680 segment. Here, the intra-continental RPTZ is thus clearly prolonged by an intra-oceanic fracture
681 zone along which some isostatic or kinematic readjustments occurred after the Argentina's SDR
682 development.

683 The magnetic anomalies map of Moulin et al. (2010) indicate a strong landward offset of the
684 M4 north of the RPTZ (Fig. 2). South of the RPTZ, the M4 is located in the outermost margin
685 domain whereas in the north, it is the M2 which lies at the SDR seaward boundary. This might
686 signify that SDR north of RPTZ could be younger due to a significant delay of the rift propagation
687 through the transfer zone (Stica et al., 2014). Franke et al. (2007) localized the major transfer zone
688 along the Salado TZ. Our mapping together with the offsets of the magnetic anomalies pattern
689 suggests that the segment comprised between the Salado and Rio de la Plata TZ may have
690 constituted another « buffer kinematic » as proposed for the Santos Basin north of the FFZ (Moulin
691 et al., 2010).

692 On the opposite margin, the SDR form en-echelon lens-shape basins of 300 to 400 km long,
693 which are offset by about 70 km from each side of the Lüderitz TZ (Fig. 13a). Therefore, South of
694 the Cape Cross & CFZ conjugate transfer zones, the SDR architecture is divided in two first-order
695 segments which coarsely correspond to the segments III & IV (Fig. 2). But, conversely with the
696 previously described RPTZ, the Luderitz TZ seems to inactivate just after the inner SDR
697 development. We observe that a double verging sequence of volcanic wedges developed in each of
698 the previously described segments, but only on the Namibian side.

699 With the exception of the RPTZ, the outer SDR successions of segments III & IV are much
700 less segmented along strike than the inner SDR (Fig. 13b). The corresponding sequences tend to

701 show horizontal, parallel and almost isopach sequences of reflections in parallel-to-the-margin
702 profiles.

703 On both conjugate margins, the SDR distribution is dominated by a considerable widening
704 of their extension toward the north, in segments V and VI (Fig. 13b). In SE Brazil, the SDR
705 seaward boundary is noticeably shifted to about 150 – 200 km toward the sea, only north of the
706 PAFZ (Fig. 2, Stica et al., 2014). In Namibia, the same offset is observed but more progressively
707 and begins north of the Cape Cross TZ (Fig. 13b).

708 The lines 4000 and 4500 cross over the Cape Cross transfer zone (Fig. 7). In these two lines,
709 the volcanic sequences of the northernmost and highly magmatic segment are strongly condensed.
710 They seal an unconformity that lies at the top of the over-tilted and eroded younger SDR (Fig. 7 and
711 Fig. 5b). Along strike, these first sequences of inner SDR are also controlled by a set of oblique-to-
712 the-margin normal faults related to the Cape Cross TZ. These transverse faults do not propagate in
713 overlying SDR, indicating that this major transtensional boundary has played during the
714 emplacement of the firsts inner-type SDR and that it was inactivated during the rest of the VPM
715 formation (also in Koopmann et al., 2014a). This along strike segmentation is less noticeable across
716 the conjugate CFZ.

717 Northward, the segment VI constitutes a large volcanic plateau supported by thick SDR
718 packages of 15 to 20 km thick (Fig. 4, 11a, 13). This plateau constitutes a morphological barrier
719 isolating the austral and central segments of the South Atlantic (Figs. 2 & 11a) and delimiting the
720 southern extend of the Aptian salt province (Torsvik et al., 2009; Chaboureaud et al., 2013). In this
721 northern segment, the wide and thick intermediate SDR are observed both tilted toward the sea and
722 toward the PAFZ and Walvis conjugate TZ that form crustal-scale large-offset transfer fault zones
723 (Abreu, 1998; Elliot et al., 2009; Cartwright et al., 2012; McDermott et al., 2018). The internal
724 reflectors dip toward the sea up to 20° in Namibia and up to 30-35° in the northern Pelotas basin
725 (Fig. 7), and the same wedges dip up to 10° toward the N and toward these EW-trending conjugate
726 transfer zones (Fig. 11a).

727 The stratigraphic markers correlated through the Namibian northern profiles (orange marker
728 in Fig. 7 and 11a) evidence that the development of the outer-type 1 of the line 4000 (Fig. 7) is
729 coeval with formation of the intermediate-SDR in the northernmost lines (Line 3000 & 2700, Fig.
730 7). This implies that both, the successive SDR wedges tend to rejuvenate toward the north, but also
731 that the magmatic activity may have continued in the north, well after the break-up of the rest of the
732 margin (Fig. 11a) which is confirmed by the ages of the drilled volcanics. In the two northernmost
733 profiles of Namibia (segment VI), a conspicuous double-verging magmatic structure (Fig. 4b and

lines 3000 and 2700 in Fig. 7) forms an early magmatic spreading center likely coeval with the incipient formation of an oceanic-type crust in segment V and further south (Fig. 4b).

The age of the SDR wedges in this segment VI is almost unknown but would be bracketed between 124 and 113 Ma according to the correlation proposed by Stica et al. (2014) and Bueno et al. (2007, Fig. 2). The Walvis Ridge was formed on the northern flank of this wide volcanic plateau (Fig. 11a). The ridge is characterized by an EW alignment of circular shape volcanoes. The oldest Ar-Ar ages obtained on the WR is 114 Ma (Fig. 2, Rohde et al., 2013). They come from the N flank of the ridge and outside of the seismic survey used in this study. Nevertheless, a similar age (113 Ma) was obtained on the topmost volcanic succession of the Florianópolis High (e.g. Bueno et al., 2007 and Stica et al., 2014). In accordance with the tectonic calendar proposed by Stica et al. (2014), it is thus likely that the SDR of the northernmost segment of the austral South Atlantic were forming during all the Aptian (Segment VI, Fig. 2 and 11a).

6. SDR and basement coeval structuration

In the following section, we emphasize the apparent genetic relationships that link the evolution of (1) the crustal basement mode of deformation and thinning, (2) its seismic characteristics, and (3) the SDR types throughout the identified crustal domains. Where they are available, the density and velocity properties of each domain are integrated to this description, to more accurately characterize basement nature.

6.1. Proximal Domain and Necking

On both conjugate margins, the proximal grabens and half-grabens, located landward of the NL, are scarce and limited both in depth and width, indicating the minor stretching recorded in the inner part of the margin.

In Namibia, the only well-developed proximal graben develops against the southern branch of the Cape Cross TZ (Clemson et al., 1997; Corner et al., 2002; Fig. 2) which obliquely crosscuts the 4500 seismic line (Fig. 7). A well reaching the uppermost succession of this graben (2313-05-1) indicates that the succession comprises basaltic flows and volcanodetrital sediments. Elsewhere in Namibia, the proximal syn-rift successions are generally thinner than 2 km (see also McMillan, 2003; Jungslager, 1999; Broad et al., 2012). It is noteworthy that the different extensional structures observed on the Namibian shelf tend to develop against normal faults that trend parallel to the basement's fabric and which dominantly dips toward the sea. This is well displayed in Namibian

lines 4000 (Fig. 5b), 6000 (Fig. 6b), and 6500 (Fig. 7). This strongly suggests a dominant control of the structural inheritance during the early stretching of the margin.

In Namibia, the upper crust in the Proximal Domain and Proximal Necking displays the seismic characters of a strongly deformed basement with what appears to be tectonically stacked thrust sheets and folds. This is well evidenced in line 7000 where a fold of 15 km wavelength lies in the middle crust above a barely deformed package of strong reflections. Between lines 7500 and 6500, the proximal domain is thus the most likely composed of the deformed units of the Gariep Belt (Fig. 3a). Between the Luderitz and Cape Cross segments boundaries, orogenic structures are less pronounced (Fig. 5b and lines 6000 to 5000 in Fig. 7). We inferred that this apparently less deformed upper crustal fabric reflects the change of orientation of the orogenic fabric. In the offshore prolongation of the Damara belt, seismic lines would be parallel to slightly oblique to the main orogenic fabric. North of Cape Cross, along the Kaoko Fold Belt, the upper crust displays some tight fold in a layered upper crustal unit (Fig. 5b). Further northward, the Proximal Necking and the NL tends to get closer to the coastal zone (e.g., Fig. 9b) precluding a relevant description of inherited structures.

In Uruguay-Brazil, proximal rift structures are also very seldom and mainly represented by landward dipping normal faults associated to limited vertical offset. The upper part of the continental crust is quite transparent and appears as little deformed (Fig. 4a to 6a). It rests onto a typical strongly reflective and layered middle and lower crust (Mooney & Mesner, 1992). The seismically homogenous character of the upper crust is almost constant from the Polonio High in Uruguay until the FFZ to the north. In northern Pelotas Basin, an isopach package of mildly powerful reflectors reaching 2-3 km in thickness is interpreted as a pre-rift succession preserved at the top of the crust (purple reflectors sequence in lines 70 & 60 in Fig. 7). The corresponding succession is discordant on the underlying basement fabric. It is also tilted against a set of continent-ward dipping normal faults. As formerly proposed by Stica et al. (2014), it is likely these isopach seismic units correspond to the offshore equivalent of the pre-SDR Paraná basaltic plateau.

When approaching to the NL and beneath the firsts inner SDR, the upper crustal fabric tends to become more reflective in both conjugate margins (Fig. 10a, b, c). The basement fabric is often seen tilted against a dense network of normal faults which accommodate the strong flexure toward Thinned Domain (e.g. lines 40, 4700, 90 in Fig. 7). It is also common to observe a dense network of subvertical to slightly tilted high-amplitude reflections cross-cutting the background fabric (Figs. 4a & 5a). These reflections are interpreted as sheeted-dyke complexes emplaced in the top of the crust. Some analogous highly injected flexure zones are well documented in SE Greenland and Mozambique coastal flexures. In these two cases, more than 50 % of magmatic dilatation is

800 associated with extensional faulting to accommodate the syn-SDR flexure of the upper part of the
801 continental crust (Klausen and Larsen, 2002; Lenoir et al., 2003; Klausen, 2009).

802 The inherited continental crust of the Proximal Domain can also be intruded both by some
803 decakilometric intrusive complexes that are well evidenced in Namibia (e.g. Fig. 3). Off these
804 intrusive centers, the velocity vertical gradient of the upper and lower continental crust seems
805 preserved with available Vp and density values that range from 5.9 to 6.8 km.s⁻¹ and from 2.7 to 2.9
806 between the surface and the Moho respectively (Fig. 15a, Bauer et al., 2000 & 2003). Under the
807 Messum and Cape Cross complexes, the authors estimate the density ranges from 2.9 to 3.0 and
808 they found the velocity increases of ~0.3 and ~0.5 km/s relative to the host rock, but with a similar
809 gradient till the Moho (Bauer et al., 2003). Similar igneous centers with dyke swarms lie all along
810 the W and E Greenland coastal flexure (Callot et al., 2002) and they denote the beginning of a
811 pervasive but heterogenous modification of the composition of the pre-rift crust. Off these localized
812 igneous centers, the proximal continental basement keeps its original characteristics till the
813 Proximal Necking and the entrance into the SDR domain (Fig. 14b).

814 Therefore, Proximal Domain and Proximal Necking are always associated with brittle upper
815 crust deformation expressed either by clear grabens filled with volcanic and detrital sediments or by
816 innermost SDR. The mechanical behavior of Uruguayan (line 4500) and SE-Brazil (line 90) lower
817 crust was described and analyzed by Geoffroy et al. (2015) and Clerc et al. (2015). Clerc et al.
818 (2015) observed numerous high amplitude reflections within the middle/lower crustal layers around
819 which the surrounding crustal fabric is deflected. They proposed that the resulting sigmoidal
820 geometry characterizes the occurrence of ductile shear bands (yellowish intra-crustal lines on our
821 interpretations). The 2D convolution seismic modelling performed by Wrona et al. (2019)
822 consolidate the Clerc et al.'s conclusions, showing that several seismic reflections
823 converging in the same direction and forming large-scale lenses are characteristics
824 for kilometric-scale shear zones. The studied conjugate seismic lines exhibit similar structural
825 pattern in both Proximal Domains and Necking (Fig. 14). The internal deformation of the lower
826 crust fabric suggests the occurrence of a simple shear fabric. The middle/lower crust displays
827 curved reflections that define dome- or lens-shape structures which are also consistent with a ductile
828 behavior (Figs. 7 & 14). This ductile strain appears distributed through all the lower crustal section
829 more than localized along single crustal-scale detachment.

830 From all these observations it can be concluded that the tectonic evolution of the Proximal
831 Domain and Proximal Necking are controlled mainly by active faulting that tends to be more
832 magmatic with the appearance of the inner-type SDR and strong, sheared reflections within the

833 lower crusts. The inner SDR rests on an intruded upper crust crosscut by landward dipping normal
834 faults that flatten atop or into a ductily deformed middle-lower crust (see the brittle/ductile
835 transition line in Figs. 4 to 7). Moreover, the Proximal Necking is a domain relatively symmetric on
836 both margin with a variable width of 30 to 80 km (Fig. 9b).

837

838 6.2. Thinned Domain

839 In both conjugate margins, the Thinned Domain is always covered by inner and also partly
840 by intermediate SDR (Figs. 7 & 16). We also recognize some packages of arcuate diverging
841 reflectors developed within aborted volcanic rift. These DVS are always located in the Thinned
842 Domain and exclusively on the Namibian margin (Figs. 7 & 16).

843 In this central domain of the studied VPM, the basal lower crust tends to show very high
844 amplitude and long reflections that could be isolated or formed 1 to 4 km thick packages of either
845 stacked reflectors or parallel to the Moho shearing. The most impressive example likely occurs in
846 the northern Pelotas basin (line 090, Fig. 14 and 4a) or in southernmost profiles of Namibia (lines
847 7000 and 7500 in Fig. 7). In the northern Pelotas profile, Geoffroy et al. (2015) distinguish these
848 stacked powerful reflections as a specific mafic lower crust (LC2, Fig. 14) from the overlying and
849 highly deformed fabric of the ductile middle-lower crust (LC1).

850 Along the margin, the basal highly reflective LC2 is not as evident neither as thick as in the
851 previous examples. These reflections could also be strongly disrupted or totally absent (km 200 of
852 the Fig. 6b), perturbing the Moho identification. These silent areas often coincide with evidences
853 for volcanic edifices emplaced in upper part of the crust. But when it occurs, the LC2 type
854 reflections are tectonically disconnected from the overlying LC1-type lower crust or is used as a
855 level of rooting for the LC1 shear bands (Fig. 14). This is consistent with the noticeable absence of
856 strong offsets of the Moho along the studied profiles. The base of the continental crust appears thus
857 as a decoupling layer between the middle/lower crust and the upper mantle (Clerc et al., 2015;
858 Geoffroy et al., 2015).

859 In the northern part, these lower crustal shear zones define some kinds of extensional
860 duplexes (Figs. 4 and 13). But in central and southern Namibia, the lower crust structure is rather
861 dominated by boudinage and an even more distributed shear (Figs. 5b, 6b and lines 5500, 6050 and
862 6500 in Fig. 7).

863 Our interpretation indicates that the inner SDR wedges development is concomitant with
864 thinning processes of the continental crust (Fig. 7 and Figs. 4 to 6). We observe that this thinning is
865 accommodated by brittle faulting of the upper crust and ductile shearing of the middle and lower

866 crust and a non-quantifiable proportion of magmatic dilatation. The seismic characteristics of the
867 upper and lower crust are the most often coherent all along Thinned Domain, resembling the
868 previously described layering from the Proximal Domains (Figs. 4, 5, 6).

869 Similar conclusion could be reach when regarding the velocity structure of the Thinned
870 Domain gave by the available refraction data (Bauer et al., 2000; Fig. 15b). Indeed, the velocity
871 structure appears as coherent all along the Thinned Domain and characterized, from the top-SDR to
872 the Moho, by a much higher velocity gradient than in the Proximal Domain. The Vp pass from 5 to
873 about 7 km.s⁻¹ throughout the extrusives package (Fig. 15b). Below, the crust shows high Vp
874 ranging from 6.9 to 7.6 km.s⁻¹ and characteristic of the so-called HVLC of VPM. This HVLC is
875 frequently presented as a two layer unit, with a 8-10 km thick upper layer with a relative low
876 velocity gradient (from 7.0 to 7.3 km.s⁻¹), overlying a thinner (2-6 km) lower layer with Vp > 7.3
877 and around 7.6 to 7.8 km.s⁻¹ (Bauer et al., 2000; Fromm et al., 2017; Becker et al., 2014). Along the
878 intersection between the Bauer et al.'s (2000) Transect 1 and the Namibian line 4500 (Fig. 15b), the
879 "upper HVLC" includes both the brittle upper crust and the top of the LC1 ductile crust (Fig. 15b).
880 Their high velocities must thus be linked to the high amount of magmatic intrusions. Downward,
881 the "lower HVLC" corresponds to the highly reflective seismically layered LC2 (Fig. 15) but also
882 includes a limited thickness of the silent area which characterizes the uppermost mantle in our
883 interpretation. It is therefore likely that the upper part of HVLC in VPM correspond to an
884 intensively injected and sheared lower crust, with possible lenses of inherited high-Vp lower crustal
885 units (as it is expected in Norway; Gernigon et al., 2006; Nirrengarten et al., 2014). Their lower part
886 might more composed of underplated magmatic mafic bodies.

887

888 6.3. Distal Necking

889 The Distal Necking is always associated to the intermediate and outer-type SDR (Fig. 16).
890 In both conjugate margins, the onset of the Distal Necking (DMHL 1, Fig. 16) well corresponds to
891 the onset of intermediate SDR south of the PAFZ and Cape Cross TZ (Figs. 16 & 7). When they
892 occur, outer SDR develop prior to crustal thin point which lies at the end of the Distal Necking
893 (DMHL 2 in Fig. 16), except in the northern segments where they develop after.

894 In South American profiles, the crustal seismic fabric of the Proximal and Thinned Domains
895 tends to be replaced by a noisier and much less reflective facies throughout the Distal Necking (Fig.
896 5a and 6a). The nature of this crust is thus difficult to constrain from seismic reflection profiles
897 only. Conversely in the Namibian profiles, we clearly recognize that intermediate SDR lie above the
898 same LC1 and LC2-type facies than in the Thinned Domain (Figs. 5 to 7). The most often, the LC1-

899 type fabric appears tilted toward the sea, the downdip terminations of the strongest reflections
900 showing some ductile drag against the LC2-type basal crust (Figs. 5, 6 & 7). These observations
901 were done all along the Namibian margin and suggests that continental basement is still within part
902 of the Distal Necking (Fig. 7).

903 The strong and seaward dipping reflections tend to root in the top mantle and may crosscut
904 the entire crustal basement (e.g., line 5000, 5500, 7500 in Fig. 7). These reflections are interpreted
905 as large-scale intrusions tardily injected within the lower crust and sheared with the host-rock (e.g.
906 Fig. 6b), during the seaward tilt of the crustal basement. The brittle upper part of this basement the
907 most often displays an opposite fabric, i.e., with powerful reflections inclined toward the continent.
908 The corresponding reflections are considered as intrusions injected in the footwall of the overlying
909 intermediate- and outer-type SDR (e.g. Fig. 6b).

910 In available refraction data, the “Lower HVLC” characterized by $7.3 < V_p < 7.8 \text{ km.s}^{-1}$ tends
911 to pinch-out along the Distal Necking (Fig. 15c). With the exception of this noticeable structural
912 closing, the velocity structure of the Distal Necking does not show peculiar difference with the
913 velocity profiles of the oceanic crust from White et al. (1992, Fig. 15e). This indicates that if a
914 certain proportion of continental crust does persist in this distal domain, it is blind for the refraction
915 data from Bauer et al. (2000).

916

917 6.4. Distal Domain

918 Off the crustal thin point which lies at the end of the Distal Necking (DMHL 2), the crust is
919 overlain directly by flat lying flows sequences (Fig. 5) in segments II to IV and by outer-SDR in
920 segments V and VI. The transition to the distal crust is characterized by a Moho presenting a
921 downward bulge-like topography that became flat lying and sub-horizontal further seaward (Fig. 7).
922 Along-strike, its thickness varies between 6 and 12 km (also in Taposeea et al., 2017), the thickest
923 crust lying in the highly magmatic northern segments (Line 3000 in Fig. 7).

924 The transition from the outermost SDR to the flat lying flows sequences (FLF) often lies at
925 the onset of this Distal Domain. The FLF are associated to a very smoothed topography of the
926 seafloor. At around 100km from the crustal thin point, the volcanic units tend to be progressively
927 tilted toward the continent against a dense normal fault network (Fig. 7). These closely spaced
928 fracturing gives rise to a much more rugose seafloor and a classical pattern of oceanic crust with
929 landward tilted blocks against oceanward dipping normal faults. This last type of crust is also
930 characterized by a thickness close of the worldwide average of the “normal oceanic crust” (6-7km
931 according to compilations of Christeson et al., 2019 and White et al., 1992, Fig. 15e).

932 From the end of the Distal Necking, the basement is generally invaded by powerful oblique
933 reflections interpreted as magmatic intrusions and/or shear bands (Becel et al., 2015; Ding et al.,
934 2018). They either display a X-shape geometry, or an alternance of areas of landward or seaward
935 dipping reflections. The coeval transformation of the seismic fabric is systematic along the margin
936 and very likely due to the overpassing of a threshold in the proportion of magmatic input.

937 The nature and origin of this distal domain is unclear so far. It may well represent a very
938 particular type of oceanic accretion or alternatively a kind of mixed mafic/continental crust
939 (Geoffroy et al., 2019, 2020).

940

941

942 7. Asymmetry level of the conjugate VPM

943 The observations made on the conjugate South Atlantic margins exhibit a significant
944 asymmetry, both in the crustal architecture, the SDR types repartition and the total volume of SDR.

945

946 7.1. Margin width

947 The margin width is one of the criteria used to evaluate the asymmetry level of a pair of
948 conjugate passive margins. The main boundaries of the continental passive margin s.s. in the case of
949 VPM are difficult to evaluate due to a lack of consensus in their formation and the definition of a
950 strict COB can be subject to strong debate (McDermott et al., 2018; Reuber et al., 2019; Geoffroy et
951 al., 2020). In this context, we consider more relevant to use some empirical but robust observations,
952 common in most lines, rather than to follow an a-priori genetic definition of the successive crustal
953 domains.

954 The Necking Line (i.e., the top-basement flexural hinge in Fig. 8), together with the DMHL
955 2 (i.e., crustal thin point) represent the most evident and repetitive features observed on most of the
956 studied seismic survey (Figs. 4 to 7). The Necking Line corresponds together to the onset of the
957 crustal necking and of the SDR domain. This clear feature can thus be used to delimit the onset of
958 the volcanic passive margin s.s. On the other hand, the DMHL 2 is used as a proxy for the end of
959 the volcanic passive margin. This boundary cannot be firmly interpreted as the seaward-most end of
960 the continental crust. Nevertheless, it represents a specific boundary in the thinning profile that
961 marks the transition between an in-thinning continental crust, and a crustal domain which is
962 observed as invaded by the magmatic seismic fabric. These two crustal lines constitute thus good
963 markers to evaluate the margin width, despite the lack of definitive explanation of their genesis.

964 VPM widths measures were performed along 5 conjugate lines spaced 200 km and trending
965 perpendicular to the average trend of the margins (Fig. 16 and Table 1). The profiles #2 to 5 are all
966 located south of the PAFZ and of the influence of the high magmatic supply of the northernmost
967 segment (profile #1 in Fig. 16 & Table I).

968 The margin width is of 109 ± 22 km in the Uruguay-Brazil, while the width of the conjugate
969 Namibian margin is 190 ± 40 km. This relative difference of about 60% outline the significant
970 asymmetry in the conjugate VPM (Fig. 9b & 16). Conversely, North of the PAFZ, the margin width
971 reaches close to 300km on the Florianópolis Plateau and about 310 km on the southern flank of
972 Walvis Ridge (S of the line 2700). Here, the margin width is thus symmetric (Table 1).

973 Looking at the different neckings (both proximal and distal), the thinning profile of the
974 basement of the South Atlantic margins is significantly asymmetric. The top-basement and Moho
975 define a strong necking of the continental crust in Uruguay/SE Brazil. This is noticeable in the
976 central and northern segments of the Pelotas Basin where the continental crust thins of more than
977 50 % in less than 70 km and still less locally (in 30km for line 040, Fig. 5a) and the two Necking
978 Domains are nearly contiguous with narrow to undifferentiated Thinned Domain (Figs. 7, 9b, 16).

979 On the opposite margin, the necking of the crust is much gentle. Off the Necking Line, the
980 crust exhibits rather a progressive and distributed thinning with local failed rifts (Figs. 7 & 9a). An
981 ultimate zone of necking lies in the outermost part of the conjugate profiles. Between the DMHL 1
982 & 2, the Moho slope suddenly increases before it reaches the shallowest depth. In Uruguay/SE
983 Brazil, this Distal Necking is much less noticeable than in Namibia, and when it is observed
984 (Uruguayan lines), it lies in a much more proximal location (Figs. 7, 9a & 16).

985

986 7.2. SDR distribution

987 When comparing now the distribution of the different types of SDR (Figs. 13b & 16), it is
988 remarkable that the asymmetry is bared by the inner and intermediate SDR which cover a much
989 wider area in Namibia than in the opposite margin (Table 1).

990 South of the PAFZ, the inner SDR define a domain with an average width of 43 ± 9 km in
991 Uruguay-Brazil against 119 ± 42 km in Namibia. These values do not change significantly in the
992 northern highly magmatic segment (Table I). This 1:2 ratio in the conjugate distribution of the
993 inner-type SDR is related to the highly different conjugate thinning gradients, characterized by the
994 strong crustal necking in Uruguay-Brazil and the more distributed thinning in Namibia (Fig. 7 & 9).

995 South of the PAFZ, the intermediate-type SDR plot on a band of relatively similar width on
996 both margins whose average is about 45 ± 10 km (Table I). However, the Namibian side presents

the segmented failed axis (DVS) developing local intermediate-type within the inner-type (Fig. 7). Independently of the width of the domain, this intermediate-type can be considered as another asymmetric element into the SDR distribution. Their extent increases dramatically to reach 180 km in northern Pelotas basin and 140 km in northern Namibia. Once again, the asymmetry in the SDR distribution tends to vanish toward the northern segment of the conjugate margins.

The outer-types SDR distribution is more hazardous to compare when considering the difficulty to distinguish the outer-type 2 from the outermost flat-lying sequences of lava (Fig. 12 & 7). A relative symmetry seems to occur between outer-type 1 SDR of the two conjugate margins (Fig. 16, Table I). This is not the case for the outermost sequences of outer-type 2 which, in the central and southern segments, are developed on a wider area on the American plate than on the African plate (Fig. 16 & Table I). This is particularly well evidenced in the central segment of the Namibian margin where flat-lying flows directly onlap the last intermediate-type SDR and their thinned crustal basement (line 5000, Fig. 7).

7.3. SDR volume

The base- and top-SDR horizons were used to calculate the thickness and estimate the SDR volume developed from the FFZ to the southern Namibian border and on the conjugate VPM. For these conjugate areas which correspond to the 2:3 of the total length of the South Atlantic VPM, the estimated SDR volume reaches $1.82 \times 10^6 \text{ km}^3$ in Namibia and $1.15 \times 10^6 \text{ km}^3$ on the South American margin. The Namibian margin would thus contain about 60 % more of erupted magma than its conjugate. Peate et al. (1992) and Milner et al. (1992) have estimated the volume of the Etendeka and Paraná Continental Flood Basalts to $0.07 \times 10^6 \text{ km}^3$ and $1.2 \times 10^6 \text{ km}^3$, respectively.

Assuming that the SDR consist entirely of volcanic material, and considering that our estimation concerns only the 2:3 of the margin, the total volume of melt reaching the surface during the formation of South Atlantic VPM would be three time the estimated volume for the present-day preserved onshore Paraná-Etendeka traps.

At the scale of the South Atlantic VPM, these results imply that the asymmetry in the magma volume and the asymmetry in the margin width are of an equivalent level, estimated by a relative difference of about 60 %.

8. Discussion

Our interpretations of long offset seismic lines provide a new view of the magmatic elements, crustal architecture and evolution of South Atlantic VPM. We propose a new definition of

the SDR packages, their distribution (Fig. 13b) along dip and strike of the margin and their correlation to the thinning profile of the crust. This allows us to propose a sequence of breaking up of the crust in this hyper-magmatic environment. In addition, the new maps of the types of SDR, their thickness profile (Fig. 13a), together with the thickness of the underlying basement (Fig. 9) and its seismic character (Fig. 4 to 7) add new elements to the already observed asymmetry of the conjugate margins. These systematic observations raised several questions on the parameters that may influence this asymmetry.

8.1. Crustal deformation and SDR formation

The definition of an intermediate facies of SDR allows to propose a refined scheme of the mode of formation of the successive SDR packages. The specific relationship between the thinning gradient of the crust and the geometric attributes of the SDR suggests a likely genetic relationship. The change from inner SDR to intermediate SDR coincides with the initiation of an ultimate necking of the crust (Fig. 16). The outer SDR domain is intimately associated to the transition from this ultimate necking to the thinner and more magmatic crust of the Distal Domain (Figs. 8 & 16). Therefore, intermediate SDR tend to develop during a transitional tectono-magmatic regime, between inner and outer SDR.

The Proximal Necking is related to a mechanical thinning of the crust as in any other margins (Sutra et al., 2013; Tugend et al., 2015), where active faulting of the upper crust controls the formation of extensional basins. Here they are filled with wedged volcanic sequences, qualified as inner SDR (Fig. 10). These inner SDR rest on an intruded brittle upper crust faulted by syn-magmatic landward dipping normal faults that flatten on a ductile middle-lower crust of continental nature as suggested by the velocity profiles from refraction data (Bauer et al., 2000). The middle-lower crust is intruded and sheared as imaged by multi-channel data (this study; Clerc et al., 2015).

The diverging and mildly arcuate shape of the intermediate SDR suggests they are related to a more advanced stage of rifting relatively to the inner SDR stage. Their large flexuration, the sheared structures observable in the lower crust and the appearance of a lower crustal unit in refraction and multi-channel data tend to suggest a much higher amount of magmatic material in association to a stronger thinning of the crust. The increase of internal curvature and the seaward migration of their pinch-out (Fig. 10) also implies a loss of flexural strength of the crust during their development (White & Smith, 2009). The arcuate shape of SDR together with the seaward migration of the successive wedges might be explained by the symmetrical development of flexural basins on each side of a sub-aerial spreading axis (Hinz, 1981; Mutter, 1982; Corti et al., 2015;

1063 Buck, 2017; Morgan and Watts, 2018). However, the seismic profiles exhibit some low-angle
1064 normal faults forming small tilted blocks (Fig. 10d) associated to the development of the
1065 intermediate SDR. This observation would suggest a sequential extensional faulting of continental
1066 basement associated to a migration of the magmatic activity (Gibson and Love, 1989; Eldholm et
1067 al., 1995; Geoffroy, 2005; Mjelde et al., 2007; Stica et al., 2014; Geoffroy et al., 2015, 2020; Clerc
1068 et al., 2015) on top of a heavily intruded and sheared ductile crust that tends to shallow beneath
1069 these sequences (Fig. 7).

1070 Along most of the margin, intermediate SDR are followed by outer SDR type 2, or in few
1071 places, directly by FLF sequences. This progressive thinning of the outermost extrusives is the
1072 likely diagnostic of the coeval decrease of the magmatic production. This evolution coincides with
1073 the transition between the Distal Necking and the Distal Domain which seems to be nearly to fully
1074 composed of magmatic material as suggested by refraction data (Fig. 15c; Bauer et al., 2000) or
1075 gravimetric inversion (Maystrencko et al., 2013).

1076 The outer SDR type 1 are confined to the northern area (Fig. 16). It is thus obvious that
1077 some specific genetic conditions are met for their development. The high curvature of these SDR
1078 suggests that the lava pile is deflected on a partially molten or a very weak basement. This is also
1079 inferred by the very low dip of the structural discontinuity which separates the SDR from lower unit
1080 (Fig. 5b & 10e). A higher thermal gradient and/or magmatic production than southward is likely,
1081 due to the proximity of the Walvis Ridge and Rio Grande Rise. The weak behavior of this outer
1082 basement could derive from a pervasive partial melting of the continental basement (Kjøll et al.,
1083 2019) and from the total replacement of the continental crust by a gabbroic mush, at the manner of
1084 the oceanic crust (Pindell et al., 2014; Quirk et al., 2014).

1085 One of the major unresolved questions concerning VPM formation is whether the outer SDR
1086 developed after the continental breakup and are therefore oceanic (Hinz, 1981; Mutter, 1982; Corti
1087 et al., 2015; Buck, 2017; Morgan and Watts, 2018), or above a heavily intruded continental crust
1088 (Eldholm et al., 1995; Stica et al., 2014; Geoffroy et al., 2015, 2020; Clerc et al., 2015; Senkans et
1089 al., 2019). The final breakup point that defines the end of the continental crust (or Continent-Ocean
1090 Boundary, COB) is therefore highly debated. The lithospheric extension that forms the VPM is not
1091 iso-volumetric. Massive magma accretion/intrusion accompanies the tectonic stretching of the
1092 continental crust since the earliest SDR emplacement. Onshore field analogues demonstrate that the
1093 crustal dilatation caused by diking can reach up to 50 % in the most proximal SDR (Klausen and
1094 Larsen, 2002). The refraction data (Bauer et al., 2000) indicate the velocity/density of the
1095 underlying crust evolve from usual continental velocity trends from the Proximal Domain to much
1096 faster velocity beneath the Thinned Domain, suggesting an increasing magmatic content in the

1097 crust. But, despite this magmatic input, the analyzed seismic data show that the lower crustal (LC1
1098 and LC2) fabric of the Proximal and Thinned Domains (dark gray reflectors on our interpretations)
1099 stays consistent till beneath the domain of the intermediate SDR (Fig. 7). This suggests that the
1100 continental basement is engaged during the intermediate SDR formation and the part of the Distal
1101 Necking.

1102 Considering this compositional and rheological evolution of conjugate VPM, the occurrence
1103 of a boundary between the continent and the ocean is highly unlikely and define its precise location
1104 is, for the least, highly model-driven. The composition of the basement under outer SDR and FLF
1105 stays an opened question. These extrusives could be related to a higher magmatic production than
1106 for a typical oceanic crust, above a fully magmatic crust, or to the continuation of the ductile
1107 stretching of the continental lower crust associated to a continuous magmatic dilation (e.g. W Indian
1108 VPM; Geoffroy et al., 2019, 2020).

1109 A peculiar and consistently observed, element is the Crustal Thin Point (CTP). It is always
1110 located at the end of the Distal Necking (Fig. 7). In the north (segment V and VI), it coincides with
1111 the transition from intermediate to outer SDR. South of the Cape Cross FZ, the CTP is more located
1112 either at the transition between outer SDR and FLF domains or directly within the FLF domain. In
1113 the exception of the northernmost segment, the CTP is parallel to the magnetic anomalies trend
1114 (slightly seaward of the M4 anomaly, DMHL 2 in Fig. 16), suggesting that this geological feature
1115 could be a synchronous event along the margin, bracketed between 131 and 129 Ma for the
1116 segments III to V (Fig. 17). The geodynamic model from Moulin et al. (2010; Fig. 17) demonstrates
1117 a strain acceleration, from less than 1 to 7 cm/yr full spreading rate, between 133 and 128 Ma.
1118 Between 128 and 124 Ma, the strain decreased to stabilized around 1.5 cm/yr full spreading rate.
1119 This would suggest a genetic correlation between the CTP formation and the strain evolution. The
1120 fact that the CTP line crosscuts the different SDR and FLF domains tend to demonstrate that the
1121 coeval strain acceleration (Fig. 17) would result from the complete strength release between the
1122 South Atlantic conjugate margins at the final breakup of the northern segments of the margin. This
1123 strain acceleration would therefore be inherent to the VPM formation but would have been delayed
1124 till the final rupture of the South Atlantic segment. At austral South Atlantic scale, we therefore
1125 propose to use the Crustal Thin Point as the edge of the South Atlantic conjugate VPM.

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1127

1128 **8.2. Austral South Atlantic VPM asymmetry and forcing parameters**

Using the CTP as the a priori edge of the studied VPM, their conjugate architecture appears rather symmetric between the Porto Alegre FZ and the Florianópolis FZ (Fig. 2). Conversely, from the Ventana TZ to the Porto Alegre FZ (Fig. 2), the architecture of South Atlantic conjugate VPM is characterized by a significantly narrower margin along the American plate than along the African plate. The observed asymmetry is moderate, reaching an average value of 60 % for the margin width and the SDR total volume and distribution (Table I). Some aborted rift zones are preserved on the Namibia/South Africa wide margin only (Figs. 13b, 16; Norcliffe et al., 2018). Outside these failed volcanic rifts, the extensional structures show an opposite dip on the two conjugate margins (Clerc et al., 2018). More generally, our observations show that the tectono-magmatic development of these conjugate VPM do not operate on a single and major fault. The upper crustal faults are seen decoupled on the ductile lower crust, which itself, underwent a distributed shear.

South of the Ventana TZ, the SDR-types distribution and their global extension suggest a possible inversion in the polarity of the asymmetry. Some DVS are found on the wide American margin side (Fig. 13b). These structures were not observed at crustal scale due to the lack of deep seismic profiles in this area and are thus not included in our discussion.

The general asymmetry of the South Atlantic VPM has already been highlighted in previous studies (Mohriak et al., 2002; Mohriak & Fainstein, 2012; Blaich et al., 2009, 2011 and 2013). Becker et al. (2014) also noted that the HVLC is much more voluminous and long beneath the African margin than beneath the American plate. Becker et al. (2016) also observe that SDR cover a wider area on the African margin. More recently, McMaster et al. (2019) observed that the Florianópolis Dyke Swarm on the Brazilian side is restricted to a 50 km wide coastal strip, while the dykes along Namibia are found up to 250 km away inland.

The development of asymmetric VPM is an old issue that was also evidenced in NE Atlantic. The conjugate profiles of the SE Greenland and Hatton Bank display a similar level of asymmetry comparatively to South Atlantic VPM, with a coarsely 1:2 to 1:3 ratio in SDR width and total volume of extrusive (Hopper et al., 2003; Smith et al., 2005; White and Smith, 2009). White and Smith (2009) consider that the NE Atlantic asymmetry would be most likely caused by the initial continental stretching. They proposed that the asymmetric stretching of the continental crust is inherited from previous stretching events (Permo-Triassic, Late Jurassic-Early Cretaceous) prior to the final event in the Upper Cretaceous-Paleocene. Alternatively, Hopper et al. (2003) suggested a more autogenic/magmatic phenomenon inspired from the continuous oceanic ridge migration process proposed for asymmetric seafloor spreading. This mechanism was recently inferred by Reuber et al. (2019) in the austral South Atlantic.

A significant pre-magmatic thinning cannot be proposed for the asymmetric development of Austral South Atlantic VPM. Even if some Mesozoic extensional events are known in the south of the austral South Atlantic, they develop oblique rifts that have not thinned the lithosphere in the trend of the margin (Fig. 3b; Frizon de Lamotte et al., 2015; Lovecchio et al., 2020). The last documented and significant thermal event is the collapse of the Pan-African belts in the Ordovician (Helibron et al., 2008) evidenced by the emplacement of granites and associated exhumation.

Despite of the absence of major rifting events prior to the Lower Cretaceous one, it is admitted that crustal or lithospheric-scale compositional heterogeneity are present within ancient orogenic belts, where different paleogeographical units have been accreted together. However, the works of Frimmel et al. (2013), Will et al. (2014) and Will & Frimmel (2018) suggest that the opening of the South Atlantic occurred along the same paleogeographic unit corresponding to the axis of the Neoproterozoic Marmora back-arc basin (purple units in Fig. 3a). In our study, we observed that the tectono-magmatic asymmetry is still significant when the rift crosscuts the northern Gondwanide front and the related Paleozoic fold belts (Paton et al., 2016; Fig. 3). It is thus unlikely that crustal-scale heterogeneities would be continuous across two diachronous and perpendicular orogenic belts and therefore control the global asymmetry of the conjugate margins.

Several recent works in thermomechanical modeling (Brune et al., 2014 and 2016a; Theunissen & Huisman, 2019) demonstrated that a weak behavior of the continental crust might favored, to a certain point, an asymmetric development of passive margins independently of the presence of inheritance. Brune et al. (2014 and 2016a) evidenced that the asymmetry in the central South Atlantic passive margins is mainly promoted by the process of lateral rift migration (Ranero & Perez-Gussinye, 2010) and that the width of the wider margin increases linearly with rift velocity (i.e., southward). This process requires the formation of a low-viscosity pocket at the rift side. Also, in their models, extremely weak crusts (i.e. with a thick ductile layer) tend to develop extremely wide and symmetric conjugate margins because of the absence of any localization of the crustal thinning.

The observations made in this paper, such as the symmetrical sheared lower crust, the large, flat-lying, detachment faults and the inability to locate the deformation (as evidenced by the westward rift axis shift), tend to support the globally weak behavior of the crust during the breakup of the Austral South Atlantic. This raises the question of what may control the weakening of the crust and the general localization of potential low-viscosity pocket toward the South American side (narrow margin). We recognize an architecture that may fit the extreme weak rheological configuration of the Brune et al.'s models (2016a) for the northernmost segment (VI in Fig. 2), characterized by the development of symmetric and very wide conjugate VPM. However, along the

1196 2500 km long margins of the austral South Atlantic, the studied profiles do not exhibit significant
1197 variations in width. Thus, we do not observe the correlation that Brune et al.'s (2014) observe
1198 between the plates velocities and the margin width. Moreover, lateral migration of active rift axis
1199 was observed and modelled in the context of magma-poor margins. This process does not involve
1200 the input of magma nor the melt migration through the lithosphere. This difficult task is often
1201 neglected or not properly modelled in passive margin simulation, whereas, in the context of VPM, it
1202 is very likely a dominant control factor which influences the mode of deformation and the rheology
1203 of the rift (Yamasaki and Gernigon, 2009, 2010).

1204 Our interpretation suggests that the architecture of South Atlantic VPM is acquired during
1205 their development and would thus be mainly caused by an asymmetric stretching of the continental
1206 crust controlling the asymmetric repartition of the extrusives. The observed failed rifts (DVS in
1207 Figs. 7, 13b & 16) and the related westward rift jump have accompanied the enlargement the
1208 deformed zone on the African side at least, within the segments III & IV (Fig. 2). This hesitation in
1209 the main rift localization ends with the development of a conjugate, coarsely symmetric, thick and
1210 long intermediate SDR.

1211 We propose that the westward dipping inherited crustal orogenic fabric could have helped to
1212 promote the development of primary west dipping half graben observed in the Proximal Domain of
1213 both conjugate margins (Fig. 7). This structural reactivation must be limited to the Precambrian
1214 basement domain and is highly speculative south of the Gondwanide orogenic front.

1215 Then, it is inferred that the general and early asymmetric development of the Atlantic rift is
1216 essentially the result of the globally weaker rheological behavior of the South African lithosphere
1217 due to an more intense thinning of the lithosphere under the Africa, resulting in an asymmetrical
1218 thermal regime. This would have promoted the higher magma production under the Africa and the
1219 asymmetric distribution observed of both the HVLC (Becker et al., 2014) and dyke swarms
1220 (McMaster et al., 2019).

1221 A recent geochemical study suggests that the Parana-Etendeka traps related melts must have
1222 been generated beneath a highly attenuated lithosphere of about 50km in thickness (Jennings et al.,
1223 2019). This implies that the LAB topography could have been highly instable and very shallow
1224 since the earliest Cretaceous on the American side in the northernmost part of the Austral South
1225 Atlantic. A more efficient thermal erosion and the shift of the magmatic activity toward the present-
1226 day location of the breakup axis might be favored by compositional or structural heterogeneities is
1227 the lithospheric mantle inherited from either or both the Pan-African orogeny north of the
1228 Gondwanide front and the Karoo-Ferrar LIP in the southernmost segments. This mechanism has

1229 been proposed to explain geochemical and lava emplacement heterogeneities of the CAMP over
1230 Florida, West Africa and northern South America (Nomade et al., 2002).

1231 In summary, the early stages of crustal deformation and lithospheric mantle erosion might be
1232 driven by the Pan-African inheritance. The late westward rift jump that characterizes the
1233 asymmetrical development of the austral South Atlantic might derive from the progressive
1234 enlargement of the thinning area of the lithospheric mantle beneath the rift axis and from a
1235 westward shift of the magma pathway throughout the upper mantle (e.g., Mittelstaedt et al., 2008 &
1236 2011; Yamasaki & Gernigon, 2010). In one hand, the general weakening of the crust, consequence
1237 of a strong thermal regime associated to the early intense lithospheric mantle erosion, might be
1238 enough to explain the autogenic development of the asymmetry as suggested by the
1239 thermomechanical models (Brune et al., 2016a). In another hand, in the highly magmatic systems,
1240 the distribution of the deformation is thought to be strongly influenced by the distribution of the
1241 magma (Corti et al., 2002, 2004, Yamasaki and Gernigon, 2009).

1242 The migration of the magmatic activity can promote a significant weakening of the
1243 lithosphere away from the previous rift axis and induce a delocalization of the deformation toward
1244 the future break-up axis (Yamasaki and Gernigon, 2010). Also, the combination of thermal
1245 relaxation of the thinned crust and associated volume of magma may also conduct to a rheological
1246 strengthening of the early rift system delocalizing the deformation to another zone of weakness
1247 (VanWijk & Cloetingh, 2002; Yamasaki and Gernigon, 2009).

1248 Of course, these two mechanisms, could have played all together to promote a rift jump. In
1249 another hand, the development of multiple rift axes could be considered as an autogenic evolution
1250 of the highly thinned and hot lithosphere. But in this case, what would have controlled their
1251 coalescence and consistent westward migration?

1252 Obviously, a relative motion of the lithospheric plates above the hot asthenosphere would
1253 represent a seductive explanation. Unfortunately, we did not have found consistent kinematic
1254 models that would be suitable for the observed rift jump. Moreover, the published models
1255 concerning the location of a hypothetical South Atlantic plume or hot spot are often contradictory
1256 (compare Fromm et al., 2017 and Reuber et al., 2019 for instance).

1257 When considering that the location of the magmatic activity is constrained by the area of
1258 significant thinning of the lithosphere, there is no peculiar reasons to generate a sudden and
1259 consistent westward offset of the magma pathway from an in-development initial zone of rifting.

1260 Therefore, it is likely that the assumed very wide thinning zone of the lithosphere may
1261 favorize, by thermal relaxation of its eastern corner, the progressive abandon of the initial rift axis.

1262 Then the westward offset of the new deformation area would help to focus the magmatic activity at
1263 the locus of the final break-up.

1264 The asymmetry of the VPM tends to vanish north of the PAFZ (segment VI, Fig. 13). This
1265 quite symmetric architecture also occurs where the margins are the widest. This architecture
1266 coincides with the areas where the magmatic input in the crust appears as the highest (Figs. 4b and
1267 13a). Thus, a higher than elsewhere thermal gradient related to mantle anomaly is inferred both by
1268 the eruption of the Paraná-Etendeka CFB, the thick SDR development, the large amount of
1269 underplating and by the higher thicknesses of the oceanic crust (Line 3000 in Fig. 7, Taposeea et al.,
1270 2017). This high thermal gradient may have strongly weakened the crust, thickening its ductile part
1271 to an extreme level. This mechanism does not exclude a combination to other factors such as a
1272 chemical change of the crust or lithospheric mantle nature (magmatic arc in Fig. 3).

1273 We propose that this highly magmatic area of the segment VI and of the associated
1274 WR/RGR may have played a dominant role in the asymmetric architecture of the southern domain.
1275 The hot and very weak northern area could have played the role of mechanical barrier, preventing
1276 the initial rift propagation and promoting its progressive cooling. The resulting rift jump would
1277 have been eased toward the west, because of an already thinned lithosphere, resulting in a higher
1278 thermal gradient.

1279 Numerous questions remain to accurately deconvolute the forces involved in the 4D
1280 development of this VPM. What is the thinning amount of the lithospheric mantle when the traps
1281 and first grabens develop at the surface? The evolution of the shape of a thermal anomaly while the
1282 lithosphere is extended, is clearly mysterious. Also, is the mantle dynamic related to the South
1283 Atlantic opening dominated by one thermal anomaly which could have been channelized
1284 southward, or by an elongated upwelling zone related to a first-order convective cell (e.g. Husson et
1285 al., 2012), or again by a hotter than normal ambient upper mantle due to the Pangea related
1286 continental insulation (Coltice et al., 2007)?

1287 Moreover, despite a better imaged crustal architecture, the precise timing of the N-S
1288 propagation of the Atlantic rift suffers from the lack of absolute dating on the successive SDR
1289 wedges.

1290

1291 **9. Conclusions**

1292 The recent long offset seismic profiles of the austral South Atlantic conjugate volcanic
1293 passive margins evidence that the different geometries and internal structures of SDR are directly

1294 related to the thinning mode of the underlying crustal basement and to the co-evolution of its
1295 rheological behavior.

1296 In addition, the studied seismic profiles evidence a clear asymmetry both in the SDR
1297 distribution and crustal domains. This asymmetry is mainly bore by inner SDR and the newly
1298 defined intermediate-type SDR. The observable Moho slope inflexions and the overlying crustal
1299 thickness profiles also define a significant asymmetry characterized by a wide African margin
1300 facing to a narrow South American margin. Double verging sequences of flows also reveals that
1301 westward rift jumps have accompanied the development of a wider African margin.

1302 Using these new and extensive observations we propose an evolution scheme for the
1303 formation for the South Atlantic austral segment VPM (Fig. 18) sketched here for the segment V
1304 (Fig. 2):

1305 a) The rift emplaced over the re-equilibrated Pan-African orogenic crust (Fig. 18a). This
1306 crust is considered to be 25-30 km thick prior to the rifting (more than 300 Ma after the last thermal
1307 event) and quite heterogenous both along-dip and -strike with the presence of several inherited
1308 structural domains from the Pan-African and Gondwanide Orogeneses (magmatic arc, back-arc,
1309 retro wedge and foreland; Basei et al., 2018). The Atlantic rifting located in a dominantly W-
1310 dipping pervasive orogenic fabric, thought to be at the front of the E-dipping suture zone and the
1311 overlying magmatic arc (Fig. 3a; Will and Frimmel, 2018; Basei et al., 2018).

1312 b) Uppermost Jurassic/Lowermost Cretaceous: The old foldbelt was stretched and affected
1313 by the first normal faults (Fig. 18b). Along dip, the structural inheritance can explain the
1314 asymmetric vergence of the early and proximal grabens which are dominantly developed by
1315 negative inversion of W-dipping orogenic fabric along both conjugate margins (Figs. 5 & 6).
1316 Considering the Mid Mesozoic rifts that developed through the SW Gondwana (in Argentina,
1317 Malvinas and Outeniqua areas) the Mid/Late Jurassic stretching phase of the Atlantic rift developed
1318 more likely in reaction to far-field stresses than to active mantle upwelling (Frizon de Lamotte et
1319 al., 2015; Lovecchio et al., 2020). However, the Latest Jurassic-Early Cretaceous grabens, from a
1320 second extensional phase, are filled with continental detrital and volcanodetrital sediments. During
1321 that period, the lithospheric mantle is thinned at least by two in order to produce the melt for the
1322 large traps (Jennings et al., 2019) and dyke swarms that emplaced during the
1323 Berriasian/Valanginian. The continental crust seems to be barely thinned during that period.

1324 c) 135-133Ma (M10-M9): The crust began to thin (Proximal Necking, Fig. 18c). This
1325 thinning is recorded by brittle deformation in the upper-crust and the emplacement of the inner-type
1326 SDR. Meanwhile, the lower crust is strongly intruded. It suffered coevally a strong shear evidenced
1327 by ductile shear lenses imaged in seismic lines. The structural inheritance seems to no longer

influence the fault pattern which shows an opposite dip on the two conjugate margins. At that stage, the lithospheric thinning was preferentially located beneath the African side, participating to the global warming and consecutive weakening of the overlying lithosphere.

d) 133-132Ma (M9-M7): The intermediate-type SDR began to emplace while the continental crust was drastically thinning (Fig. 18d). During this period, the deformation and magmatism failed to focus, deserting proto-rift axes in the east. The dominant factor controlling this westward rift jump is not clearly defined. The inability to locate both deformation and magmatism could be characteristic of weak crust rifting related to a high thermal regime and likely a shallow asthenosphere. It could also be the consequence of a progressive cooling and strengthening of the initial rift resulting in a westward shift of the tectono-magmatic activity.

e) 132-131Ma (M7--M4): during the onshore emplacement of the late intermediate- and first outer-type 1 SDR (Fig. 18e), magmatism and associated deformation definitely shift westward and localized close to the South American side leading to the observed asymmetry. This moment could be considered as the break-up initiation of the conjugate margins as the magmatism finally localized at a constant ridge forming a more symmetric magmatic taper (distal necking).

f) 131-130Ma (~M4-M2): The observed strain acceleration generating the crustal thin point ends the distal neck. As the South American and African plates drifted apart, the magmatic production was balanced by plates velocity. The magmatic input slowly decreased changing the conditions of emplacement of the extrusives from outer-type 1 SDR to outer-type 2 SDR, (Fig. 18f) ultimately evolving to FLF type sequences.

g) 130-126 Ma (M2-M0): Since 130 Ma (Hauterivian-Barremian transition, Fig. 18g), a more classical oceanic crust emplaced clearly emphasizing the drifting between the two newly formed geodynamic plates.

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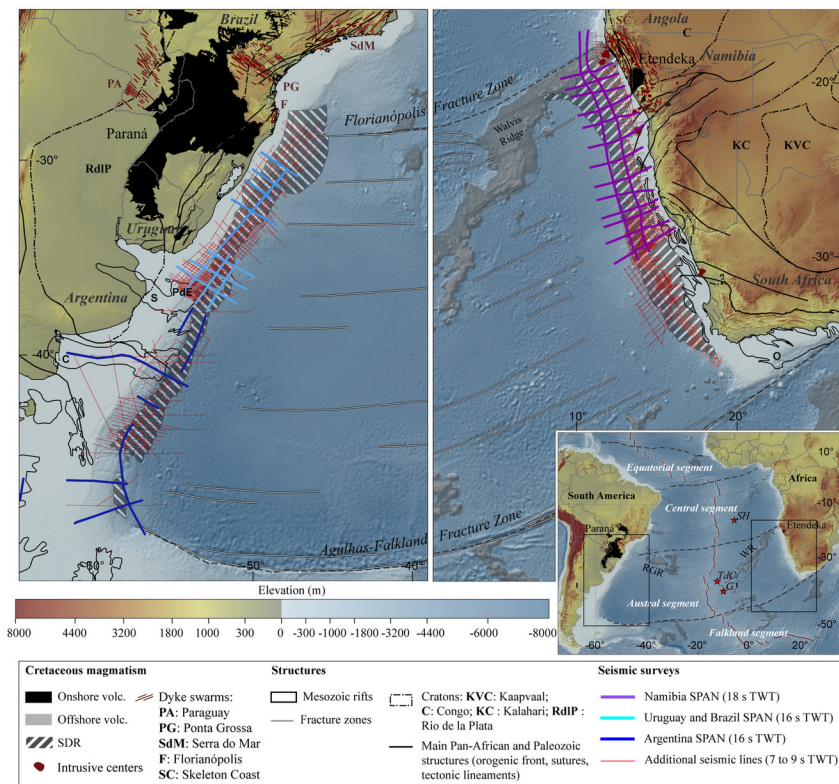


Figure 1. Map of the SDR extend (this study) with the 2D seismic surveys used to define the crustal architecture of the South Atlantic conjugate volcanic margins.

Background data: topography/bathymetry ETOPO1 grid. The South Atlantic oceanic segments and related morphostructures are reported on the inset map (from Moulin et al., 2005). WR/RGR: Walvis Ridge/Rio Grande Rise, Islands; SH: Saint Helena, TdC: Tristan da Cunha, G: Gough. The oceanic fracture zones are from Matthews et al. (2011). The dyke swarms are from Almeida et al. (2013) and McMaster et al. (2019). Mesozoic basins: C: Colorado Basin, S: Salado Basin, PdE: Punta del Este Basin, O: Outeniqua Basin. See Fig. 3 for the legend of the onshore geological background.

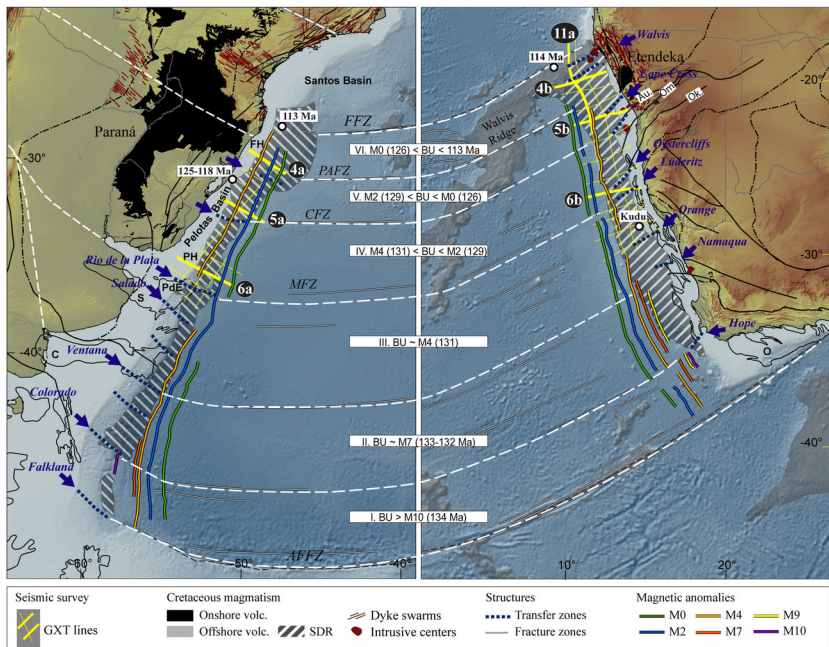


Figure 2. Map of the SDR extend (this study) with the conjugate profiles interpreted in Figs. 4, 5 & 6 (the thick yellow lines). The sub-segment ages are from magnetic anomalies (Rabinowitch & LaBrecque, 1979; Moulin et al., 2010; Koopmann et al., 2014b; Hall et al., 2018; Collier et al., 2017).

The first- and second-order along strike segmentation of the margin is figured with the major transfer zones (blue dotted lines and related arrows) recognized by Clemson et al. (1997), Franke et al. (2007), Blaich et al. (2009), Koopmann et al. (2014a), Soto et al. (2011) and Stica et al. (2014). In the Pelotas basin, they refer to the facing oceanic fracture zones (Chuí and Porto Alegre FZ in Stica et al., 2014). Microplates limits from the kinematic model of Moulin et al. (2010) are reported with onshore white dashed lines. The first-order segmentation of the South American margin is extrapolated to the conjugate plate (offshore white dashed lines), by following the oceanic fracture zones from Matthews et al. (2011). The indications for the break-up age in each first-order segment corresponds to the age (according to the timescale of Gradstein et al., 2012) of the first magnetic anomaly resting on a constant thickness crust (this study). Note the M9 and M4 anomalies overlapping the SDR. The ages in white boxes correspond to the available Ar-Ar datings obtained on Brazilian SDR sequences and on a Walvis Ridge volcano (see text for references). Fracture Zones: FFZ: Florianópolis Fracture Zone, PAFZ: Porto Alegre FZ, CFZ: Chuí FZ, MFZ: Meteor FZ, AFFZ: Agulhas-Falkland FZ. Basement highs: PH: Polonio High, FH: Florianópolis High. . The Pan-African structures reactivated during the South Atlantic opening (from Corner & Durrheim, 2018): Au.: Autsieb, Om.: Omaruru, Ok.: Okahandja. See Fig. 3 for the legend of the onshore geological background.

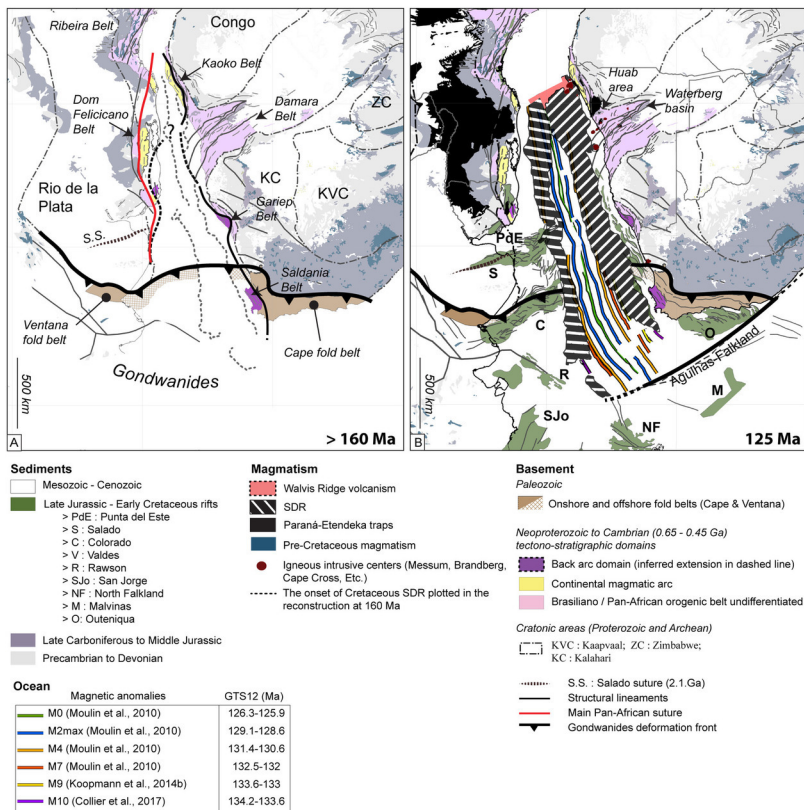


Figure 3. A: Pre-opening reconstruction of the SW Gondwana (kinematic model from Moulin et al., 2010); B: Reconstruction at 126 Ma (~M0) with the magnetic anomalies identified in the South Atlantic Ocean and the SDR domains mapped in this study. The trans-Atlantic connections between analogous tectono-stratigraphic units are from and Pangaro & Ramos (2012), Frimmel et al. (2013) and Paton et al. (2016). Mesozoic rifts are from Parsiegla et al. (2009), Becker et al. (2012), Lovecchio et al. (2018) and this study. GTS12: time scale from Gradstein et al. (2012).

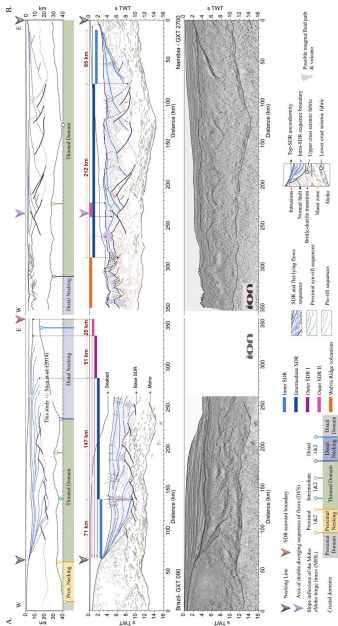


Figure 4. Conjugate 090 vs. 2700 ION GXT lines line drawing and interpretation with corresponding time-depth conversion. The conjugate lines are located in the segment VI of Fig. 2.

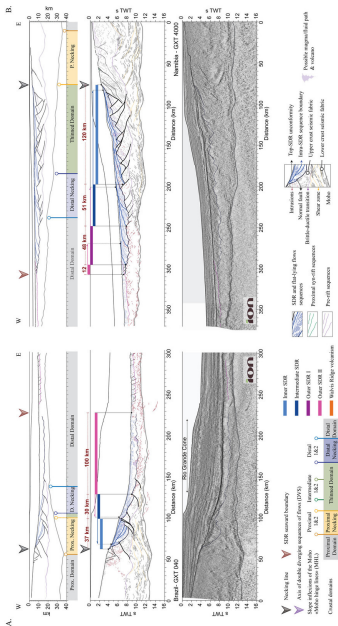


Figure 5. Conjugate 040 vs. 4000 ION GXT lines line drawing and interpretation with corresponding time-depth conversion. The conjugate lines are located in the segment V of Fig. 2.

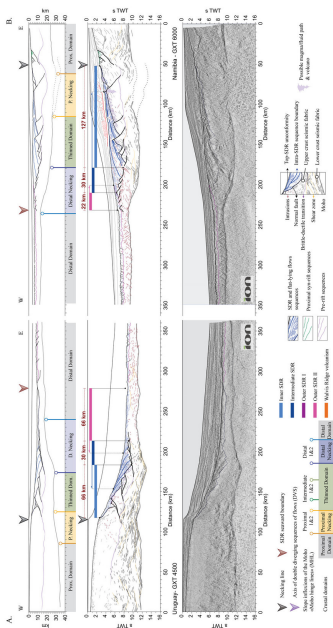


Figure 6. Conjugate 4500 vs. 6000 ION GXT lines line drawing and interpretation with corresponding time-depth conversion. The conjugate lines are located in the segment IV of Fig. 2.

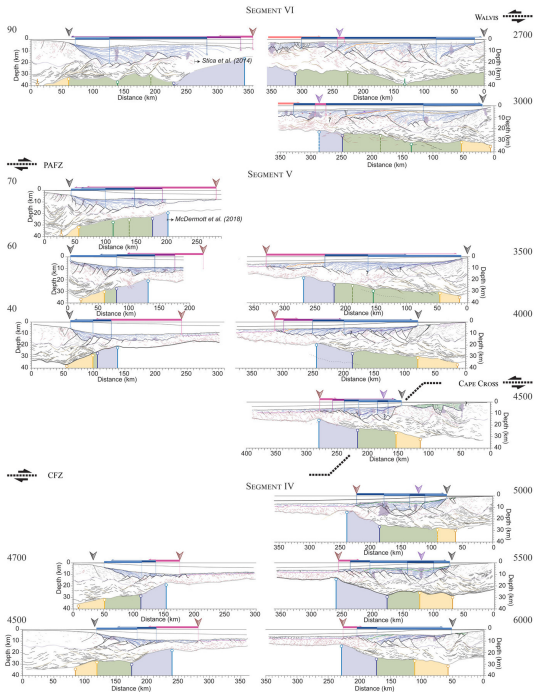


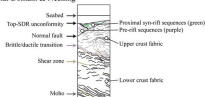
Figure 7. Interpretations of the depth-converted seismic profiles for the entire 2D ION survey (location in Fig. 2). The inset map displays the seismic survey reconstructed at 126 Ma and 133 Ma according to the Moulin et al. (2010) kinematic model. The conjugate lines displayed in Figs. 4 to 6 are underlined in yellow.

LINE DRAWING

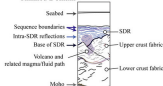
VOLCANIC ARCHITECTURE

CRUSTAL ARCHITECTURE

Proximal Domain & Necking



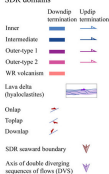
Thinned Domain



Distal Domain & Distal Necking



SDR domains



Crustal domains

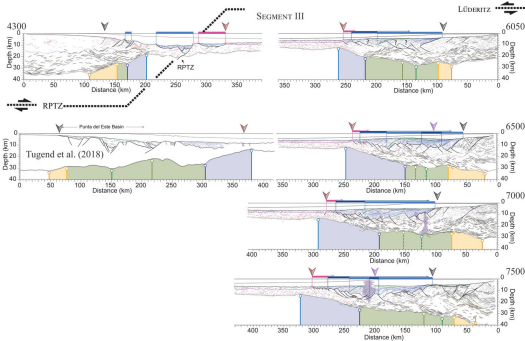
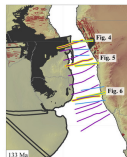
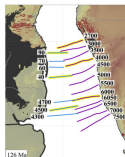


Figure 7. (continue)

DOMAINS

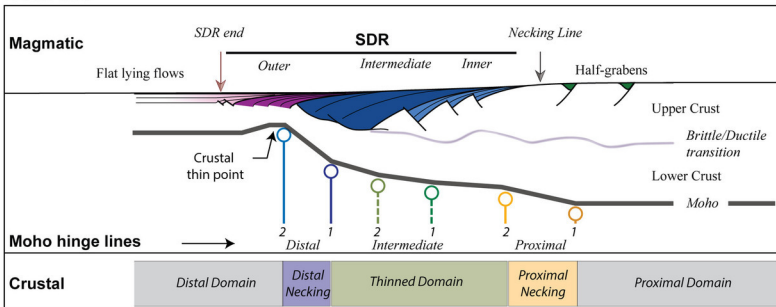


Figure 8. Synthetic scheme illustrating the distribution of the main structures and of the different magmatic and crustal domains as observed on the South Atlantic conjugate VPM (Fig. 7).

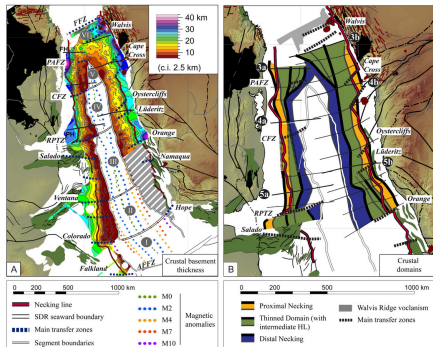


Figure 9. Maps reconstructed at 126 Ma (M0). See Fig. 3 for the legend of the geological background. A: Thickness of the crust preserved beneath the SDR. PH: Polonio High, FH: Florianópolis High. B: Location of the main crustal domains and Moho hinge lines. In A., the first-order oceanic segments corresponds to those of the displayed in Fig. 2.

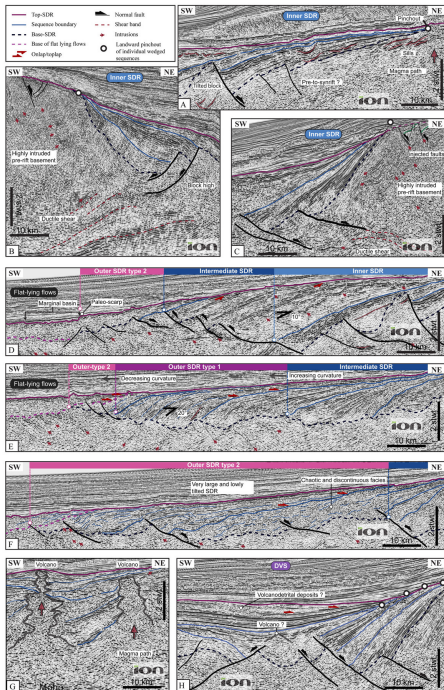


Figure 10. Illustration of the main structure types and of seismic facies encountered within the SDR domain.

A-B-C: Inner-type SDR: Different examples of inner SDR forming a planar-diverging wedge related to continent-ward dipping normal faults. Note the important lateral and vertical variations of the reflectivity within these seismic units. D: Intermediate-type SDR: Large and thick wedge with slightly arcuate-diverging reflections (up to 10° dip). Up-dip terminations display seaward migrating pinchout of successive wedges. The base of the wedge lies on an unclear surface which appears intensively intruded (red arrows) and cut up by a complex network of continentward-dipping faults. This SDR is overlain by a slightly thickening seaward-dipping sequence of planar-diverging reflectors (above the pink dotted line). This ultimate sequence of flows (outer-type 2) oversteps the hanging wall of the underlying SDR and laterally merges with a sequence of flat-lying flows. Toward the continent, the intermediate-type wedge overlies a set of shorter and over-tilted inner-type wedges (bottom right). The intermediate-type SDR develops above a 11 km thick highly intruded basement. Note the numerous figures of intrusions in the interior of the wedge (red arrows). E: Outer-types 1 and 2: The outer-type 1 is characterized by a set of strongly arcuate-diverging sequences of reflectors deflected above a sub-horizontal to slightly continent-ward dipping basal surface. This outer SDR forms a 5-6km thick wedge of seaward-dipping sequences flows. The reflections tend to onlap the previously emplaced sequences with a backstepping pattern which suggests a coeval subsidence of the inner-more domain. The last emplaced sequence of flows (outer-type 2) onlap a residual topography formed by the previous SDR sequence flexure. They overly a highly injected basement which underwent a very limited extension. F: 100 km long and 4 km thick outer-type 2 SDR showing a planar-diverging pattern with a slightly tectonized substratum and a low lateral thickening (cf. line 3500 in Fig. 7). G: Evidences for deep-seated volcanic constructions emplaced throughout the inner SDR of the Namibian margin (line 4500 in Fig. 7). Note the « vertical column » of highly disrupted seismic facies. This disturbed acoustic facies is likely due to a magma and/or fluid extensive impregnation of the basement. H: Example of double verging sequences of SDR wedges.

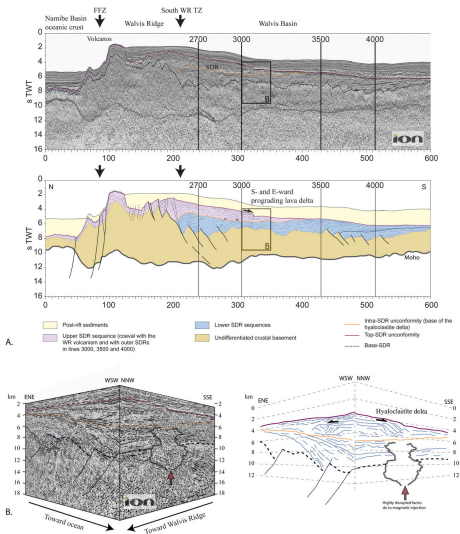
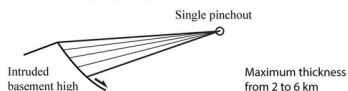


Figure 11. A: Margin parallel profile (location in Fig. 2) and corresponding schematic interpretation illustrating the crustal and volcanic architecture of the northern Namibian VPM and Walvis Ridge. Two superimposed sequences of SDR, separated by a major unconformity (orange marker, also in Fig. 7). The youngest SDR (pink) are correlatable with the sequences that lie at the top of the Walvis Ridge and which are likely intra-Aptian (see text for explanation). The two SDR sequences dip and thicken northward, against an E-W transfer zone (Walvis TZ in Fig. 2), and westward (see in-dip profiles in Figs. 4 & 5). A lava delta is interpreted southeastward, downlapping the orange marker (cf B). B: Intersecting seismic profiles showing the 3D architecture of the southeastward prograding lava delta. The subaqueous volcano-detrital construction develops against the southern flank of Walvis Ridge and within a subsiding down to 1000m depth basin. This indicates that the margin central segment (n° V, Fig. 2) south of Walvis Ridge has started its subsidence and is drowning whereas subaerial flows continue to develop SDR wedges toward the N and the W.

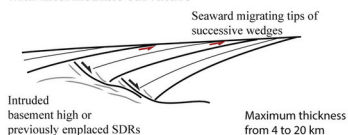
Inner SDRs

Planar-diverging wedge

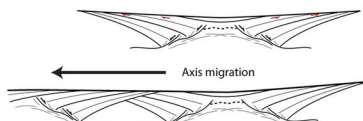


Intermediate SDRs

Arcuate-diverging thick wedge with intermediate curvature

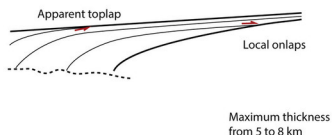


Double verging sequences (DVS) of volcanic wedges



Outer SDRs

Type 1 - Arcuate-diverging thick wedge with high curvature and sub-horizontal basal surface (maximum thickness from 5 to 8km)



Type 2 - Various patterns observed for the last emplaced SDRs

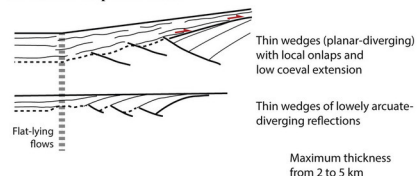


Figure 12. Typology applied to the SDR domain and based on the structural diversity of the SDR successive wedges (modified from Planke et al., 2000).

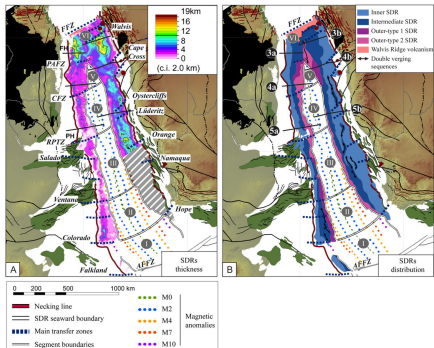


Figure 13. Maps reconstructed at 126 Ma (M0). See Fig. 3 for the legend of the geological background. A: Total SDR thickness. B: Distribution of the different types of SDR. The first-order oceanic segments corresponds to those of the displayed in Fig. 2.

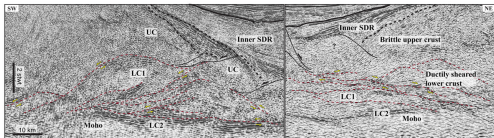
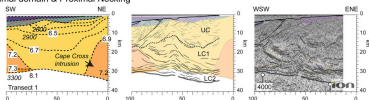
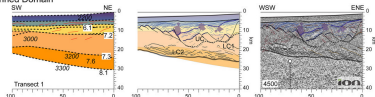


Figure 14. Conjugate view of the ductile and pervasive deformation within the lower crust. Shear bands (red dotted lines) are distributed through the reflective lower crust. The internal fabric display folds and dome-shape structures which is also diagnostic for a solid-state flow of the deep part of the VPM. Note the apparent sense of shear is consistent with continentward-dipping normal faults crosscutting the upper crust (UC) and the seaward tilt of the overlying SDR. See the text for explanation of LC1 and LC2 lower crust.

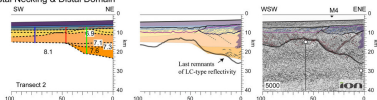
A. Proximal domain & Proximal Necking



B. Thinned Domain



C. Distal Necking & Distal Domain



Bauer et al.'s (2000) velocity model (P-waves)

— Seismic reflectors (Bauer et al.'s line drawing)

----- Isovelocity lines (with value in km.s^{-1})
with corresponding density values in g.cm^{-3} (only for Transect 1)

Velocity profiles (in C.):
— Landward profile
— Mid profile
— Seaward profile
— White et al. (1992)
Seismic velocity limits
for mature oceanic crust

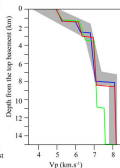


Figure 15. Velocity structure of the Namibian volcanic margin. Bauer et al. (2000) refraction profile (Transect 1 & 2, in left column) plot on top (on central column) of the closest seismic lines (4000 in A, 4500 in B and 5000 in C). A: Velocity model (Transect 1) projected on the Proximal Domain of the line 4000 (location in D). B: Velocity model (Transect 1) projected on the Thinned Domain of the line 4500 (location in D). C: Velocity model (Transect 2) projected on the Distal Necking and Distal Domain of the line 5000 (location in D). D: Map of the location of the displayed seismic (yellow lines) and refraction (thick black lines) data. The undifferentiated SDR domain is plotted in grey-whites stripes together with the magnetic anomalies as picked by Moulin et al (2010) and the Necking Line and the SDR-seaward boundary. E: Depth-Vp plot for the three profiles shown in C. and the seismic velocity limits for mature oceanic crust (grey shading) from White et al. (1992). This diagram is from Taposeea et al. (2017, redrawn).

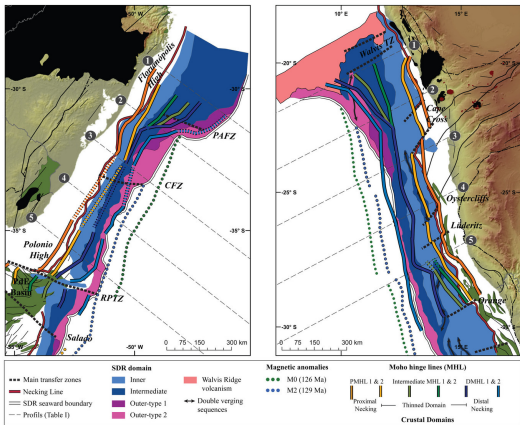


Figure 16. Location of the main crustal domains and related Moho hinge lines over the distribution of the SDR in Uruguay-SE Brazil and Namibia conjugate VPM. The conjugate profiles in dashed lines refer to the measures reported in Table I.

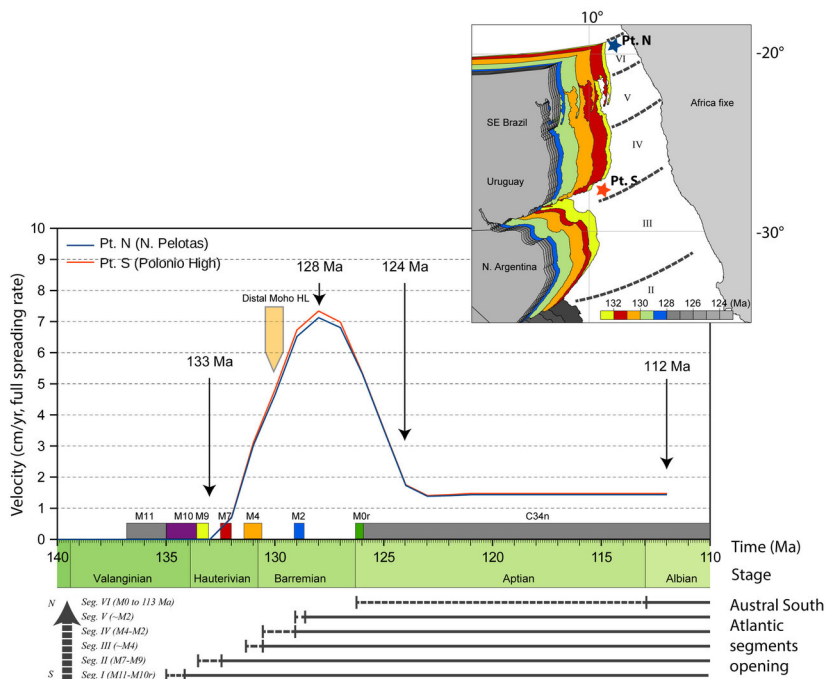


Figure 17. Full spreading rates evolution in austral South Atlantic according to Moulin et al.'s (2010) rotation poles and kinematic model. The two velocities curves are calculated for the N and S extremities of the “la Plata” microplate (Uruguay-SE Brazil) of Moulin et al. (2010) (see the inset map for location). It is noteworthy that no significant difference is observed between these two reference points. The main velocity changes are pointed with an arrow. The age indicated for the distal Moho hinge line 2 and associated crustal thin point is inferred, assessing it corresponds to an isochron located between M4 and M2 anomalies (Fig. 16). The successive openings of the six major SA segments are reported below the graph, in accordance with the data and studies quoted in Fig. 2. The dotted lines segments illustrate the uncertainties related magnetic anomalies duration. The inset map show an animation of the SW-ward drift of the South American microplates of Moulin et al. (2010), between 133 Ma and 124 Ma, relatively to a fixed Africa.

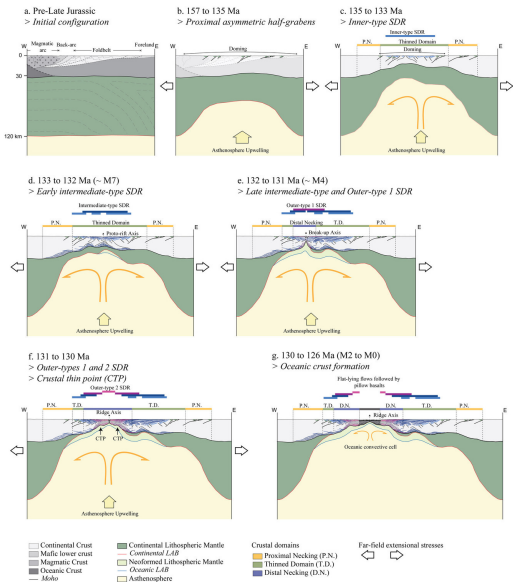


Figure 18. Conceptual evolution of South Atlantic austral segment VPM. The initial stage structural domains from Panafrican belt are redrawn from Basei et al. (2018). LAB: Lithosphere-Asthenosphere Boundary. The proposed timing is adapted for the tectono-magmatic development of the segments IV and V (Fig. 2 for location and Fig. 17 for timing constraints).

Feuille1

Profiles number (from N to S)	Margin width (km)		Inner (km)		Intermediate (km)		Outer-type 1 (km)		Outer-type 2 (km)	
	Ur-Br	Na	Ur-Br	Na	Ur-Br	Na	Ur-Br	Na	Ur-Br	Na
1	300	310	51	100	180	140	40	-	20	99
2	123	244	44	168	60	60	43	30	25	60
3	84	150	32	72	38	27	-	34	88	-
4	97	193	42	136	40	40	-	-	25	9
5	131	172	55	100	41	49	-	-	63	20
Average ± S.D. (all)	147 ± 88	213 ± 64	45 ± 9	115 ± 37	72 ± 61	63 ± 45	42 ± 2	32 ± 3	44.2 ± 30	47 ± 41
Average ± S.D. (2 to 5)	109 ± 22	190 ± 40	43 ± 9	119 ± 42	45 ± 10	44 ± 14	43 ± 9	32 ± 3	50 ± 31	30 ± 27

Table I. Comparison of conjugate profiles based on the criteria of the width of the margin (column 1) and of the width of the different SDR domains (columns 2 to 5). See Fig. 16 for location of c

each profile.