

Postrift stress field inversion in the Potiguar Basin, Brazil – Implications for petroleum systems and evolution of the equatorial margin of South America

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4 1 **Postrift stress field inversion in the Potiguar Basin, Brazil – implications for**
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6 2 **petroleum systems and evolution of the equatorial margin of South America**

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24 **Abstract**

25 Rifting and related normal stress regime in the equatorial continental margin of Brazil
26 ceased during the Late Cretaceous, when the stress regimes in eastern South America and
27 West Africa changed to induce strike-slip or reverse motion. In this study, we explore the
28 postrift tectonic, geomorphic, magmatic, and sedimentary responses to stress changes
29 using the Potiguar Basin, the easternmost basin in the equatorial margin of Brazil, as a
30 case study. We use field and topographic data, 2D seismic reflection lines, vertical
31 electric soundings, and geochronological and borehole data to constrain the stress
32 evolution of the Potiguar Basin from the Late Cretaceous to the Quaternary, discussing
33 the role of basin inversion on sedimentation and landforms. Our results indicate the
34 presence of two strike-slip stress regimes after rifting. The first stress field (SF1)
35 occurred from Late Cretaceous to the middle Miocene and consisted of a N-S-oriented
36 maximum subhorizontal compression and an E-W-oriented extension. The second stress
37 field (SF2) took place from the middle Miocene to the present day and included
38 subhorizontal E-W to NW-SE compression combined with N-S and NE-SW
39 subhorizontal extension. Emplacement of volcanic rocks occurred along transtensional
40 faults, with a principal peak during SF1 at 20-30 Ma and a subordinate peak during SF2
41 at 5-10 Ma. In response to shortening during SF2, a 70-km-long and 50-km-wide dome
42 formed, where marine Miocene strata were uplifted to ~250 m asl. This uplift induced the
43 displacement of alluvial channels away from the dome. Anticlines formed by
44 transpression along the main NE-SW-striking faults during both SF1 and SF2 acted as
45 traps in the petroleum system. Similar shifts and stress field inversions documented in
46 other areas of the Brazilian continental margin are consistent with the Neogene rise of the

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4 47 Andes and may have implications for reconstructing the tectonic history of the Equatorial
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6 48 Atlantic margin of South America.
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11 50 **Keywords:** Neotectonics, basin inversion, stress field, equatorial margin, tectonic uplift
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15 52 **1. Introduction**

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19 53 The postrift evolution of continental margins is, in theory, characterized by thermal
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21 54 subsidence related to lithospheric cooling and associated thickening ([McKenzie, 1978](#)).
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24 55 The postrift stage can evolve into a passive margin, which was initially viewed as being
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26 56 marked by tectonic quiescence. However, many passive margins worldwide have been
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29 57 subjected to inversion, such as in eastern Canada ([Withjack et al., 1995](#)), the North Sea
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31 58 ([Doré et al., 2008](#)), the western Ghats in India ([Gunnell and Fleitout, 2000](#)), Rockall
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33 59 Trough off Great Britain ([Boldreed and Anderson, 1993](#)), southeastern Brazil ([Cobbold et](#)
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35 60 [al., 2001](#)), and West Africa ([Hudec and Jackson, 2002](#)).
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41 62 The processes that drive the postrift evolution of a passive margin remains elusive, and
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43 63 two main points are still open to debate. First, the first-order stress fields on most present-
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46 64 day continental margins are well established, but they are not necessarily the same as
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49 65 those that existed during the early postrift stage, which are still poorly documented (e.g.,
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51 66 [Dyksterhuis et al., 2005](#); [Sippel et al., 2009](#)). Second, there are several examples of rift-
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53 67 related basin inversion on passive margins (e.g., [Müller et al., 2014](#)), but geophysical data
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56 68 (e.g., seismic reflection sections) have been unable to unravel most postrift structures,
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58 69 which are either very shallow or below seismic resolution.
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6 71 Most basins on the conjugate margins of eastern South America and West Africa are
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9 72 considered to be examples of the ‘passive’ extensional basin type described by [McKenzie](#)
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11 73 ([1978](#)). On the equatorial margins of these continents, rifting has evolved from wrench
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14 74 tectonics in the Late Jurassic to the Early Cretaceous to a passive margin stage that began
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16 75 in the Late Cretaceous (e.g., [Moulin et al., 2010](#)). After continental rifting, these basins
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19 76 underwent tectonic quiescence, during which the tectonics were presumably controlled by
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21 77 thermal subsidence ([Matos, 1992](#); [Maurin and Guiraud, 1993](#)). However, flexural effects
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24 78 on passive margins offset by oceanic fracture zones might not follow the classic thermal
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26 79 subsidence model described by [McKenzie \(1978\)](#), and later critically assessed by [Braun](#)
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28 80 [et al. \(2013\)](#).

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33 82 At the basin scale, changes in plate motions on the equatorial margins of South America
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36 83 and Africa have changed the stress field and affected the fault geometries and kinematics
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38 84 as much as the resulting basin stratigraphy ([Fairhead and Binks, 1991](#)). Nonetheless,
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41 85 many studies have failed to explain the evolution of these conjugate margins after
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43 86 continental breakup. Despite mounting evidence of basin inversion ([Cobbold et al., 2001](#);
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45 87 [Marques et al., 2014](#); [Nogueira et al., 2015](#)) as well as folding and uplifting of Neogene
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48 88 and Quaternary coastal sediments (e.g., [Rossetti et al., 2013](#); [Gandini et al., 2014](#)), linking
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51 89 the normal faulting stress regime during the breakup of Pangea to the present-day
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53 90 intraplate strike-slip to reverse stress regime is still a challenge. In addition, several
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55 91 consequences of this stress field inversion such as topographic and sedimentary changes
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58 92 have not been fully assessed.

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94 The paleostress evolution of the equatorial margin is not well known, especially after
95 rifting ceased in the Cretaceous. Predicting stress fields along sheared continental
96 margins is difficult because little is known about the mechanical interaction between the
97 two plates. For example, the present-day intraplate stress field in South America is
98 controlled by its collision with the Nazca Plate and the rise of the Andes ([Garcione et al.,](#)
99 [2008](#); [Marques et al., 2014](#); [Nogueira et al., 2015](#); [Assumpção et al., 2016](#)). However, to
100 what extent the spatial and temporal variations of the upper-crustal deformation in the
101 central Andes are related to shifts in the intraplate stress field of South America remains
102 an unsolved question. Therefore, understanding the controlling factors of stress evolution
103 during the postrift period is important for unravelling the geological evolution of the
104 Atlantic margin.

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106 This work aims to understand the evolution of the stress fields between the normal fault
107 stress regime during the breakup of Pangea and the present-day intraplate strike-slip
108 stress regime in equatorial South America. The main objectives of this study are (1) to
109 determine the evolution of the stress field during the postrift period; (2) to identify the
110 timing of tectonic inversion; (3) to investigate the effects of inversion on Neogene-
111 Quaternary sedimentation and landforms; (4) to investigate the implications of stress field
112 inversion in the petroleum geology system of the Potiguar Basin; and (5) to assess the
113 importance of the Potiguar Basin to better understanding the evolution of Equatorial
114 Brazil's offshore region.

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4 116 We selected the Potiguar Basin in the equatorial margin of Brazil as a case study for
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6 117 several reasons. First, it is composed of a long stratigraphic record comprising
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8 118 sedimentary units that span from the Neocomian (~138 Ma) to the Late Quaternary in an
9
10 119 aborted rift located on the passive continental margin of northeastern Brazil (e.g., [Matos,](#)
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12 120 [1992; de Castro et al., 2012](#)). Second, it is located in a region that was the last segment of
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14 121 the South American Plate to breakup from the African Plate ([Azevedo, 1991; Matos,](#)
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16 122 [1992](#)). A significant portion of the seismicity in South America is also concentrated in
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18 123 this region, where the highest ground peak acceleration in intraplate South America is
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20 124 recorded ([Shedlock and Tanner, 1999](#)). Finally, it contains a variety of present-day stress
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22 125 markers such as focal mechanisms, breakouts, and image logs (e.g., [Ferreira et al., 2008;](#)
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24 126 [Bezerra et al., 2011; Reis et al., 2013](#)), so it is a good area to correlate the paleo- and
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26 127 present-day stress fields ([Fig. 1](#)). The multidisciplinary investigation carried out in the
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28 128 Potiguar Basin allowed us to better understand the postrift evolution of Equatorial
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30 129 Brazil's offshore region, improving the current knowledge about the opening of the
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32 130 Atlantic and related plate reconstructions.
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132 **2. Tectonic and stratigraphic settings**

133 *2.1. Rift-to-drift evolution of the Potiguar Basin*

134 The Potiguar Basin is located in the eastern part of the Brazilian equatorial margin. This
135 basin is an aborted rift located on the narrowest continental margin of Brazil ([Fig. 1A](#)),
136 and is limited to the south and west by Archean and Proterozoic crystalline basement,
137 respectively ([Van Schmus et al., 1995; Souza et al., 2013, 2016](#)). The Potiguar Basin is
138 composed of a main NE-SW-oriented central graben that is ~6 km deep and ~50 km
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4 139 wide, bounded by two horsts (Fig. 2). The central graben formed at 130 Ma, when ductile
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6 140 Precambrian shear zones were reactivated during continental rifting (Matos, 1992; De
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9 141 Castro et al., 2012). Rifting evolved from NW-SE-oriented extension in the Neocomian
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11 142 to E-W-oriented extension in the Barremian (Matos, 1992). Most of the rift faults in the
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14 143 Potiguar Basin strike NE-SW, whereas the transfer faults that accommodate
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16 144 displacements between different rift segments strike NW-SE (Bertani et al., 1990; De
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19 145 Castro and Bezerra, 2015). Postrift reactivation of these faults propagated them into Late
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21 146 Cretaceous to Cenozoic strata, although with lower slip rates (Bezerra and Vita-Finzi,
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23 147 2000; Nogueira et al., 2010).

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28 149 The sedimentary filling of the Potiguar Basin consists of three major depositional
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30 150 sequences: rift, transitional, and postrift. The first two sequences are buried and cannot be
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32 151 directly investigated in the field (Fig. 2). However, the postrift sequence crops out in the
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34 152 basin and contains stratigraphic markers that can be used to constrain the ages of the
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36 153 paleostress fields.

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43 155 The lithologies, locations, and ages of four postrift sedimentary and a volcanic unit are
44
45 156 well constrained in the Potiguar Basin. The Açu and Jandaíra Formations record the first
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47 157 phase of the opening of the South Atlantic, which occurred in the Late Cretaceous. The
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49 158 Açu Formation contains siliciclastic alluvial deposits (Fig. 2) deposited from the Albian
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51 159 (110 Ma) to the Cenomanian (94 Ma), whereas the Jandaíra Formation records carbonates
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53 160 that formed in tidal to shallow shelf environments from the Turonian (93 Ma) to the
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55 161 Campanian (80 Ma) (Pessoa Neto et al., 2007). These units cap transitional and rift
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4 162 sequences in the central graben but lie directly over the crystalline basement in the
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6 163 eastern and western horsts (Fig. 2) (Sousa Filho et al., 2000). Soares et al. (2012) and
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9 164 Alves et al. (2017) have shown these “transitional” sequences to be associated with
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11 165 continental breakup.
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16 167 The Barreiras Formation caps the Late Cretaceous units and consists of several
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18 168 siliciclastic lithologies deposited in various marine-influenced transitional environments
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21 169 (Rossetti et al., 2013). This unit is present in several sedimentary basins along the
22
23 170 Brazilian coast and is particularly well exposed in numerous coastal cliffs up to ~40 m
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25
26 171 high. The age of the Barreiras Formation is based on its stratigraphic relationships with
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28 172 other Oligo-Miocene sedimentary units in the western part of the Brazilian equatorial
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31 173 margin (e.g., Rossetti, 2004). In addition, pollen data are consistent with an early-middle
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33 174 Miocene age for this unit (e.g., Arai et al., 1988; Arai, 1997). This age was confirmed by
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36 175 Ar/Ar and U/Th-He chronology, which constrained the depositional ages to between 23
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38 176 Ma and 17 Ma (Lima, 2008; Rossetti et al., 2013).
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43 178 The Quaternary sediments are mostly composed of sandy to gravelly alluvial deposits
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45 179 with optically stimulated luminescence ages ranging from 400 ka to 0.4 ka (Moura-Lima
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48 180 et al., 2011). These deposits occur as alluvial sediments along recent valleys and
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51 181 paleovalleys (Moura-Lima et al., 2011) and as colluvial sediments along fault scarps
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53 182 (Gurgel et al., 2013).
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4 184 The Cenozoic volcanic and hypabyssal rocks are related to the Macau volcanism (Fig. 2).
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6 185 They cut across the Late Cretaceous units, the Barreiras Formation, and the crystalline
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9 186 basement. In the present study, we focused on the magmatic units that exhibit wide
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11 187 expression in the Potiguar Basin, which includes both alkaline and tholeiitic basalts (Sial,
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13 188 1976; Almeida et al., 1988; Souza et al., 2003, 2005) that have K-Ar and ^{40}Ar - ^{39}Ar ages
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15 189 ranging from 47 to 7.4 Ma with a prominent peak at 30-10 Ma and minor peaks at 50-40,
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17 190 40-30 and 10-0 Ma (Tab. 1 and Fig. 3). There are also volcanic bodies in the Potiguar
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19 191 Basin with older ages at 93 Ma and 70-50 Ma but these are limited to small volcanic
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21 192 bodies (Souza et al., 2004), lacking a wider expression proving they influenced the
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23 193 deformation pattern of the sedimentary basin. Based on the chemistry of inclusions in
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25 194 these units, Sial (1976, 1977) and Rivalenti et al. (2000) estimated that these basaltic
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27 195 magmas originated at pressures of 19 to 27 kbar or depths of approximately 60-90 km.
28
29 196 There are several interpretations for the origin of these volcanic rocks, including: (1) the
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31 197 surface track of the Fernando de Noronha plume (Fodor et al., 1998; Rivalenti et al.,
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33 198 2000); (2) the influence of an oceanic fracture zone (Almeida et al., 1988); (3) the result
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35 199 of either an internal readjustment within the South American plate during its westward
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37 200 migration, or the pressure release of arched zones that formed in the Cretaceous during
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39 201 the opening of the South Atlantic Ocean (Sial, 1976; Carneiro et al., 1988); or (4) the
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41 202 manifestation of the upwelling flow in an edge-driven convection mode (Knesel et al.,
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43 203 2011). Despite its controversial origin, these workers agree that the magma from these
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45 204 intrusions ascended from the uppermost portion of the upper mantle to the very shallow
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47 205 continental crust along deep fault systems.
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2.2. Present-day stress field and seismicity of the eastern equatorial margin

The present-day stress field in the Potiguar Basin and its crystalline basement is well constrained by three types of stress indicators: focal mechanisms, borehole breakouts, and image logs. In general, strike-slip faulting predominates, but normal and reverse faulting also occur. The focal mechanisms of events at depths between 1 and 12 km indicate that the maximum horizontal compression (S_{Hmax}) trends E-W in the eastern part of the basin, shifts to NW-SE in the central and western parts of the basin, and then shifts again to E-W between the Potiguar and Paraiba basins (e.g., [Ferreira et al., 1998, 2008](#), [Bezerra et al., 2011](#)) ([Fig. 1B](#)). The inversion of focal mechanisms in the eastern and western parts of the region indicates a strike-slip faulting regime with the maximum compressive stress (σ_1) subhorizontal and parallel to the coastline, the minimum stress (σ_3) orthogonal to the coastline, and a subvertical intermediate compressive stress (σ_2). The stress axis σ_1 shift to ENE-WSW-oriented and σ_3 NNE-SSW-oriented away from the Potiguar Basin in the western and southern parts of the study area ([Lima Neto et al., 2014](#); [Oliveira et al., 2015](#)) ([Fig. 1B](#)). The borehole breakout orientations are consistent with those observed in the focal mechanisms. In the central part of the Potiguar Basin, breakouts indicate a NW-SE-oriented S_{Hmax} , although a scattered pattern occurs offshore in the continental platform ([Lima et al., 1997](#)) ([Fig. 1B](#)). The image logs are also consistent with both stress indicators described above. However, the image logs indicate that a normal stress regime is present in the central part of the basin at depths of 0-2.5 km. These stresses change to a strike-slip/normal regime below 2.5 km, which is consistent with the focal mechanisms recorded in the crystalline basement from 1 to 9 km. This stress decoupling has been considered to be an indication of inversion and bending of the upper layers in the

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4 230 Potiguar Basin (Reis et al., 2013). All of these stress indicators consistently indicate a
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6 231 maximum horizontal compression roughly parallel to the coastline, which has been
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9 232 interpreted as the effect of the density contrast between the continental and oceanic crust
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11 233 (Assumpção, 1992; Ferreira et al., 1998; Oliveira et al., 2015).
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16 235 Strain data from a permanent GPS network have also been reported for the Potiguar
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18 236 Basin and adjacent regions (Marotta et al., 2015). The velocity direction trends
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20 237 predominantly west and north with maximum rates of 4.0 ± 1.5 mm/year and 4.1 ± 0.5
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22 238 mm/year for the x and y components, respectively. The largest contractional strain was -
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24 239 $0.072838 \times 10^{-6} \pm 2.32 \times 10^{-10}$ /year and was oriented NW-SE in the Potiguar Basin, and
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26 240 the maximum extensional strain was $0.109552 \times 10^{-6} \pm 3.65 \times 10^{-10}$ /year and was oriented
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28 241 NE-SW, also in the Potiguar Basin. These results indicate that the greatest strain and
29
30 242 variation in velocity are located in the crystalline basement near the Potiguar Basin and
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32 243 also that the shortening/compression direction is consistent with the maximum horizontal
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34 244 stress indicated by the three stress markers presented above (Marotta et al., 2015).
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38 246 Historical and instrumental seismicity data show that the border of the Potiguar Basin is
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40 247 seismically active. These data also indicate that this basin exhibits a relatively high level
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42 248 of seismicity compared to other 'passive' margin basins in eastern South America. The
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44 249 Potiguar Basin has been affected by sequences of earthquakes that can last 10 years or
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46 250 more and have body-wave (m_b) magnitudes up to 5.2 (Bezerra et al., 2007; Assumpção et
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48 251 al., 2014). The seismicity and the location where the stress has been released affect the
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50 252 crystalline basement around and underneath the Potiguar Basin at depths from 1 to 12 km
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253 (Takeya et al., 1989; Ferreira et al., 1998; do Nascimento et al., 2004; Lima Neto et al.,
254 2013, 2014; Oliveira et al., 2015).

255

256 **3. Methods**

257 *3.1. Paleostress analysis*

258 Paleostress tensors were reconstructed based on a detailed field study and the analysis of
259 fault populations. The field investigation included 271 outcrops that were grouped into 30
260 sites, where slip data for 1127 faults were measured. In a few locations, joint-stylolite
261 geometries and their cross-cutting relationships were analyzed using drone images with 1
262 cm pixel sizes to constrain the fault-slip data. The fault-slip data consisted of the
263 orientations of fault planes and related striations, slip senses, and the quality of the
264 kinematic indicators. A variety of kinematic indicators were used to determine the slip
265 senses of faults, including mineral steps, tension gashes, Riedel shears, conjugate shear
266 joints, and overgrowths of differently oriented fibers on the same fault (Hancock, 1985;
267 Petit, 1987; Angelier, 1994). In addition, stylolites were utilized as paleostress indicators
268 when their occurrence allowed a statistical analysis.

269

270 The following criteria guided the investigation of the relative chronology of the stress
271 markers in the field: (a) cross-cutting relationships between faults and joints; (b)
272 overprinting relationships between striae on a fault plane; (c) the ages of the stratigraphic
273 units affected by faults and joints; (d) the persistence and consistency of kinematic
274 indicators along a fault plane; and (e) splitting of inhomogeneous fault-slip data from
275 kinematically compatible sets. We applied all of these criteria to identify successive

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4 276 tectonic events and their related stress fields. Where sparse data, we analyzed outcrops
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6 277 separately but merged the data from neighboring outcrops in similar stratigraphic units
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9 278 that were not separated by fault zones. We then determined the stress tensors by
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11 279 combining the stratigraphic information with the relative order of the fault–slip data sets.
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14 280 We only used the automatic separation of the fault-slip data to cross-check the results
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16 281 derived from the criteria presented above.
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21 283 Stress tensors were determined using several steps. Each reduced stress tensor comprises
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23 284 the directions and plunges of the principal stress axes ($\sigma_1 \geq \sigma_2 \geq \sigma_3$) and the stress ratio $R =$
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25 285 $(\sigma_2 - \sigma_3) / (\sigma_1 - \sigma_3)$ (Angelier, 1979). We used the direct inversion method (Angelier, 1990,
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27 286 1994) based on Bott’s principle (Bott, 1959), which we processed in the Tectonics FP
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29 287 package (Otner et al., 2002). In addition, we imposed a quality standard for the fault-slip
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31 288 data. For example, faults with striae deviating by more than 10° from the fault plane were
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33 289 disregarded, and only corrected fault sets were used to calculate the orientations of the
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36 290 principal axes.
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41 292 3.2. Seismic, geoelectric and topographic investigations

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43 293 Seismic, geoelectric and borehole data were utilized to investigate the architecture of the
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45 294 Potiguar Basin and the postrift tectono-sedimentary fill in the central part of the basin.
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48 295 The data included three previously stacked and migrated 2D seismic sections and three
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50 296 exploratory wells from the Brazilian Petroleum Agency (ANP) database. In addition, 30
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53 297 vertical electric soundings (VESs) were performed along a 55 km profile near one
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56 298 seismic line and oriented roughly perpendicular to the rift faults. We used the VESs to
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4 299 define geoelectric layers and the internal geometry of the uppermost part of the basin,
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6 300 where the seismic lines have poor resolution. The soundings were spaced 2.0 km apart,
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9 301 and all of the measurements were taken using Schlumberger's electrode array with current
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11 302 electrode half-spacings ($AB/2$) ranging from 1.5 m to 600 m. We used a Tectrol
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13 303 resistivity meter with a maximum power of 500 W, which was able to provide the
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16 304 apparent resistivity with high accuracy for estimated investigation depths of 250-300 m.
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18 305 Estimates of the resistivity and thickness were calculated from the VES data using the
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20 306 IPI2Win software developed by [Bobachev \(2003\)](#). Interactions using this code were
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22 307 carried out automatically and interactively until the calculated model satisfied a minimum
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24 308 difference between the measured and calculated data.
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310 The results are presented in a geoelectric pseudo-section derived from 1-D inversions of
311 each of the 30 VESs, which were applied separately to obtain the resistivity distribution
312 in the shallow subsurface (Fig. 4). Seismic and well information were used to constrain
313 the 1-D inversion and interpret the geoelectric pseudo-section. The root-mean-square
314 misfits between the observed and calculated geoelectric curves (RMS errors in Fig. 4) are
315 in the range of 0.25 - 12.7%.

316

317 **4. Results**

318 *4.1. Postrift paleostress fields*

319 Two paleostress fields were identified based on the criteria outlined in Section 3. We list
320 the site locations, amount of data, and stress tensors for each stress field in [Table 2](#). Each

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4 321 stress fields are presented below with the general stress parameters, and the analysis of
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6 322 the deformation structures and their significance.
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11 324 Stress field 1 (SF1) was identified in both the siliciclastic units of the Açu Formation and

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14 325 the carbonates of the Jandaíra Formation. Most sites revealed polyphase deformation

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16 326 reaching two stress fields. A total of 22 stress tensors represent SF1, 14 of which are

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18 327 strike-slip tensors, 2 are normal tensors, and 6 are reverse stress tensors. The stress

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21 328 tensors indicate a subhorizontal N-S-oriented σ_1 and a subhorizontal E-W-oriented σ_3

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23 329 (Fig. 5, Table 2). SF1 is also marked by a set of N-S-striking mode I joints and

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26 330 orthogonal E-W-striking vertical stylolites. The maximum horizontal compression (σ_1) is

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28 331 consistently oriented N-S with a $\sim 10^\circ$ variation in azimuth and a $\sim 5^\circ$ variation in plunge.

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31 332 Generally, σ_2 is subvertical, and σ_3 is generally subhorizontal with an E-W-trend, but

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33 333 these two axes shift position at several sites (Fig. 5). The mean concentration of stress

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36 334 tensors indicates the following: 357/15 σ_1 , 268/98 σ_2 , and 087/60 σ_3 (Fig. 5). The stress

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38 335 ratios (R) for SF1 range from 0.8234 to 0.2185. The changes from the main strike-slip

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41 336 regime to the local normal and reverse regimes are due to shifts between σ_2 and σ_3 or

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43 337 between σ_1 and σ_2 . Faults of SF1 are pervasive in the basin, as attested by remote sensing

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46 338 and field data from current and previous studies (Bezerra et al., 2009, 2014).
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51 340 SF1 was followed by a second stress field (SF2). Where a relative chronology could be

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53 341 established, the faults and joints of SF2 systematically postdate those of SF1. Less than

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56 342 50% of the faults from either SF1 or the rift phase were reactivated during SF2. We

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58 343 obtained 21 stress tensors (Fig. 6, Table 2), of which 15 indicate a strike-slip regime, 1
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4 344 indicates a normal fault regime, and 5 indicate a reverse fault regime. These stress tensors
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6 345 also indicate partitioning between strike-slip and normal faulting. SF2 shows a
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9 346 subhorizontal E-W-trending σ_1 , a subvertical σ_2 , and a subhorizontal N-S-trending σ_3 in
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11 347 the eastern part of the basin. However, σ_1 and σ_3 rotate clockwise in the western part of
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14 348 the basin, where σ_1 trends WNW-ESE, and σ_3 trends NNE-SSW at four sites. The mean
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16 349 SF2 of the stress tensors yielded 267/18 σ_1 , 270/84 σ_2 , and 023/63 σ_3 (Fig. 6) with R
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19 350 ranging from 0.9928 to 0.0481 (Table 2). The variation in R is larger than that observed
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21 351 in SF1. Approximately 50% of the major faults of SF2 are the same as those of SF1,
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24 352 suggesting that the reactivation of pre-existing faults was the mode of failure during SF2
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26 353 at several sites.
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31 355 The differences in the directions of the maximum compression (σ_1) for the strike-slip
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33 356 faults of SF1 (N-S trends) and SF2 (mostly E-W trends) indicate that the paleostress
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36 357 tensors for each stress field can be grouped as a single tectonic episode. However, the
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38 358 permutation between σ_2 and σ_3 and the wide range in R in both stress fields indicate that
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41 359 the relative magnitudes of the principal stress axes vary, although they exhibit the same
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43 360 orientations.
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47 48 362 *4.2. Major postrift faults and volcanic units*

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51 363 In the postrift period, the major structures of the main graben and horsts in the rift phase
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53 364 were reactivated, and new faults were generated. We focus on the central part of the
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55 365 basin, where the major faults are closely associated with basaltic volcanic units (Fig. 7).
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4 366 These faults are mainly NW-SE-striking dextral faults and NE-SW-striking sinistral
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6 367 faults.

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11 369 The location, geometry, and age of the volcanic rocks in the Potiguar Basin and the
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13 370 adjacent basement indicate their relationship to the major faults. Most of the basaltic
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15 371 rocks in the onshore area of the Potiguar Basin form NW-SE-oriented bodies along the
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17 372 NW-SE-striking Afonso Bezerra Fault and other parallel faults (Fig. 7A). Some of the
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19 373 volcanic rocks are deformed by right-lateral fault offsets (Fig. 7B), which indicate
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21 374 magmatic injection mainly during SF1. This is observed on the eastern contact of the
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23 375 Serra Preta basaltic body, which is marked by a right-lateral strike-slip fault of SF1
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25 376 (numbered 1 in Fig. 7B). Other volcanic bodies, such as the Serra Aguda plug (numbered
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27 377 2 in Fig. 7B), are intersected by the NW-SE-striking faults. As shown in Table 1, the
28
29 378 existing K-Ar ages of the Serra Aguda and the Serra Preta bodies are 24.6 ± 6.0 Ma and
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31 379 18.0 ± 0.5 to 13.9 ± 2.0 Ma, respectively. These results constrain the end of the SF1 event
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33 380 and the onset of the SF2 event to between 25 Ma and 14-18 Ma.
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44 382 *4.3. Geophysical evidence of tectonic inversion*

45 383 The analysis of several rift structures is relevant for reconstructing the evolution of the
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47 384 postrift phase. The three seismic lines reveal an internal geometry composed of two main
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49 385 half-grabens that dip to the SE and are separated by a central basement horst (Figs. 8, 9A,
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51 386 B, C). Intracrustal listric normal faults controlled the graben geometry. The border faults
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53 387 and other faults nearby were reactivated by uplifting the most recent layers (RF in Fig.
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55 388 9A, B, C). A subhorizontal unconformity marks the stratigraphic contact between the
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4 389 synrift and postrift sequences (U in Fig. 9A). Evidence of tectonic inversion is present in
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6 390 the central part of the main graben, where subvertical listric faults exhibit reverse
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9 391 displacements in the rift layers. Each seismic line contains evidence of reverse faulting,
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11 392 although the throws of these fault are much smaller than those of the rift faults (RF in
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14 393 Fig. 9A, B, C).

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18 395 The main graben observed in the seismic lines (Fig. 9A) coincides with the area of
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20 396 highest topography in the basin, which we defined as the Mel Dome (Figs. 8; distance
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22 397 from 2 to 26 km in Fig. 9A). A section derived from the interpretation of a geoelectric
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24 398 survey along this dome to a depth of 300 m combined with well data reveals details not
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26 399 observed in the seismic lines (Fig. 10). This section combined the results of 1D inversion
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28 400 of the resistivity data from all 30 VESs. Figure 4 displays the individual inversions of
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30 401 VESs 1, 12, 21, and 30, whose geoelectric patterns are characterized by three to six
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32 402 resistivity layers with a large resistivity range (10 to 15500 Ω .m). The upper high-
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34 403 resistivity layer may be caused by dry soil on the surface, which is apparently absent in
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36 404 VES30 (Fig. 4D). VES12 and VES21 are located at the Mel Dome and display a
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38 405 sequence of very high- (<10000 Ω .m) and low- (\sim 10 Ω .m) resistivity layers related to
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40 406 silty and clayey units of the Barreiras Formation. The underlying low- to medium- (30 to
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42 407 750 Ω .m) resistivity layers represent the carbonate units of the Jandaíra Formation. The
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44 408 interpreted geoelectric section (Fig. 10) indicates that the highest portion of the Mel
45
46 409 Dome is in the half-graben, which is controlled by a border fault (at a distance of 28 km
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48 410 in Fig. 10). We identified a basal layer in the wells that represents the postrift siliciclastic
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50 411 sequence (Açu Formation; yellow layer in Fig. 10). In this area, the carbonate Jandaíra
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4 412 Formation (blue layer in Fig. 10) and the Miocene siliciclastic Barreiras Formation
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6 413 (brown and green layers in Fig. 10) are thicker in the central part of the dome, suggesting
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8 414 that the postrift sedimentary unit was originally deposited in a topographic depression
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10 415 that was later uplifted by tectonic inversion. The marine Barreiras Formation forms the
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12 416 top of a gently folded dome (the Mel Dome) with elevations ranging from 0 m along the
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14 417 shoreline to at least 200 m in the central part of the dome.
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21 419 *4.4. Surface evidence of tectonic inversion*

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23 420 The Potiguar Basin is marked by tablelands of Cretaceous limestones that dip 2°-3° to the
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25 421 north. These tablelands are interrupted by a fold in the central part of the basin, which
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27 422 corresponds to the Mel Dome. The dome has a NE-oriented elliptical shape and is ~50
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29 423 km long, ~35 km wide, and 273 m high. We distinguish this dome (Fig. 11) from those
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31 424 related to volcanic activity, which are circular and smaller. The Mel Dome caused the
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33 425 shoreline to advance more than 10 km seaward. This position of the Barreiras Formation
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35 426 constitutes a unique feature of the Brazilian coast.
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44 428 In the postrift phase of a sedimentary basin, the rivers would be expected to drain towards
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46 429 the centre of the basin due to thermal subsidence. However, the Mel Dome exhibits radial
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48 430 drainage, where the rivers drain away from the central trough in the directions of two
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50 431 valleys: Assu to the east and Mossoró to the west (Fig. 11). Uplift of the Mel Dome
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52 432 stimulated renewed sedimentation to form the Quaternary alluvial deposits. In addition,
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54 433 the traces of the rift faults are oblique to the Assu Valley, but they match the Mossoró
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4 434 Valley, where they form topographic breaks, indicating that the deformation continued
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6 435 into the Miocene-Quaternary (Fig. 8).
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11 437 **5. Discussion**

13 438 *5.1. Ages of stress fields*

16 439 The Potiguar Basin has been affected by two nearly orthogonal stress fields since the
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18 440 Campanian. The stress tensors shifted almost orthogonally from SF1 to SF2. Both stress
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20 441 fields exhibit subhorizontal compression oblique to the rift faults. The relationship
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22 442 between the faults and the stress field indicates oblique-slip movements of the major rift
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24 443 faults in both stress fields. Transpression and transtension coexisted during both strike-
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26 444 slip regimes.
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33 446 Our study is the first to present paleostress evidence for two stress fields during the Late
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35 447 Cretaceous-Cenozoic on the South American equatorial margin. Both stress fields are
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37 448 younger than the rifting, and their ages can be constrained by stratigraphic correlations.
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39 449 Rifting along the equatorial margin of South America started in the Neocomian (ca. 140
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41 450 Ma) (Matos, 1992; Gasperini et al., 2001) and ceased in the Albian (ca. 110 Ma) (Masle
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43 451 et al., 1988). This time period represents the maximum age of SF1. In the Potiguar Basin,
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45 452 deformation structures related to SF1 are located in the Açu and Jandaíra formations and
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47 453 in the Miocene Barreiras Formation (23-17 Ma) at a few sites. This indicates a minimum
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49 454 early Miocene age for SF1.
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4 456 SF1 was followed by SF2, which affected the Miocene and Quaternary sedimentary units.
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6 457 This shift in the stress field occurred from the early to middle Miocene. During SF2,
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9 458 reactivation occurred along the varied geometric style of the SF1 phase. The fault-slip
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11 459 data indicate continuous deformation along faults from the middle Miocene to the
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14 460 present. The shift from SF1 to SF2 also corresponds to the end of the major pulse of the
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16 461 Macau volcanism (Fig. 3) and to the onset of the deposition of the Barreiras Formation
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19 462 (Rossetti et al., 2013).
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23 464 Mainly during SF1, faults allowed mantle magmas to reach the upper crust and controlled
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25 465 the emplacement of several shallow volcanic and hypabyssal bodies (e.g., dikes, sills,
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27 466 plugs, laccoliths, and lopoliths) within Phanerozoic basins in NE Brazil (Souza et al.,
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29 467 2013; Damaceno et al., 2017). The basaltic intrusions indicate relative weakness along
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31 468 the Afonso Bezerra fault system because failure is necessary to open faults and channel
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33 469 mantle-derived melts to reach the upper crust (Lowrie et al., 1986; Mladenović et al.,
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35 470 2015).
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43 472 The seismic line and wells indicate that pulses of inversion younger than Neogene-
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45 473 Quaternary occurred in the basin during SF2, and that the rift faults propagated into the
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48 474 postrift units and folded the Miocene and Quaternary strata (Figs. 9, 10). The coastal
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50 475 marine Barreiras Formation reaches an elevation of at least 200 m, which is well above
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52 476 its average elevation of nearly 70 m near the Brazilian coast (Rossetti et al., 2013). The
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54 477 presence of these deposits at elevations up to 250 m asl at sites at the Mel Dome and at
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57 478 depths below the modern sea level in offshore areas indicates a significant variation in
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4 479 height that cannot be explained solely by sedimentation. The Quaternary sediments that
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6 480 cap the Barreiras Formation are at elevations as high as 270 m in the Mel Dome (Fig. 8).
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9 481 The resulting interpretation of these upper postrift layers indicates that the Mel Dome is a
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11 482 broad anticline. In addition, fold hinges near and parallel to the rift faults are consistent
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14 483 with our results and have been described in several oil fields along the eastern rift border
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16 484 faults (Souto Filho et al., 2000). The dome and adjacent alluvial valleys indicate that the
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19 485 Potiguar Basin is currently experiencing tectonic inversion and that the postrift units,
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21 486 mainly those in the central part of the basin, are being uplifted and exhumed. We interpret
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24 487 the Mel Dome as the surface expression of the tectonic squeezing of the basin. We
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26 488 propose that the present-day dome relief and the radial drainage pattern are related to
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29 489 positive tectonic inversion since the Neogene that is caused by a horizontal maximum
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31 490 compression oriented orthogonal to the NE-striking rift faults.
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36 492 Inversion of the normal faulting stress regime during rifting to a strike-slip stress regime
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38 493 in the postrifting phase was previously proposed for the Araripe (Marques et al., 2014)
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41 494 and Rio do Peixe (Nogueira et al., 2015) basins, which are located 350 km and 200 km
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44 495 south of the equatorial margin, respectively (Fig. 1B). In the Araripe Basin, inversion
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46 496 caused uplift of marine deposits to continental conditions and caused basin dissection by
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48 497 alluvial incision (Marques et al., 2014). Although both basins show evidence of E-W- to
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51 498 ENE-WSW-trending subhorizontal compression, they did not record an initial stress field
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53 499 with N-S-trending subhorizontal compression, such as in the Potiguar Basin. In addition,
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55 500 the stress fields in the Araripe Basin (located 200 km SW of the Potiguar Basin; Fig. 1B)
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58 501 and Rio do Peixe Basin (located 120 km south of the Potiguar Basin; Fig. 1B) in
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4 502 northeastern Brazil inverted from normal stress fields during rifting to strike-slip stress
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6 503 fields during the postrifting phase. This shift occurred from the Late Cretaceous to the
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9 504 Miocene, when the NW-SE-extension transitioned to ENE-WNE horizontal compression.
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11 505 The exact timing of this shift could not be determined due to the lack of temporal markers
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14 506 in these basins (Marques et al., 2014; Nogueira et al., 2015), which is a chronological
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16 507 problem that is resolved by our data.
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20 21 509 *5.2 The influence of the stress fields on the petroleum system of the Potiguar Basin*

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23 510 The petroleum systems of the Potiguar Basin have two main source units. The first is the
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25 511 lacustrine black shales of the Pendência Formation (rift phase), located onshore in the
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28 512 Potiguar grabens. This source had its peak oil generation during the Campanian, which
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31 513 migrated to the same unit (Mello et al., 1988; Bertani et al., 1990). The second oil source
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33 514 is the marine-evaporite and lacustrine black shales and marls of the Alagamar Formation
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36 515 (transition between rift and postrift), which is responsible for more than 70% of the oil
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38 516 that accumulated in the basin. The oil from this second source was generated in the
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41 517 offshore portion of the basin during the Miocene-Holocene (Souto Filho et al., 2000).
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43 518 This oil migrated along the major fault systems (Figs. 2, 8) and filled the postrift
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45 519 reservoirs of the Alagamar and Açu formations (Fig. 8) until it reached the farthest
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48 520 reservoir in the onshore portion of the basin, more than 100 km from the offshore source
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51 521 (e.g., Mello et al., 1988; Bertani et al., 1990; Souto Filho et al., 2000; Penteadó et al.,
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53 522 2007). The present study indicates that the reactivated NE-SW-striking rift faults (Areia
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55 523 Branca and Carnaubais systems; Figs. 2 and 8) played a major role in the migration of oil
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58 524 in the postrift period during the SF1 and SF2 stress fields.
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6 526 The stress data we present in this study area are consistent with previous descriptions of
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9 527 the traps of the Alagamar-Açu petroleum system in the main reservoirs of the postrift
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11 528 units. Two main NE-SW-striking fault systems that cut across the postrift units
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14 529 concentrate the oil fields: the Areia Branca and the Carnaubais (Fig. 8). Both fault
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16 530 systems are postrift reactivations of major rift fault systems (Bertani et al., 1990). The
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19 531 drag of postrift layers along these fault systems in a strike-slip regime folded and formed
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21 532 anticlines. A shale unit with good lateral continuity at the top of the Açu Formation acted
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24 533 as a regional seal that trapped the oil along the NE-oriented hinges of the anticlines
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26 534 ([Souto Filho et al., 2000](#)). The stress data presented in this study indicate that the
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29 535 maximum compressions of both SF1 and SF2 were subhorizontal and oblique to the main
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31 536 NE-SW-striking fault systems, which would induce compression and fold dragging,
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34 537 consistent with the creation of anticlines that formed the traps for the Alagamar-Açu
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36 538 petroleum system. In addition, studies in the basin support the hypothesis that most of the
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39 539 oil generated in the Alagamar Formation migrated along these ENE-WSW-striking
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41 540 fractures that formed and have remained open since the Eocene ([Souto Filho et al., 2000](#)).
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44 541 These fractures are consistent with mode-I tensional fractures that formed during SF2,
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46 542 mostly after the Miocene.
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50 544 A pattern of deformation favorable for hydrocarbon accumulation similar to the one
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53 545 presented in this study was identified by [Gomes et al. \(2014\)](#) in the offshore part of the
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56 546 Potiguar Basin. In addition, Cenozoic anticlines related to mass movements were
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58 547 identified, for example, by [Kruger et al. \(2012\)](#) in the Barreirinhas Basin and Pellegrini
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4 548 and [Ribeiro \(2018\)](#) in the Pará Maranhão Basin ([Fig. 1A](#)). Despite the difference in the
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6 549 deformation mechanism of these mass movements compared to those described in the
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9 550 present study, these anticlines are the likely sites for hydrocarbon accumulation.

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14 552 *5.3 Origin of the stress fields and implications for reconstructing the postrift history of*
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16 553 *the equatorial margin of South America*

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19 554 The onset of postrift stress fields in the Potiguar Basin is consistent with major events on
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21 555 the equatorial margin and indicates how knowledge of the former can contribute to an
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23 556 understanding of the latter. The deformation and opening of the equatorial Atlantic were
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26 557 diachronous ([Szatmari, 2000](#)). However, the chronology of the most important event is
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29 558 well established ([Matos, 2000](#)). The passive margin phase of the equatorial margin
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31 559 (drifting) occurred after the end of the thermal effect of a spreading center, when the
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33 560 tectonic influence of the fracture zones had a minor control on the generation of space for
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36 561 sedimentation (stretching phase) ([Matos, 2000](#)). The rifting phase mainly occurred in the
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38 562 Aptian-Albian but lasted until the Cenomanian ([Matos, 2000](#)). The passive margin stage
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41 563 started in the Cenomanian-Turonian (100-90 Ma), when an oceanic connection between
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43 564 the waters of the central and south Atlantic was established ([Antobreh et al., 2009](#)). This
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46 565 period also coincides with the deposition of the first continental postrift units during the
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48 566 Albian-Turonian in the Potiguar, Ceará, and Barreirinhas basins ([Matos, 2000](#); [Pessoa](#)
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50 567 [Neto et al., 2007](#)). It also coincides with the onset of the first postrift stress field in the
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53 568 Potiguar Basin active since the Campanian (this study).

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4 570 The origin of both stress fields is a matter of debate. The wrench tectonics after the
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6 571 Cenomanian was characterized by both transpression and transtension along the newly
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8 572 created transform fault zone (Matos, 2000; Szatmari, 2000). We suggest that pulses of
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10 573 transpression could have induced the N-S-trending compression and E-W-trending
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12 574 extension, which generated SF1 from the Campanian to the Paleogene in the Potiguar
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14 575 Basin. However, this hypothesis does not rule out other mechanisms. In addition, SF1,
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16 576 which was not previously recognized in the Potiguar Basin, ended during the deposition
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18 577 of the Barreiras Formation. This unit resulted from a major marine transgression in the
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20 578 Miocene, which is also recorded in several other coastal areas of South America (Rossetti
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22 579 et al., 2013). The end of this transgression was marked by unconformities in several
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24 580 sedimentary basins along the Brazilian continental margin (e.g., Pessoa Neto et al., 2007)
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26 581 and West Africa (Fairhead et al., 2013). These unconformities are related to major plate
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28 582 shifts, which led to changes in the stress field (Fairhead et al., 2013). Several studies have
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30 583 related variations in the stress field during the postrift stages to the activity of oceanic
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32 584 fracture zones along the equatorial margins of South America and Africa and major shifts
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34 585 in plate direction (e.g., Davidson et al., 2016).
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45 587 Another explanation for the end of SF1 and the onset of SF2 is a major plate shift coeval
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47 588 with the end of the Incaic period (50-40 Ma) and the onset of the Quechuan period (22-0
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49 589 Ma), which are two major phases of the Andean uplift (Garzzone et al., 2008). During
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51 590 these periods, the topographic gradients of both the Mid-Atlantic ridge (MAR) and the
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53 591 Andes cordillera have induced intraplate stresses necessary for tectonic inversion in the
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55 592 South America plate (e.g., Assumpção, 1992; Cogné et al., 2012). As a detailed
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4 593 discussion presented by Marques et al. (2014) already indicates, the stresses caused by
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6 594 the Andes and MAR have been transmitted into the interior of the South American plate,
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9 595 which behaves as effectively elastic (Marques et al., 2014).

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14 597 Based on the ages of the fault-controlled basaltic magmas, which were emplaced as
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16 598 shallow basaltic intrusions (dikes, plugs, laccoliths, and lopoliths), a major peak of fault
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18 599 reactivation occurred at 30-10 Ma. However, the recurrence of basaltic magmatism since
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21 600 the breakup of Pangea (e.g., Fodor and McKee, 1986; Mizusaki et al., 2002; Knesel et al.,
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23 601 2011; Souza et al., 2003, 2013) along the onshore equatorial margin of South America, as
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25 602 well as Pliocene to Pleistocene magmatism in the Fernando de Noronha Archipelago
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27 603 (Perlingeiro et al., 2013), suggests recurring reactivation of the fault systems. This
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29 604 recurrent basaltic magmatism and fault reactivation may be ascribed to a combination of
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31 605 factors as follows: (1) probable synchronicity with Paleogene and Neogene tectonic
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33 606 events along the Andean margin; (2) a thermal anomaly in the shallow upper mantle of
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35 607 northeastern Brazil, as deduced from geophysical data (Pinheiro and Julià, 2014); (3) a
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37 608 permanent state of melting and periodic segregation of small volumes of basaltic melts
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39 609 (Knesel et al., 2011); and (4) upwelling magma flows of small-scale plate-driven
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41 610 convection in the upper mantle (Knesel et al., 2011; Perlingeiro et al., 2013) inherited
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43 611 from the St. Helens hot spot. Thermal anomalies appear to have travelled west with the
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45 612 South American Plate when the St. Helens hot spot became active in the Cretaceous (120-
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47 613 100 Ma) (cf. Nurnberg and Muller, 1991; Wilson and Guiraud, 1992). A combination of
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49 614 hypotheses (2), (3) and (4) is the likely explanation for this recurrent magmatism because
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4 615 they form a series of processes by which magma is stored beneath the crust to form a
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6 616 thermal anomaly and ascends to the upper crust during pulses of faulting.
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11 618 The transition from SF1 to SF2 in the Potiguar Basin coincides with major events
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13 619 offshore along the equatorial margin. For example, this transition coincides with the shift
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15 620 from gravity gliding with normal faulting near the margin and thrusts at the toe of the
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17 621 continental platform in the Paleogene to E-W-oriented compression on the African and
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19 622 South American side (e.g., [Trosdtorf et al., 2007](#); [Davison et al., 2016](#)). The maximum
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21 623 horizontal compression roughly parallel to the equatorial margin during SF2 is consistent
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23 624 with the extensional-compressional tectonics system (an extensional fault system near the
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25 625 continent, a distal contractional system away from the continent, and a transitional or
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27 626 translational domain between the two that affected the sedimentary units from the Late
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29 627 Cretaceous to the Neogene (e.g., [Oliveira et al., 2012](#)).
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35 629 The origin of SF2 is better constrained due to its correspondence with the present-day
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37 630 stress field, which indicates that they a common origin. The fault-slip data are also
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39 631 consistent with the present-day strain derived from GPS data in this basin ([Marotta et al.,](#)
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41 632 [2015](#)) ([Fig. 12](#)). However, in contrast to SF2, the normal faults in the current stress field
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43 633 are mainly located in the central part of the basin and extend to a depth of 2 km ([Reis et](#)
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45 634 [al., 2013](#)). This apparent contradiction is consistent with the doming in the main graben
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47 635 area.
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4 637 The shift of S_{Hmax} from E-W- to NW-SE in the present-day stress field roughly following
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6 638 the coastline in the central part of the basin was related to the contrast in density between
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9 639 the continental and oceanic crust (Assumpção, 1992; Ferreira et al., 1998). This
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11 640 maximum compressive stress (σ_1) that has been parallel to the continental margin of the
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14 641 Potiguar Basin since the Miocene has also been observed in other passive margins (e.g.,
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16 642 Santimano and Rillen, 2011). Away from the continental margin, however, SF2 continues
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19 643 to be a strike-slip regime, but the maximum compression is oriented ENE-WSW
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21 644 (Marques et al., 2014; Nogueira et al., 2015). This change was related to the temporal
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23 645 shift in the stress field due to plate-driving forces in the Andes. This episode coincided
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26 646 with the 22–0 Ma Quechuan period of uplift (Coutand et al., 2001), when the Andean
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29 647 Plateau was exceptionally high (Garziona et al., 2008). A similar stress pattern of σ_1
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31 648 parallel to the coastline developed with the breakout along the N-S-trending coast of the
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33 649 Sergipe-Alagoas Basin 400 km south of the Potiguar Basin (Lima et al., 1997).
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38 651 The stress fields recorded in the Potiguar Basin have implications for reconstructing the
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41 652 tectonic history of the continental margin of South America, especially SF2, which
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44 653 matches the present-day stress field. For example, the present-day stress field on the
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47 654 eastern continental margin of South America is characterized by a subhorizontal
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50 655 compressive horizontal stress and a strike-slip to reverse stress regime (Assumpção et al.,
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52 656 2014). This subhorizontal compression on the continental margin of Brazil is comparable
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55 657 to that recorded in the southeastern part of the Australian passive margin and along the
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58 658 West African margin (Hudec and Jackson, 2002), where inversion anticlines indicate
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4 659 major periods of Miocene-Quaternary compression ([Schneider et al., 2004](#); [Sharp and](#)
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6 660 [Wood, 2004](#); [Hillis et al., 2008](#)).

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11 662 Repeated fault reactivation and the generation of new faults are manifested in the
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13 663 intraplate seismicity of the area adjacent to the Potiguar Basin ([Bezerra and Vita-Finzi,](#)
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15 664 [2000](#); [Bezerra et al., 2011](#)). The orientations of the rift structures in relation to the stress
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17 665 field ([Reis et al., 2013](#)), overpressuring of faults ([Bezerra et al., 2007](#)), and fault
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19 666 weakening ([Bezerra et al., 2007, 2011](#)) in this basin could be evidence of inversion.
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26 668 Considering the mean elevations of the Barreiras Formation of 67 m asl in the adjacent
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28 669 Paraíba Basin ([Rossetti et al., 2013](#)) and more than 250 m asl in the Mel Dome, it is
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31 670 reasonable to infer that more than 180 m of uplift has occurred over the last 23 Ma. This
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33 671 rate is consistent with those found in Miocene-recent anticlines in Australia, where 200 m
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35 672 of uplift has occurred since the early-mid Pliocene ([Sandiford et al., 2004](#)). Similar uplift
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37 673 rates are not predicted by the classical postrift evolution model of sedimentary basins
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39 674 (e.g., [McKenzie, 1978](#)).

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45 676 The new data presented in this study help to constrain the changes in stress along the
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47 677 continental margin of Brazil from a vertical to horizontal maximum compressive stress,
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49 678 which is consistent with the stress evolution in intraplate South America. The change in
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51 679 the stress field influenced the present-day landscape and also the Quaternary sedimentary
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53 680 record. The process we describe here might have implications for reconstructing the
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4 681 neotectonic history of the eastern continental margin of South America and other passive
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6 682 margins elsewhere.
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10 11 684 **6. Conclusion**

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14 685 The stress field evolution of the Potiguar Basin constrains the postrift history of the
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16 686 equatorial margin of Brazil based on the following conclusions:

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19 687 (a) The well-constrained chronology of sedimentary, volcanic and hypabyssal rocks
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21 688 allowed us to identify two stress fields. After rifting ceased, an intraplate strike-
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23 689 slip stress regime with N-S-trending compression and E-W-trending extension
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26 690 (SF1) occurred from the Late Cretaceous (from at least the Campanian) to the
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29 691 middle Miocene.

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31 692 (b) Stress field SF1 was replaced by a neotectonic strike-slip stress regime (SF2) with
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33 693 E-W- to NW-SE-trending compression and N-S- to NE-SW-trending extension.
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36 694 Shortening during SF2 has been accommodated by strike-slip and reverse faults
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38 695 and folding of postrift Miocene-Quaternary sedimentary units.

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41 696 (c) SF2 shaped the topography of the Potiguar Basin and caused doming of the
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43 697 central part of the basin. Basin inversion shaped the drainage basins into a radial
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46 698 pattern at the basin centre.

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48 699 (d) The onset of the present-day stress field (SF2) occurred at the end of the major
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51 700 peak of volcanism at 30-25 Ma and the beginning of the deposition of the
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53 701 Barreiras Formation at 23-17 Ma.

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55 702 (e) SF2 continues today and is consistent with stress data from focal mechanisms,
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58 703 borehole breakouts, image logs, and strain data from GPS permanent monitoring.
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4 704 (f) Both SF1 and SF2 reactivated the major onshore NE-SW-striking rift fault
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6 705 systems and generated anticlines in the postrift units. These anticlines trapped the
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9 706 petroleum of the Alagamar-Açu system that has migrated from offshore marine-
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11 707 evaporitic and lacustrine units in the transitional sequence. The oil migration
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14 708 process mainly occurred during the Campanian and Miocene-Quaternary.

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16 709 (g) SF1 and SF2 developed in a passive margin stage of the equatorial margin, when
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19 710 the connection between the equatorial Atlantic and South Atlantic was
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21 711 established. This period coincides with the wrenching phase of the equatorial
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24 712 margin.

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26 713 (h) The multitude of data presented in this study supports post-breakup deformation
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29 714 along the equatorial margin of South America, which is under compression and
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31 715 contains mainly of Neogene- to Quaternary-controlled landforms.

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15 1184 **Figure Captions**

16
17 1185 Figure 1 – (A) Topographic map of the Brazilian Equatorial Margin, showing the main
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19 1186 marginal basins (topography from the Global Relief Model from ETOPO1 – NOAA,
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21 1187 www.ngdc.noaa.gov/mgg/global/). The black line marks the coastline. (B) Map of the
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23 1188 present-day stress data of the eastern part of the Brazilian Equatorial margin superposed
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25 1189 on a shaded image derived from the digital elevation model of the shuttle radar
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27 1190 topography mission. Stress indicators are focal mechanisms ([Assumpção, 1992](#); [Ferreira](#)
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29 1191 [et al., 1998, 2008](#); [Bezerra et al., 2011](#); [Lima Neto et al., 2014](#); [Oliveira et al., 2015](#)),
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31 1192 borehole breakouts ([Lima et al., 1997](#)), and image logs ([Reis et al., 2013](#)). Sedimentary
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33 1193 basins quoted in the text: Potiguar (Pb), Ceará (Ceb), Araripe (Ab), Rio do Peixe (RPb),
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35 1194 Parnaíba (Pnb), Iguatu (Igb), Paraíba (Pab), Pernambuco (Peb), Barreirinhas (Bab). Inset:
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37 1195 South American continent with location of the study area.

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46 1197 Figure 2 – Simplified tectonic structures and stratigraphy of the Potiguar Basin. A)
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48 1198 Simplified geological map with main structures of the rift phase. These structures mostly
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50 1199 occur below the post-rift units but are marked on the map for the sake of clarity. B)
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52 1200 Cross-section of the central part of the basin (modified from [Bertani et al., 1990](#); [Bezerra](#)
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54 1201 [et al., 2000, 2011](#)).

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4 1203 Figure 3 – Histogram of ages of the Macau volcanism in the Potiguar Basin and adjacent
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6 1204 Precambrian basement. Ages sources listed in Table 1.

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11 1206 Figure 4 – Observed and calculated resistivity curves and geoelectrical models obtained
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14 1207 from 1D inversion of VES 1, 12, 21, and 30, respectively.

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19 1209 Figure 5 – Paleostress tensors of the stress field 1 (SF1) in the Potiguar Basin. Fault-slip
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21 1210 data are shown in equal-area lower hemisphere stereonet. Fault planes = continuous lines
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23 1211 and striaes; arrows on the circles = sense of movements of hanging walls (double arrows
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25 1212 for strike-slip; inward directed arrows for reverse slip; outward directed arrows for
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27 1213 normal slip). Stress tensors of maximum compressive (S1), intermediate (S2), and
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29 1214 minimum stresses (S3) as five-red, four-green, and three-blue stars, respectively. Black
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33 1216 paleostress tensors of SF1 at the upper right-hand corner.

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39 1218 Figure 6 - Paleostress tensors of the stress field 2 (SF2) in the Potiguar Basin. Fault-slip
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41 1219 data are shown in equal-area lower hemisphere stereonet. Fault planes = continuous lines
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43 1220 and striaes; arrows on the circles = sense of movements of hanging walls (double arrows
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51 1224 arrows outside stereonet = directions of compression and extension. Summary of SF2
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53 1225 stress tensors at the right-hand corner.

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9 1228 of the Potiguar Basin on a background of a digital elevation model from the Shuttle Radar
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11 1229 Topographic Mission (SRTM). A) Volcanic bodies aligned along the NW-SE-striking
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13 1230 Afonso Bezerra fault system. B) Details of a volcanic body along which a right-lateral
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15 1231 displacement of the boundary between the Açú and Jandaíra formations occurs. Number
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17 1232 of volcanic bodies on the map corresponds to site numbers presented on Table 1. The
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19 1233 solid black lines mark the boundaries between units in the Potiguar Basin. See Figure 2
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21 1234 for location. Fault traces and location of volcanic bodies from Bezerra et al. (2009). The
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23 1235 bracket black lines on both maps indicate the segments of the Afonso Bezerra fault
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25 1236 system. Key: 1, Serra Preta basaltic body; 2, Serra Aguda plug.
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33 1238 Figure 8 – Major faults generated during the Potiguar Basin rifting (white lines), and
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35 1239 location of seismic sections, wells and geoelectric survey superposed on a digital
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37 1240 elevation model (DEM) from the Shuttle Radar Topography Mission (SRTM). Blue lines
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39 1241 = seismic lines (L); yellow triangles = vertical geoelectric soundings (W). Volcanic
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41 1242 dome, VD; Mel dome, MD; Assu valley, AV; Mossoró valley, MV. Rift faults after
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43 1243 Bertani et al. (1990), Matos (1992), and Souto Filho (2000). The yellow areas indicate the
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45 1244 location of the main oil fields in the postrift unit, the Açú Formation (Bertani et al.,
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47 1245 1990). Location of area in Fig. 2.
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55 1247 Figure 9 – Seismic sections orthogonal to faults generated during the rifting and
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57 1248 reactivated in the postrift phase of the Potiguar Basin. A) Seismic section L 039. B)

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4 1249 Seismic section L 217. C) Seismic section L 194. In each section, a detail of the seismic
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6 1250 pattern and the corresponding interpretation are presented on the lower part of each
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9 1251 seismic line. (U) mark the unconformity between the rift and postrift sequences. (a'), (b'),
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11 1252 (c') are uninterpreted and (a''), (b''), (c'') are interpreted details of the sections (A), (B),
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14 1253 (C), respectively. Location of sections in Figure 8.

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19 1255 Figure 10 – Geoelectric pseudo-section along the Mel dome based on 30 vertical electric
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24 1257 that were combined with well data to allow inferences down to a 600 m depth. The
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26 1258 topography of the dome was derived from a digital elevation model of the shuttle radar
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29 1259 topographic mission data (DEM-SRTM). Location of VESs in Figure 8.

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33 1261 Figure 11 - Topography of the Mel dome and adjacent Quaternary alluvial valleys. A)
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36 1262 View from the north. B) View from the north, including radial drainage system. Key:
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38 1263 Assu valley, AV; Mel dome, MD; Mossoró valley, MV.

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43 1265 Figure 12 – Maps with the summary of the three stress fields, highlighting the maximum
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46 1266 horizontal directions of compression and extension and fault kinematics. A) Stress field
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48 1267 1. B) Stress field 2. C) Present-day stress field.

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Figure 1

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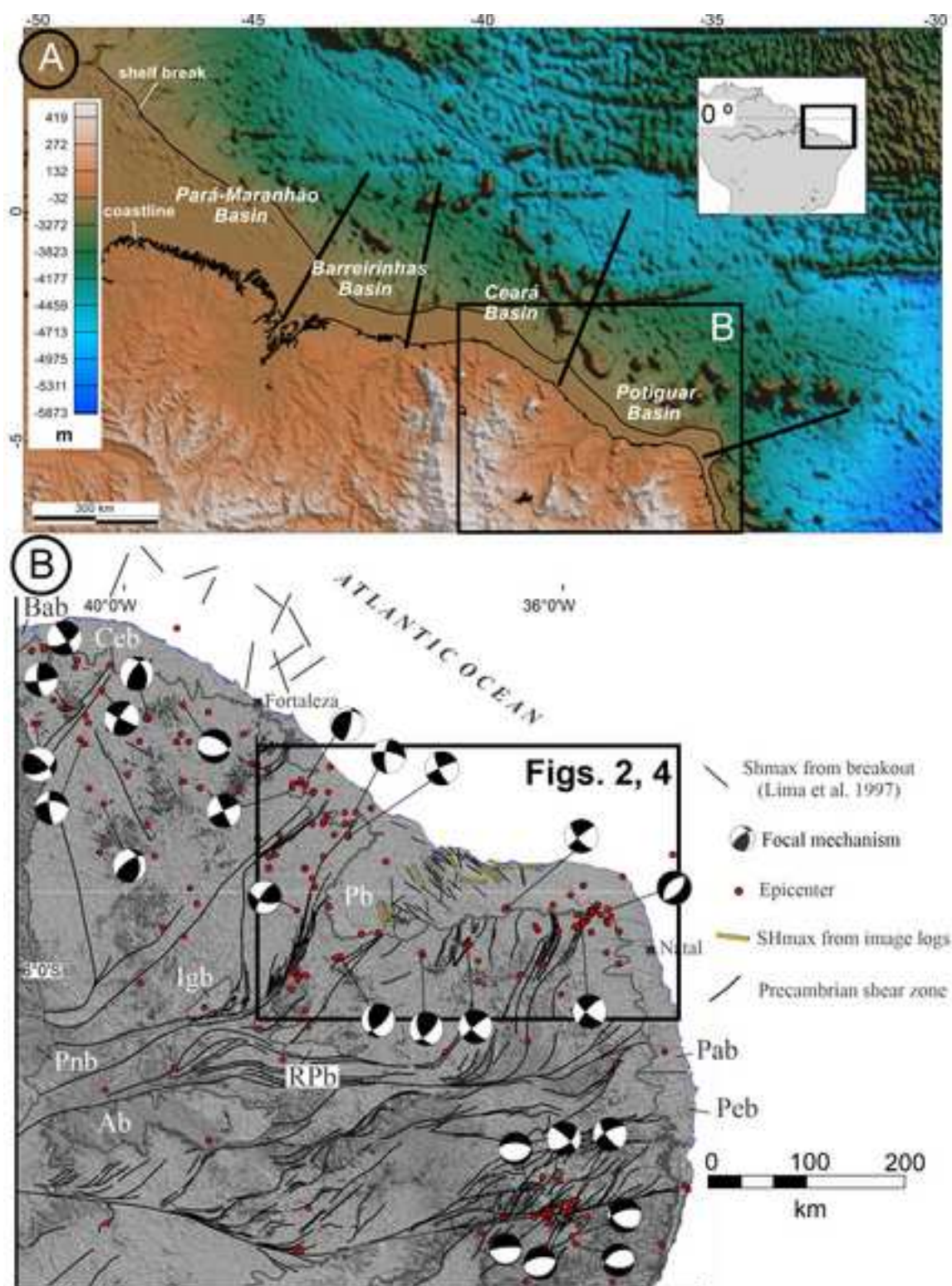


Figure 1 – (A) Topographic map of the Brazilian Equatorial Margin, showing the main marginal basins. (topography from the Global Relief Model from ETOPO1 – NOAA, www.ngdc.noaa.gov/mgg/global/). (B) Map of the present-day stress data of the eastern part of the Brazilian Equatorial margin superposed on a shaded image derived from the digital elevation model of the shuttle radar topography mission. Stress indicators are focal mechanisms (Assumpção, 1992; Ferreira et al., 1998, 2008; Bezerra et al., 2011; Lima Neto et al., 2014; Oliveira et al., 2015), borehole breakouts (Lima et al., 1997), and image logs (Reis et al., 2013). Sedimentary basins quoted in the text: Potiguar (Pb), Ceará (Ceb), Araripe (Ab), Rio do Peixe (Rpb), Parnaíba (Pnb), Iguatu (Igb), Paraíba (Pab), Pernambuco (Peb), Barreirinhas (Bab). Inset: South American continent with location of the study area.

Figure 2
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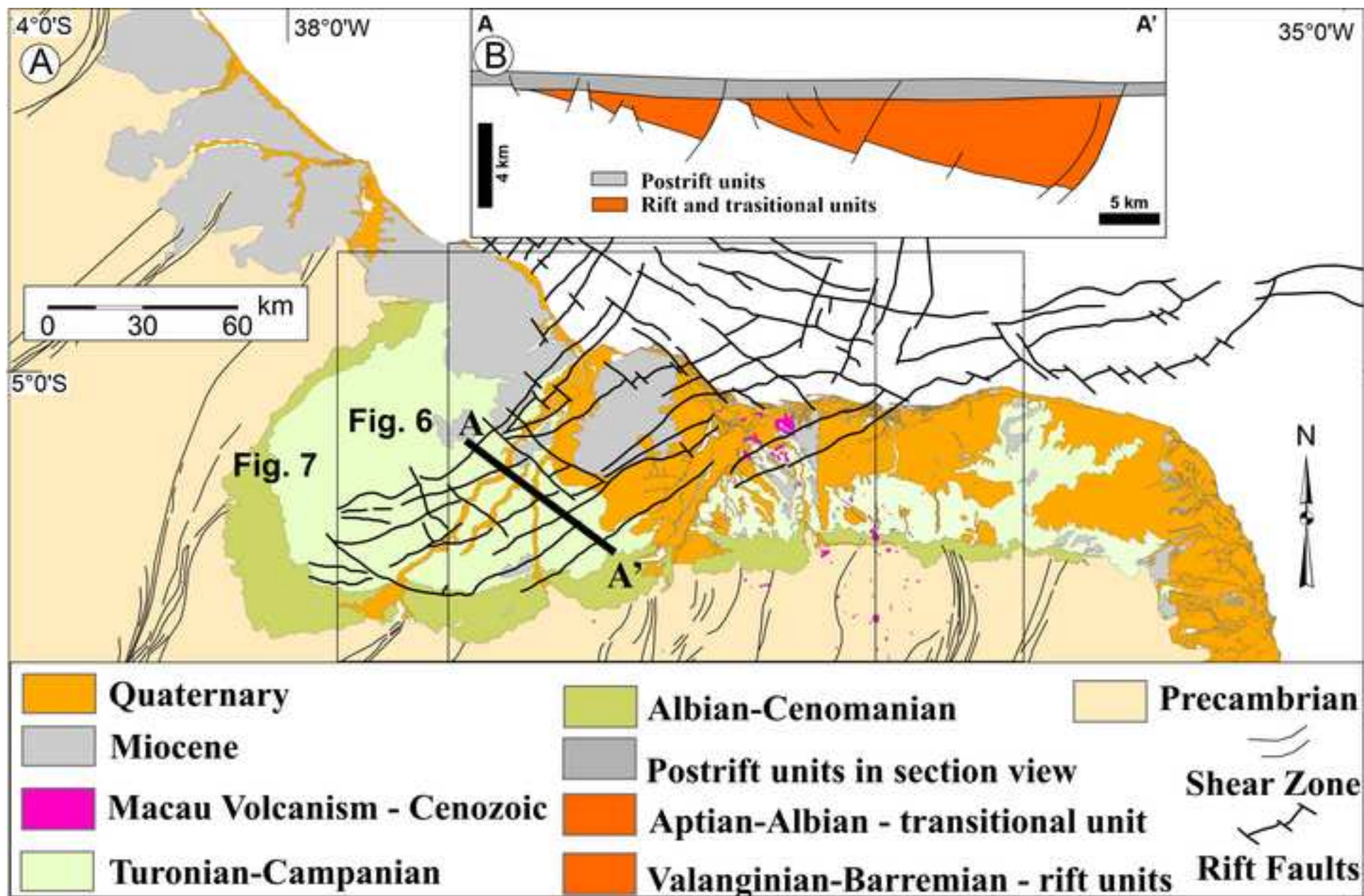


Figure 2 – Simplified tectonic structures and stratigraphy of the Potiguar Basin. A) Simplified geological map with main structures of the rift phase. These structures mostly occur below the post-rift units but are marked on the map for the sake of clarity. B) Cross-section of the central part of the basin (modified from Bertani et al., 1990; Bezerra et al., 2000, 2011).

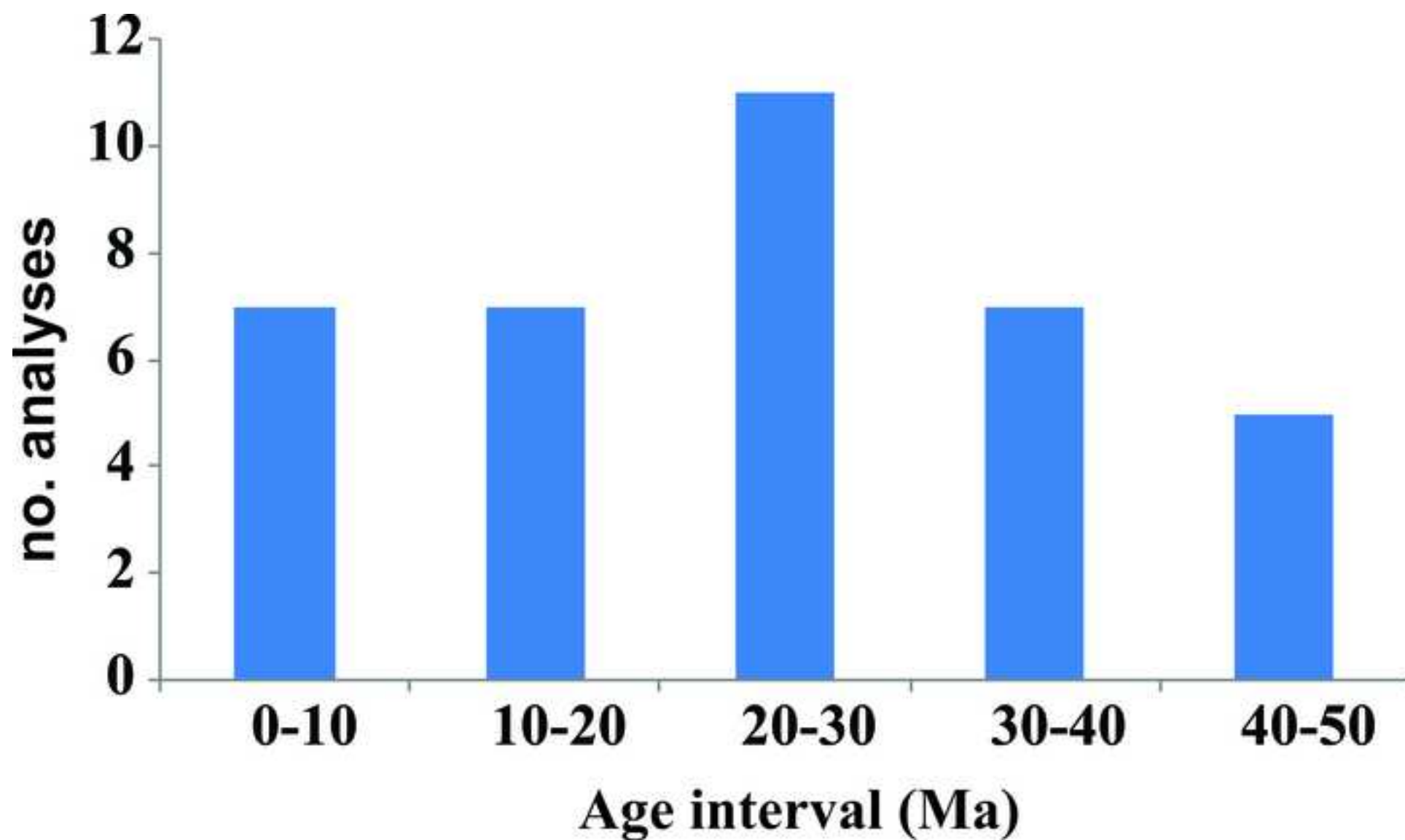


Figure 3 – Histogram of ages of the Macau volcanism in the Potiguar Basin and adjacent Precambrian basement. Ages sources listed in Table 1.

Figure 4
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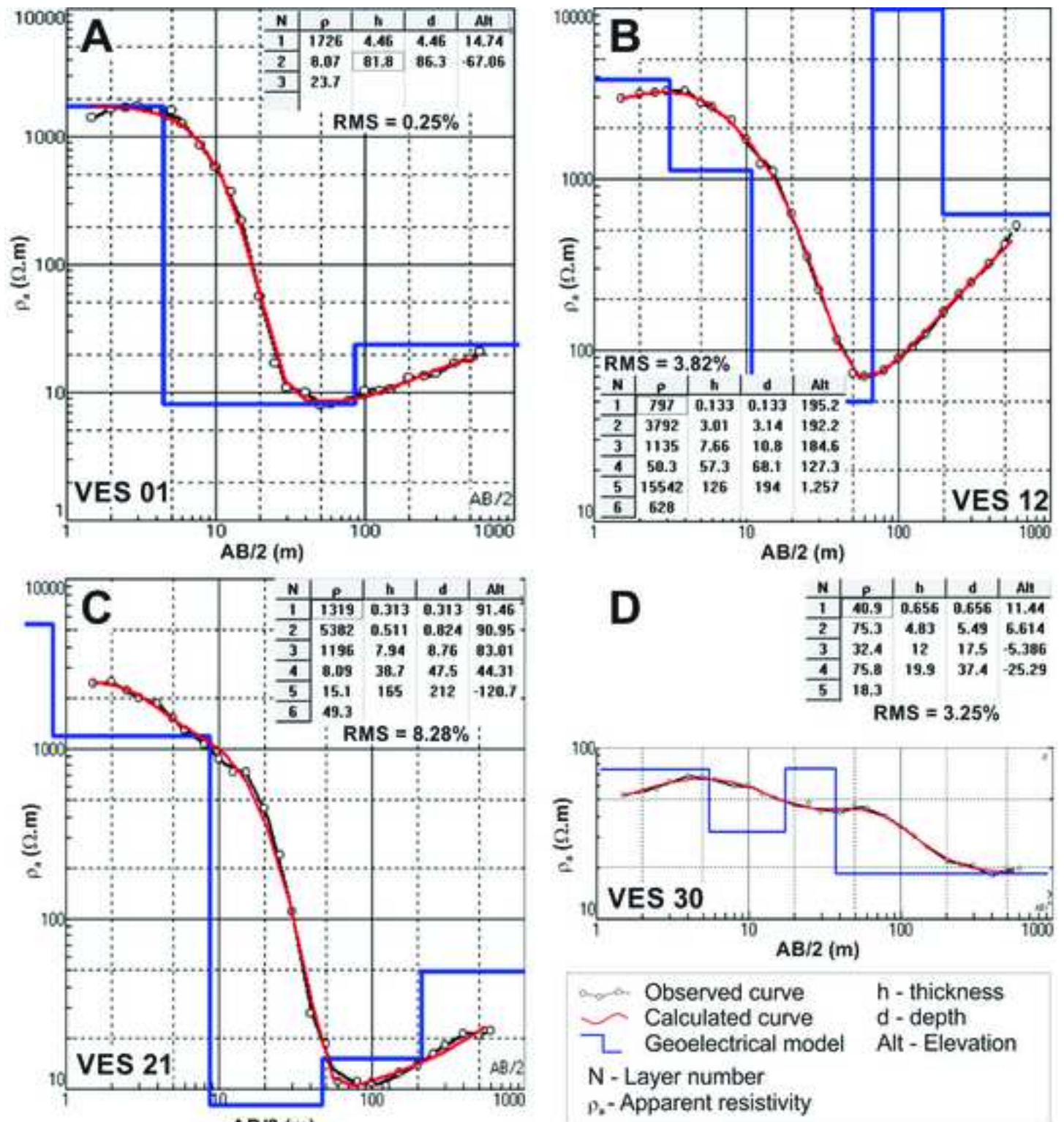


Figure 4 – Observed and calculated resistivity curves and geoelectrical models obtained from 1D inversion of VES 1, 12, 21, and 30, respectively.

Figure 5
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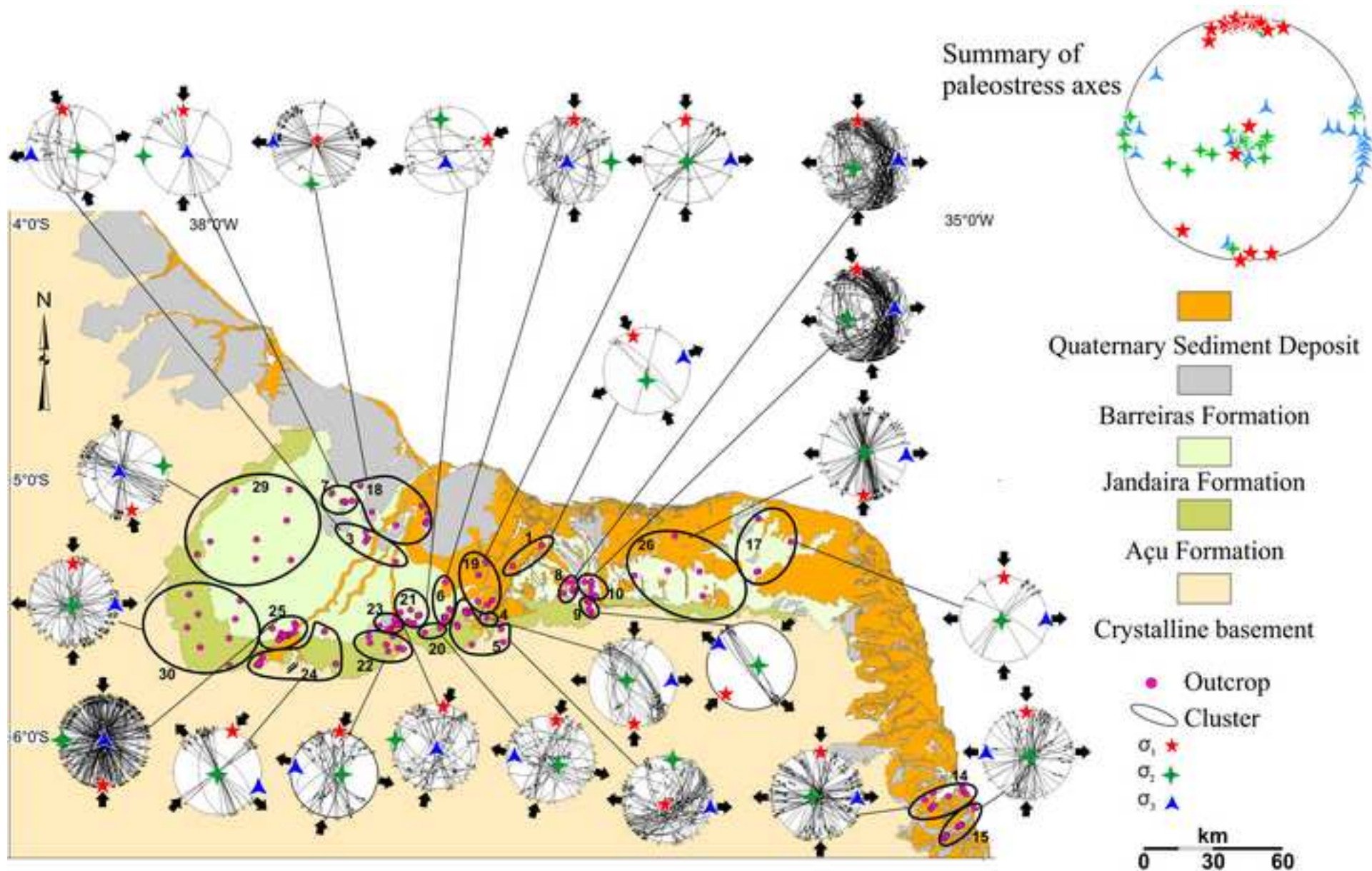


Figure 5 – Paleostress tensors of the stress field 1 (SF1) in the Potiguar Basin. Fault-slip data are shown in equal-area lower hemisphere stereonets. Fault planes = continuous lines and striae; arrows on the circles = sense of movements of hanging walls (double arrows for strike-slip; inward directed arrows for reverse slip; outward directed arrows for normal slip). Stress tensors of maximum compressive (S_1), intermediate (S_2), and minimum stresses (S_3) as five-red, four-green, and three-blue stars, respectively. Black arrows outside stereonets = directions of compression and extension. Summary of paleostress tensors of SF1 at the upper right-hand corner.

Figure 6
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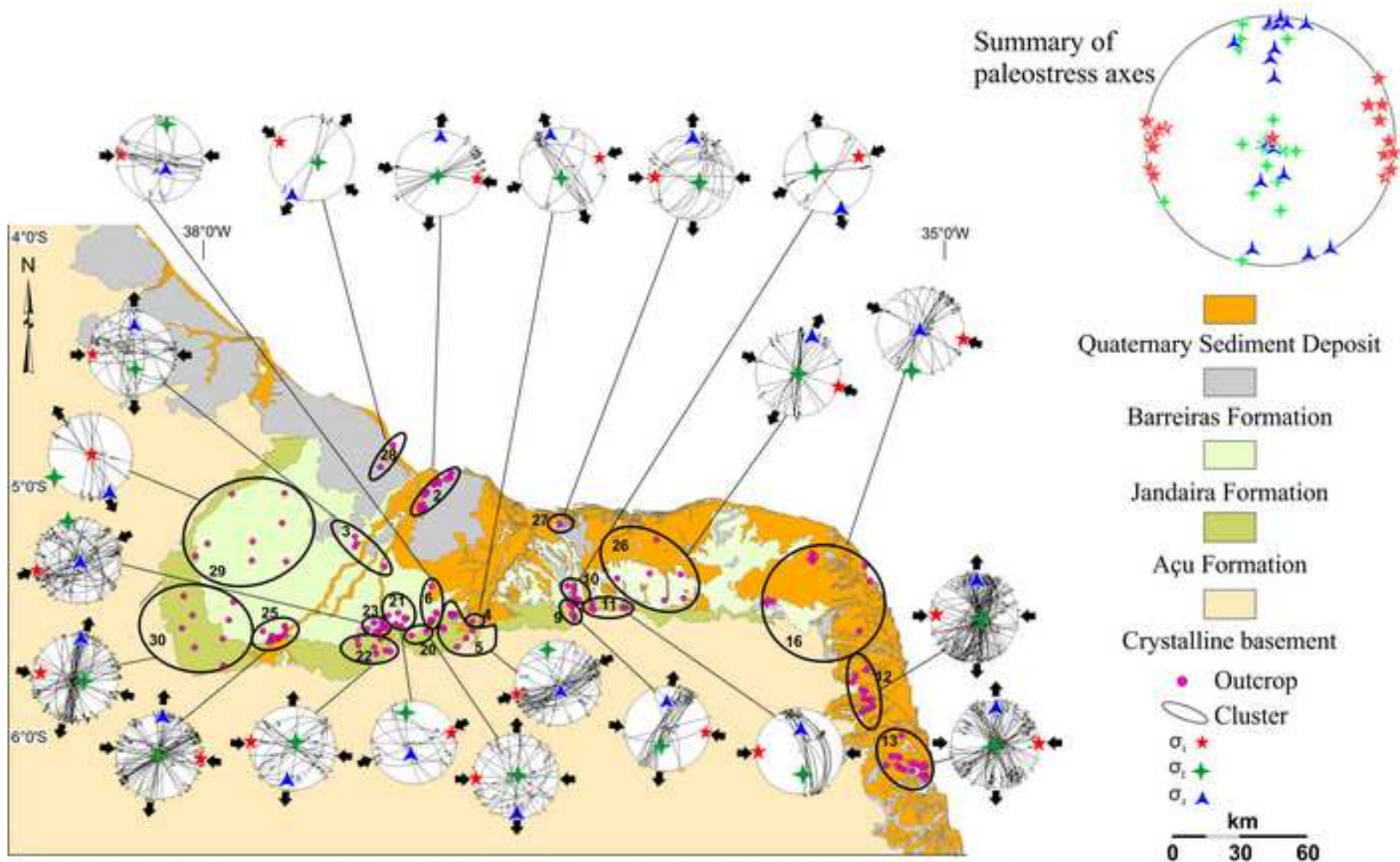


Figure 6 - Paleostress tensors of the stress field 2 (SF2) in the Potiguar Basin. Fault-slip data are shown in equal-area lower hemisphere stereonets. Fault planes = continuous lines and striaes; arrows on the circles = sense of movements of hanging walls (double arrows for strike-slip; inward directed arrows for reverse slip; outward directed arrows for normal slip). Stress tensors of maximum compressive (σ_1), intermediate (σ_2), and minimum stresses (σ_3) as five-red, four-green, and three-blue stars, respectively. Black arrows outside stereonets = directions of compression and extension. Summary of SF2 stress tensors at the right-hand corner.

Figure 7
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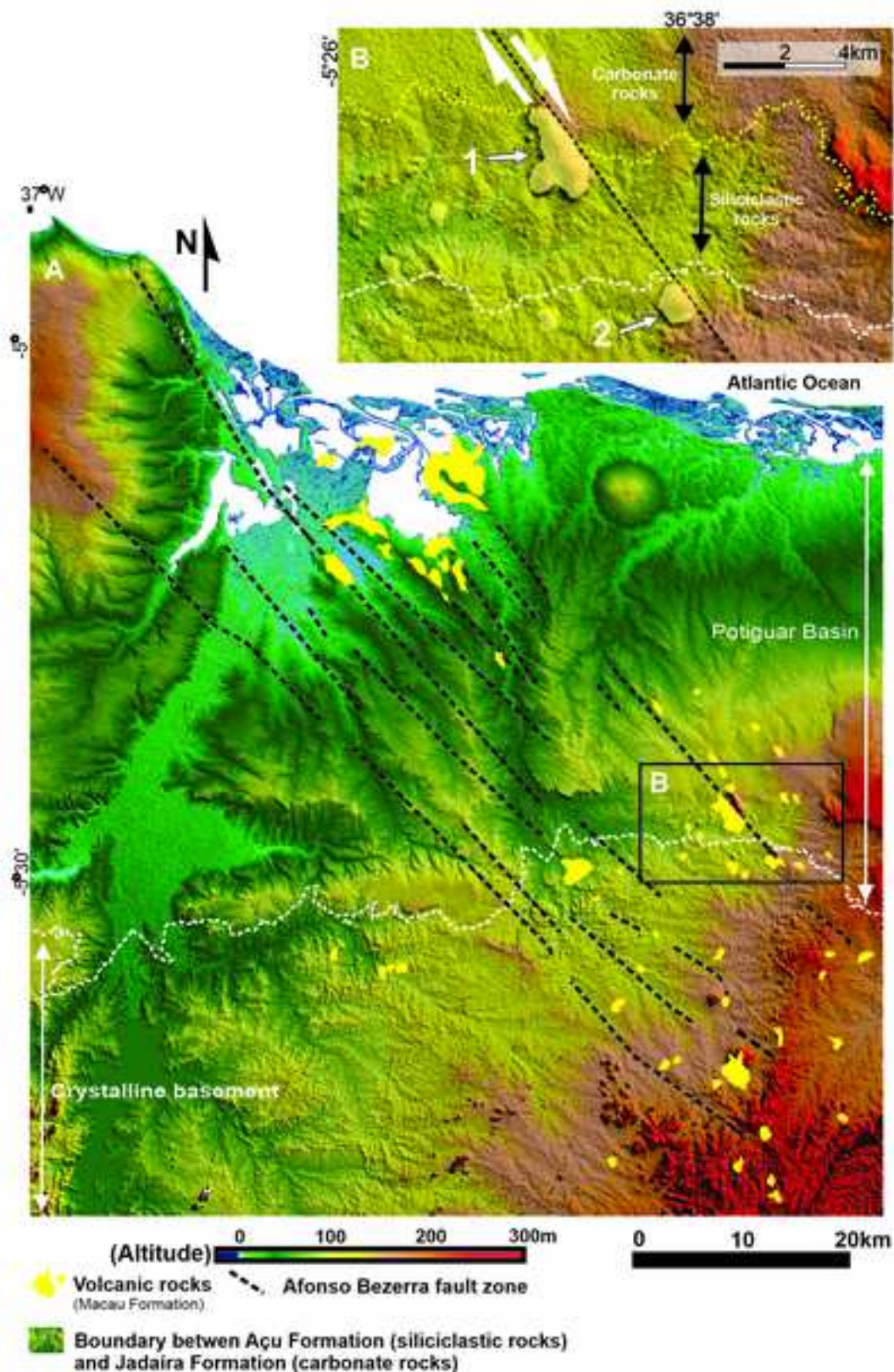


Figure 7 – Fault pattern and Cenozoic volcanic intrusions in the eastern and central area of the Potiguar Basin on a background of a digital elevation model from the Shuttle Radar Topographic Mission (SRTM). A) Volcanic bodies aligned along the NW-SE-striking Afonso Bezerra fault system. B) Details of a volcanic body along which a right-lateral displacement of the boundary between the Açú and Jandaíra formations occurs. Number of volcanic bodies on the map corresponds to site numbers presented on Table 1. See Figure 2 for location. Fault traces and location of volcanic bodies from Bezerra et al. (2009). Key: 1, Serra Preta basaltic body; 2, Serra Aguda plug.

Figure 8
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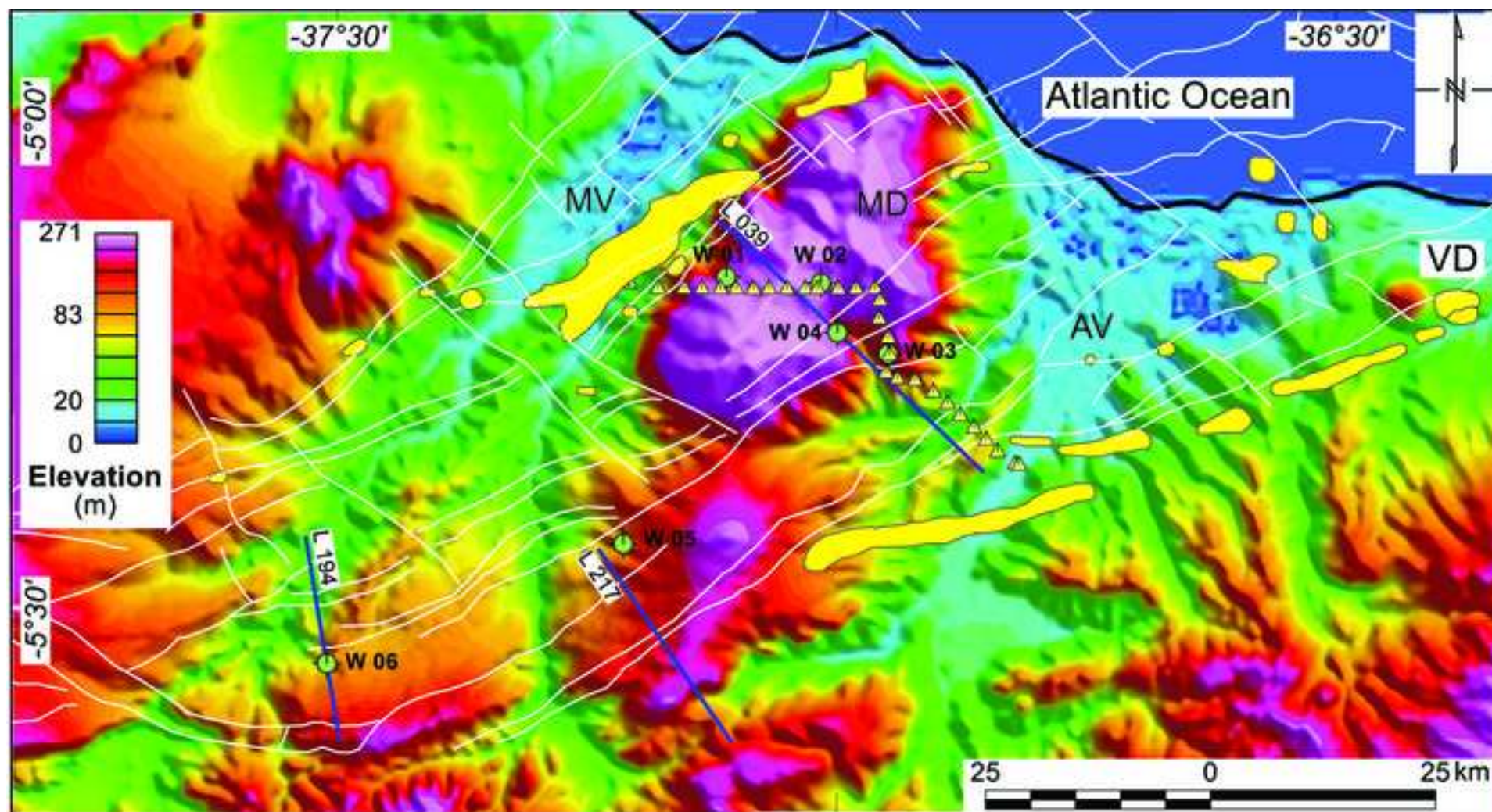


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Figure 9
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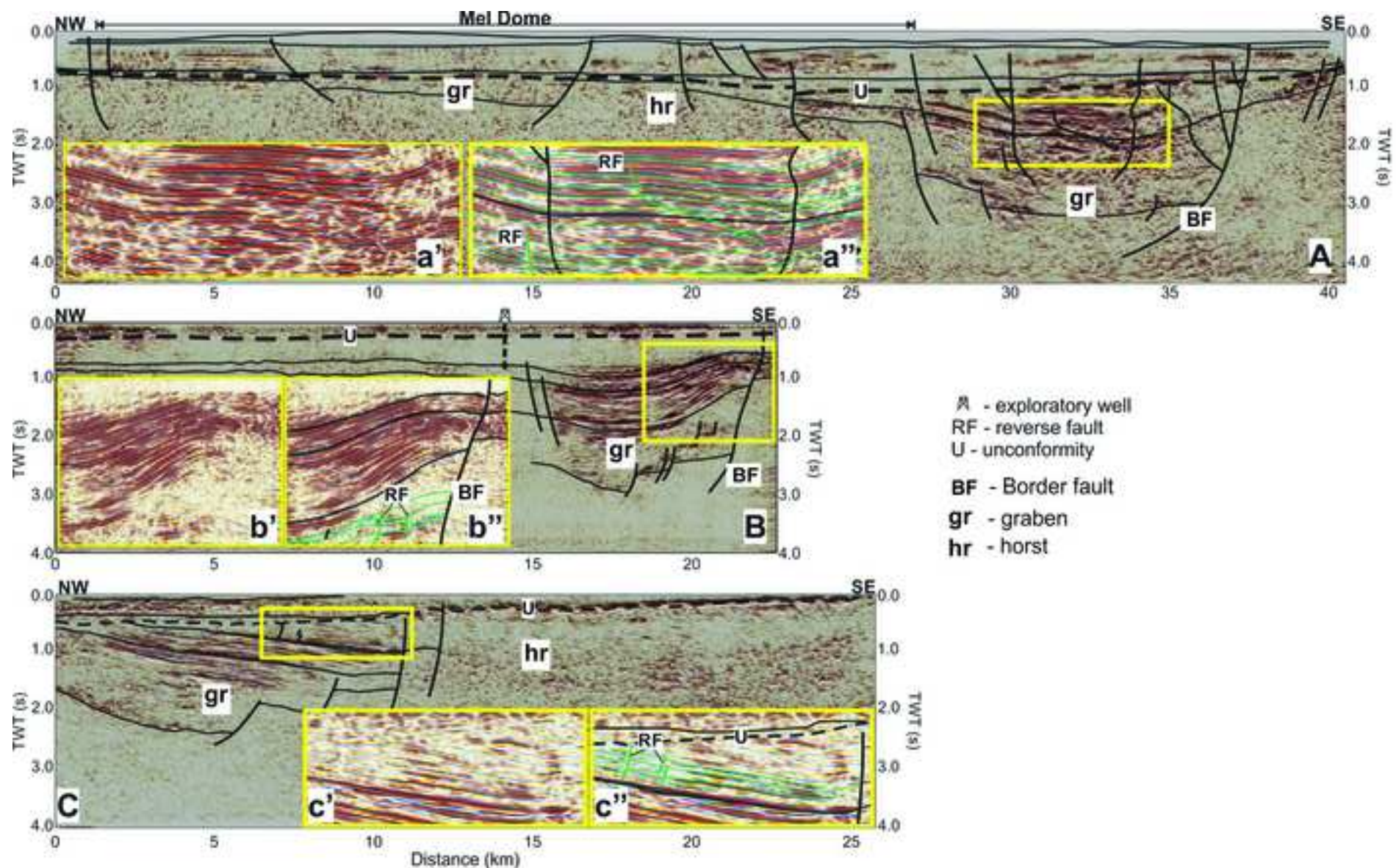


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Figure 10

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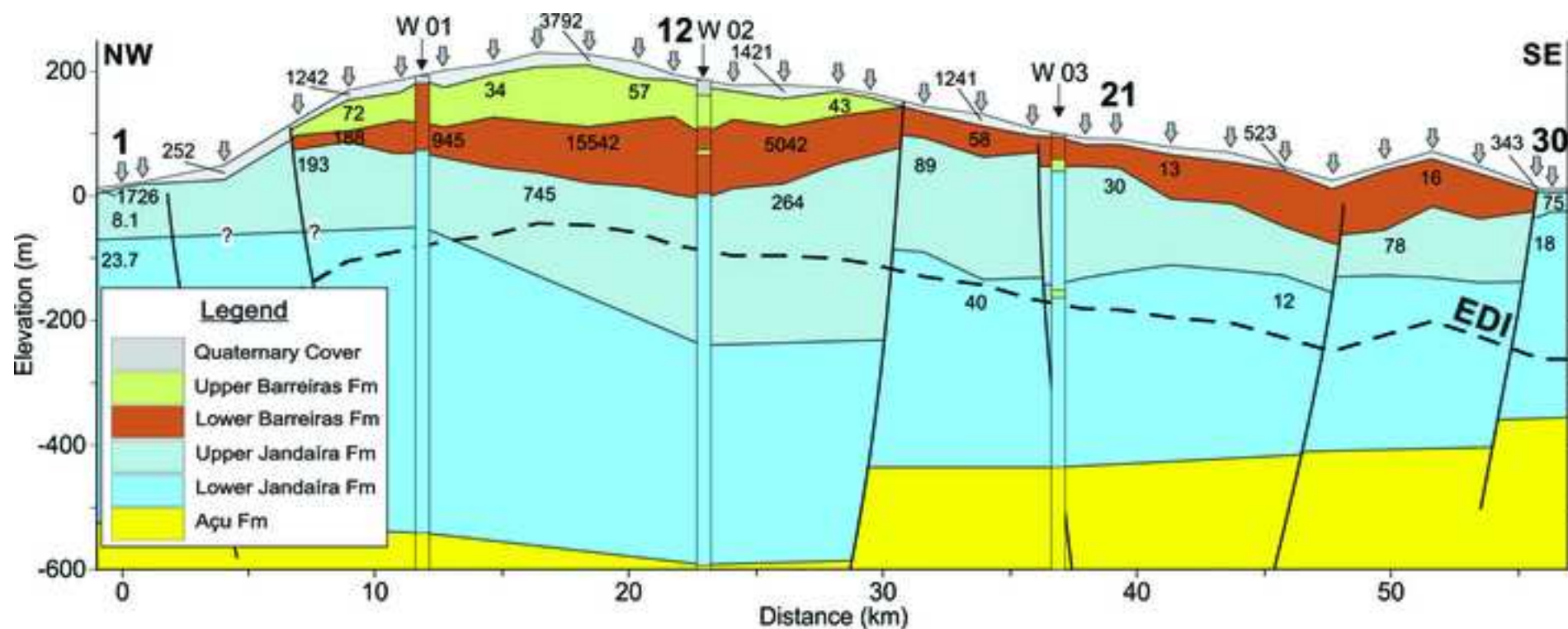


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Figure 11

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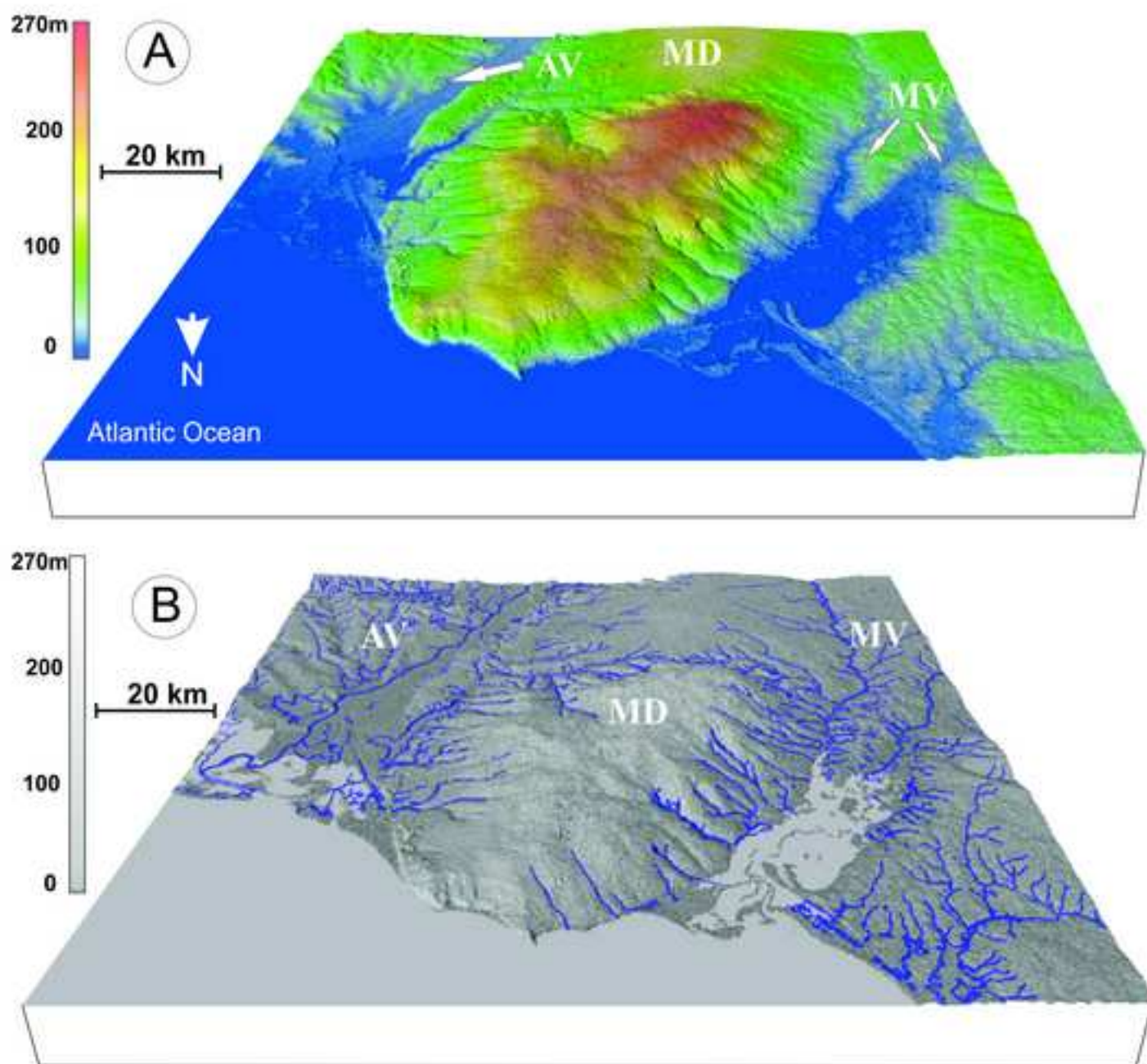


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Figure 12
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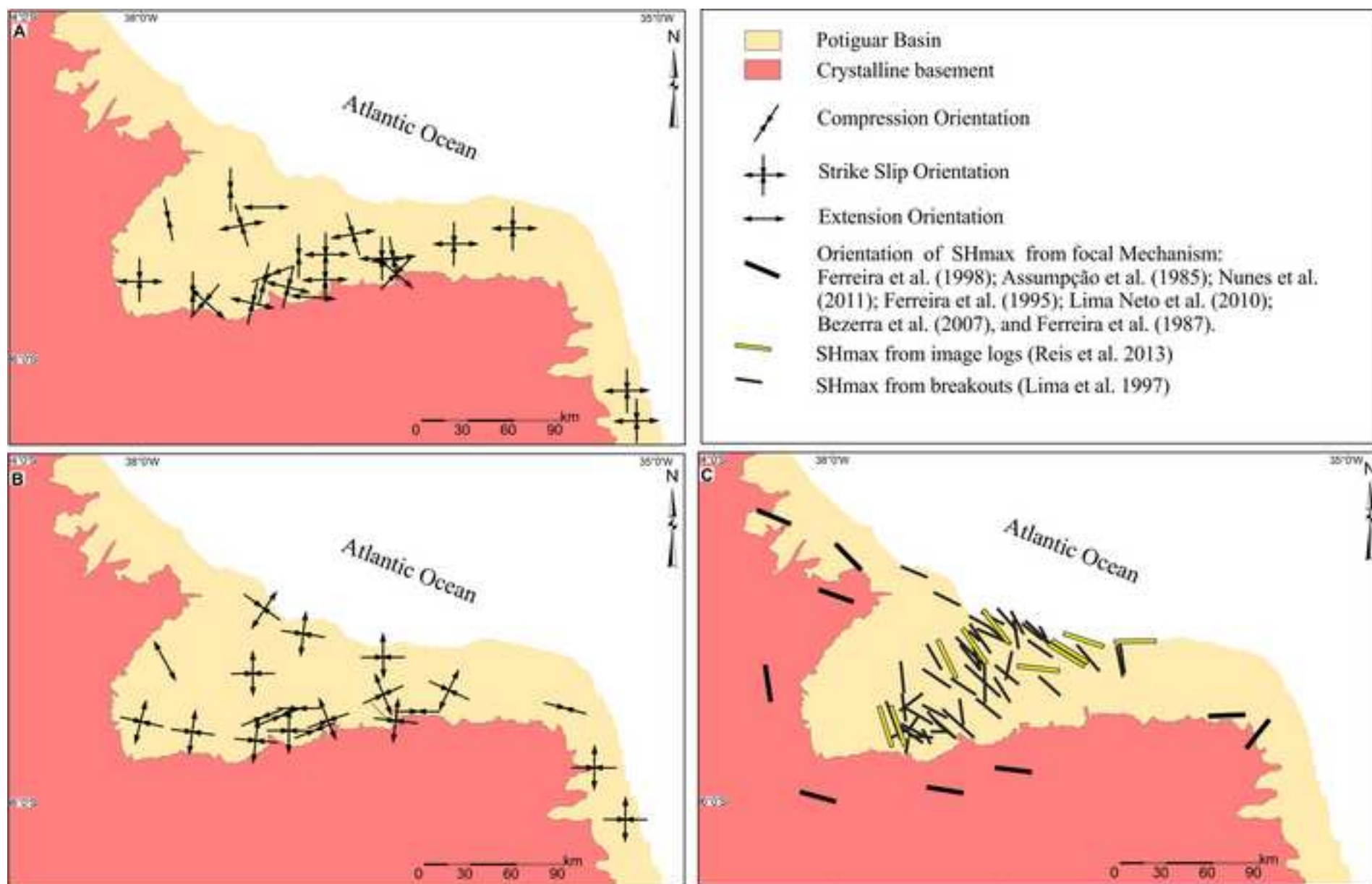


Figure 12 – Maps with the summary of the three stress fields, highlighting the maximum horizontal directions of compression and extension and fault kinematics. A) Stress field 1. B) Stress field 2. C) Present-day stress field.

Table 1 – Summary of K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages (2-sigma error) of Cenozoic basaltic rocks in the Potiguar Basin and adjacent basement. All ages are from whole rock samples, except for offshore sites, which are feldspar ages.

Site	Lithology	Age	Latitude South	Longitude West	Source
Onshore					
1	Basalt	42.0±3.0 ^(a)	05.08.00	36.30.45	Asmus and Guazelli (1981)
2	Basalt	14.2±1 ^(a)	05.10.38	36.42.49	Sial et al. (1981)
3	Basalt	18.0±1.0 ^(a)	05.20.40	36.20.15	Asmus and Guazelli (1981)
4	Basalt	18.0±0.5 ^(a)	05.27.20	36.18.00	Ebert and Brochini (1968) apud Sial (1976)
5	Basalt	12.5±1.0 ^(a)	05.28.12	36.19.12	Sial et al. (1981)
5	Basalt	13.0±1.0 ^(a)	05.28.12	36.19.12	Sial et al. (1981)
5	Basalt	13.9±2.0 ^(a)	05.28.12	36.19.12	Sial et al. (1981)
6	Basalt	24.0±6.0 ^(a)	05.30.25	36.20.25	Asmus and Guazelli (1981)
7	Basalt	8.9±0.5 ^(b)	05.38.34	36.20.01	Knesel et al. (2011)
8	Basalt	7.5±0.3 ^(b)	05.46.06	36.17.49	Knesel et al. (2011)
8	Basalt	7.6±0.4 ^(b)	05.46.30	36.17.48	Knesel et al. (2011)
9	Basalt	27.7±0.6 ^(a)	05.46.38	36.24.16	Sial et al. (1981)
9	Basalt	7.9±0.3 ^(b)	05.47.37	36.26.14	Knesel et al. (2011)
10	Basalt	20.0±1.0 ^(a)	05.40.10	36.20.00	Asmus and Guazelli (1981)
11	Basalt	24.6±0.8 ^(b)	05.42.05	36.19.15	Souza et al. (2003)
11	Basalt	23.7±1.2 ^(b)	05.42.05	36.19.15	Araújo et al. (2001)
11	Basalt	19.7±0.8 ^(a)	05.42.05	36.19.15	Cordani (1970)
12	Basalt	24.2±0.3 ^(b)	05.58.40	36.13.45	Menezes et al. (2003)
13	Basalt	31.7±1.0 ^(a)	05.58.00	36.21.49	Sial et al. (1981)
13	Basalt	30.4±1.0 ^(a)	05.58.00	36.21.49	Sial et al. (1981)
14	Basalt	7.4±0.4 ^(b)	05.54.38	36.27.14	Knesel et al. (2011)
14	Basalt	7.8±0.3 ^(b)	05.54.38	36.27.14	Knesel et al. (2011)
15	Basalt	29.7±0.6 ^(b)	05.58.00	36.21.45	Araújo et al. (2001)
16	Basalt	29.3±2 ^(a)	05.58.00	36.21.45	Sial et al. (1981)
16	Basalt	29.6±3 ^(a)	05.58.00	36.21.45	Sial et al. (1981)
Offshore					
19	Basalt	19.0±6.0 ^(a)	05.18.42	37.28.04	Mizusaki and Saracchini (1991), apud Mizusaki et al. (2002)
20	Basalt	28.0±7.0 ^(a)	04.54.54	36.17.59	Mizusaki and Saracchini (1991), apud Mizusaki et al. (2002)
21	Basalt	29.0±1.8 ^(a)	04.55.00	36.15.00	Fodor and McKee (1986)
21	Basalt	36.3±2.0 ^(a)	04.55.00	36.15.00	Fodor and McKee (1986)
22	Basalt	42.0±3.0 ^(a)	04.55.36	36.28.04	Mizusaki (1989)
23	Basalt	37.0±7.0 ^(a)	04.55.39	36.14.32	Mizusaki and Saracchini (1991), apud Mizusaki et al. (2002)
23	Basalt	37.0±9.0 ^(a)	04.55.39	36.14.32	Mizusaki and Saracchini (1991), apud Mizusaki et al. (2002)
23	Basalt	47.0±1.0 ^(a)	04.55.39	36.14.32	Mizusaki and Saracchini (1991), apud Mizusaki et al. (2002)
23	Diabase	44.0±4.0 ^(a)	04.55.39	36.14.32	Mizusaki and Saracchini (1991), apud Mizusaki et al. (2002)

24	Basalt	44.6 ±6.6 ^(a)	04.55.40	36.14.33	Mizusaki (1989)
25	Basalt	33.0±1.0 ^(a)	04.56.41	36.09.57	Mizusaki (1989)
25	Basalt	32.0±1.0 ^(a)	04.56.41	36.09.57	Mizusaki and Saracchini (1991), apud Mizusaki et al. (2002)

^(a) K-Ar age, ^(b) ⁴⁰Ar/³⁹Ar age.

Table 2 - Stress tensors from fault-slip data. site, outcrop location or outcrop groups shown in Figures 4 and 6; N^o = number of measurements used to compute the stress tensors; Phase = relative age of stress tensor; Unit age = age discussed in section 2.1 (Tur = Turonian; Cam = Campanian; Alb = Albian; Cen = Cenomanian; σ_1 , σ_2 , and σ_3 = dip direction and plunge of striae; R = relative ratio between principal stress axes; stress regime: SS = strike-slip; N = normal; R = reverse.

Site	N ^o	Phase	Unit age	σ_1	σ_2	σ_3	R	Stress regime
1	4	1	Tur-Cam	340/15	175/75	071/04	0.4669	SS
2	18	2	Miocene	094/01	197/84	004/06	0.0616	SS
3	15	1	Tur-Cam	354/05	107/78	263/11	0.3556	SS
	28	2		270/05	169/63	003/26	0.1258	SS
4	28	1	Alb-Cen	177/06	297/78	086/10	0.573	SS
	13	2		072/06	185/74	341/15	0.7847	SS
5	66	1	Alb-Cen	211/77	009/12	100	0.2948	N
	40	2		253/03	344/16	154/74	0.3567	R
6	39	1	Tur-Cam	008/04	098/04	232/85	0.794	R
	16	2		276/12	010/18	154/68	0.0481	R
7	11	1	Tur-Cam	356/00	266/02	086/88	0.3711	R
8	132	1	Tur-Cam	351/07	245/66	084/23	0.6267	SS
9	14	1	Alb-Cen	214/10	083/75	306/11	0.727	SS
	26	2		096/09	198/54	360/34	0.4333	SS
10	190	1	Tur-Cam	350/03	255/59	082/31	0.7876	SS
	8	2		070/17	263/72	161/04	0.6993	SS
11	26	2	Albian-Cen	267/07	171/43	004/46	0.7049	SS
12	74	2	Late Quaternary	273/08	129/81	004/06	0.3178	SS
13	53	2	Late Quaternary	090/04	224/84	360/04	0.5198	SS
14	64	1	Miocene	006/05	240/81	096/07	0.2498	SS
15	15	1	Miocene	357/06	158/84	267/02	0.5176	SS
16	39	2	Late Quaternary	103/00	193/02	007/88	0.9928	R
17	7	1	Tur-Cam	001/02	220/87	091/02	0.4459	SS
18	40	1	Miocene	021/81	186/09	276/02	0.8234	N
19	25	1	Tur-Cam	357/02	184/88	087/00	0.4212	SS
20	34	1	Alb-Cen	008/12	135/72	275/14	0.3508	SS
	22	2		269/02	027/85	179/04	0.4087	SS
21	21	1	Tur-Cam	343/06	074/06	212/82	0.5985	R
	11	2		079/11	342/24	192/63	0.2025	R
22	37	1	Alb-Cen	007/00	099/78	277/12	0.5421	SS
	21	2		279/00	009/76	189/14	0.4664	SS
23	39	1	Tur-Cam	012/09	281/05	162/79	0.1617	R
	54	2		257/02	347/04	139/86	0.7380	R
24	27	1	Alb-Cen	019/03	236/87	109/02	0.3313	SS
25	182	1	Tur-Cam	182/00	272/00	041/90	0.9810	R

	57	2		095/01	256/89	005/00	0.6945	SS
26	75	1	Tur-Cam	182/00	288/89	092/01	0.1336	SS
	37	2		107/03	274/87	017/01	0.8908	SS
27	17	2	Miocene	270/12	119/77	001/06	0.7584	SS
28	18	2	Miocene	060/00	327/87	150/03	0.1577	SS
29	51	1	Tur-Cam	167/03	077/08	274/81	0.2185	R
	14	2		039/88	240/02	150/01	0.0906	N
30	37	1	Albian-Cen	004/04	204/86	094/01	0.4825	SS
	56	2		277/17	109/72	008/03	0.9723	SS