Tectono-stratigraphic evolution of the rift and post-rift systems in the Northern Campos Basin, offshore Brazil

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Abstract
The Campos and Santos Basins have been a focus of subsurface studies since the discovery in 2006 of large accumulations of hydrocarbons in the Early Cretaceous rift and post-rift strata below Aptian evaporites. In this study, regional 2D seismic lines, a 3D seismic survey and well data were interpreted to reconstruct the tectono-stratigraphic evolution of the rift and post-rift stages in the northern sector of the Campos Basin. Detailed 3D seismic interpretation and geological modelling were used to subdivide the pre-salt sedimentary record. This revealed a diachronous strain distribution along the Guriri Fault System (GFS), a roughly NE-SW striking fault-bounded horst that was a focus of rift-related deformation. The syn-rift succession is interpreted to be the product of two episodes of rifting (RP1 and RP2) with contrasting fault activity patterns and lithostratigraphy. Volcaniclastics and coarse siliciclastics of RP1 were deposited under WNW-ESE transtension that formed discontinuous half-grabens followed by extensive erosion and tectonic inversion. Structurally controlled bioclastic rudstones and hybrid deposits characterize RP2 and were deposited in elongated half-grabens that delimit the GFS horst, which developed under an NW-SE extension. Deposition under increasingly less tectonically active transitional and post-rift stages dominated by thermal subsidence gradually led to the healing of the rift-related structural relief. Selective reactivation of rift faults testifies active tectonic inversion through compression from immediate post-rift throughout evaporite deposition. These interpretations are put into the context of recent geochronological data of onshore dyke emplacement and the new age constraints for the end of deposition of the pre-salt sequence. Therefore, we propose an earlier onset on rifting in the Campos Basin, at the Berriasian.

KEYWORDS
Campos Basin, early cretaceous, pre-salt, rift basin, sedimentation, tectonics

1 | INTRODUCTION
Several studies of rifting of the Campos Basin in the context of the break-up of the South Atlantic Ocean during the Early Cretaceous have been made using a combination of regional 2D seismic, wellbore and potential field data (Chang et al., 1992; Ponte & Asmus, 1976; Stanton et al., 2019). However, relatively few studies have been made using...
high-resolution 3D seismic data at the scale of segments of basin-wide faults that constitute the rift system (Alvarenga et al., 2021; Fetter, 2009; Olivito & Souza, 2020). Also, systematic structural mapping is only available for the Central sector of the Campos Basin (Guardado et al., 1989a).

Detailed petrographic and geochemical studies characterized the hydrothermal diagenesis in the rift transition to post-rift pre-salt carbonates in the North of the Campos Basin, with substantial impacts on reservoir quality (Herlinger Jr. et al., 2017; Lima et al., 2020; Lima & De Ros, 2019). These publications assumed that hydrothermal fluids moved upwards along faults developed during the rift phase. Hydrothermal fluid flow via geological structures is a subject of broad relevance in subsurface resources geology (e.g., Sharp et al., 2010; Walsh et al., 2018) and is controlled by several factors, from constitution and properties of the basement and sedimentary column (Davies & Smith Jr, 2006; Hollis et al., 2017) to tectonic regime (Sibson, 1992; Wiprut & Zoback, 2000).

However, the spatial organization and temporal constraints of the rift faults thought to work as conduits for hydrothermal fluids in the pre-salt succession are poorly understood. Also, the age constraints of the end of pre-salt and onset of evaporite deposition have been recently challenged (e.g., Pietzsch et al., 2020; Tedeschi et al., 2017), demanding new investigations on the tectono-stratigraphic evolution of the pre-salt succession and onset age of rifting.

To undertake the first step towards a more comprehensive understanding of tectono-stratigraphic evolution in the North of Campos Basin, it is essential to (1) constrain in detail the strain distribution related to rift faults in the three-dimensional space and their influence on deposition of syn-rift sequences and (2) describe the depositional architecture and the deformation in the post-rift sequences. Hence, the seismic architecture of syn-rift horizons and successions is interpreted and compared with some classic publications on tectono-stratigraphy of rift basins (Gawthorpe & Leeder, 2008; Lambiase, 1990; Martins-Neto & Catuneanu, 2010; Morley, 2002; Prosser, 1993). Also, studies on the Phanerozoic brittle reactivations and dyke emplacement in the adjacent onshore Precambrian basement (Calegari et al., 2016; Santiago et al., 2020) and magmatism in the Campos Basin (Almeida et al., 2013; Matos, 2021) are considered for a comprehensive understanding of the constraints of basin evolution. Such a comprehensive approach is necessary due to the poor geochronological control for the onset of rifting.

The aims of this paper are to:

1. Establish the geometry and kinematics of the rift-related fault framework and their later inversion;
2. Describe the rift and post-rift tectono-stratigraphic framework and

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**Highlights**

- The tectono-stratigraphic framework of the presalt sequence is investigated.
- The NE-SW Guriri fault system was the locus of the rift related deformation.
- The structural and stratigraphic frameworks are the result of two phases of rifting.
- The onset of rifting occurred during the Berriasian.
- Inversion tectonics took place towards the end of the post-rift stage and during evaporite deposition.

3. Discuss our interpretations in the face of the new time constraints for the deposition of the pre-salt succession and evaporites.

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**2 | GEOLOGICAL SETTING**

The Campos Basin is located in the Central Segment of the South Atlantic Ocean, which includes the contiguous Espírito Santo and Santos basins (East Brazil) (Moulin et al., 2010). These basins originated from rifting of West Gondwana during the Early Cretaceous (Chang et al., 1992; Rabinowitz & LaBrecque, 1979) and evolved as an initially magma-poor, then magma-rich break-up stage (Morgan et al., 2020). The Central Segment exhibits some remarkable tectonic features as follows: (1) southward widening of the basin margin and therefore a strong asymmetry of the necking zone and the continent-ocean boundary, (2) the role of the Santos and São Paulo Plateau as a kinematic ‘buffer’ between the Central and the Austral segments (Moulin et al., 2010; Araujo et al., 2022) and (3) anomalous rifting without the expected magnitude of rift-related subsidence and which led to the development of a long-lived extensive shallow carbonate platform and the subsequent salt deposits (Karner & Gambôa, 2007; Pietzsch et al., 2018).

The study area is located 70 km offshore, on the continental slope of the northern part of Campos Basin. The structural map of the shallow water domain of the Central Campos Basin shows major rift fault blocks bounded by alternating east- and west-dipping NNE striking normal faults, which are laterally continuous for long distances (up to 200 km) and have throws of as much as 2500 m (Guardado et al., 1989a) (Figure 1).

The rift system in the central and northern Campos Basin has its foundations in two distinctive basement units: the Oriental Terrane of the Ribeira Belt and the Cabo Frio Tectonic Domain (CFTD) (Schmitt et al., 2008;
The former is constituted by low-P high-T Neoproterozoic para- and ortho-derived gneisses intruded by several magmatic bodies, with high-angle NE-SW fabrics generated under a right-lateral transpressional regime as a result of polyphase amalgamation from ca. 650–500 Ma. The CFTD mainly comprises Paleoproterozoic orthogneisses (2.0–1.94 Ga) tectonically interleaved with medium to high T-P Neoproterozoic para-derived gneisses, both migmatized. The metamorphism and coeval ductile structures developed between 550 and 490 Ma due to a collisional event called the Búzios Orogeny (Schmitt et al., 2004). The CFTD has a predominantly low-angle NW-SE to N-S structural grain, orthogonal to the adjacent Oriental Terrane (Schmitt et al., 2004, 2016).

Two sets of fractures and dyke swarms striking NW to NNW transverse to the basement metamorphic grain in the onshore basement have their offshore projections close to the study area (Almeida et al., 2021): the NNW-SSE Vitória-Colatina lineament (Chang et al., 1992) and the Piúma/Itaoca (Castelo) lineaments (Lourenço et al., 2016; Santiago et al., 2020) (Figure 1). In common, these structures demonstrate a polyphase evolution, with alkaline dykes being emplaced during the Cambrian, and tholeiitic dykes in the Early Cretaceous age (see age compilation in Pessano et al., 2021). The dykes are interpreted as the result of the orogenic collapse of the Araçuaí Belt during the Cambrian and the opening of the South Atlantic Ocean during Early Cretaceous respectively. Therefore, they show recurrent episodes of brittle deformation of which precise timing and kinematic relations remain unclear (Calegari et al., 2016; Lourenço et al., 2016; Santiago et al., 2020).

A relative tectonically quiescent period of more than 300 Myr followed the Búzios Orogeny. From the Late Jurassic to Early Cretaceous, prior to the South Atlantic Ocean opening, several magmatic events were recorded in the region of Western Gondwana in which the Campos Basin formed. In the present-day onshore basement domain, early Cretaceous dykes along the Vitória-Colatina lineament yield Ar–Ar ages ranging 136–128 Ma and were emplaced under slightly dextral oblique extension (Almeida et al., 2013). The Castelo dyke swarm reveals a U–Pb Concordia age of 141 ± 1.9 Ma (Santiago et al., 2020). Both dyke swarms have geochemical signatures and ages similar to other Upper Jurassic/Lower Cretaceous dyke swarms of the Brazilian coast (Almeida et al., 2013) and to the basalt floods of the Paraná-Etendeka Large Igneous Province (LIP), of which ages ranges from 135 to 131 Ma (Matos, 2021 and references therein).

The rift system formed during the onset of the Campos Basin in Central South Atlantic (c.f. Moulin et al., 2010) occurred under an E-W main extension direction, which varies from orthogonal to oblique to the present-day coastal line.

![Figure 1](image-url)
resulting in a rift system in the northern Campos Basin. The extension became increasingly oblique southwards and reached its maximum in the ultra-wide rifted margin of the Santos Basin (Araujo et al., 2022). During the early rift phase, magmatism within the Campos Basin was characterized by mafic dyke injection onshore and extrusion of continental flood basalts of the Cabiûnas Formation from the Berriasian to Barremian (Mizusaki et al., 1992; Santiago et al., 2020).

The Campos Basin sedimentary record is traditionally divided into three tectono-stratigraphic supersequences: rift, post-rift and drift (Cainelli & Mohriak, 1999; Winter et al., 2007). The rift and post-rift supersequences span from Haueterian to Aptian and represent the pre-salt succession. It begins with a volcanlastic unit that represents the Cabiûnas Formation (Figure 2). Most relevant data from this unit came from shallow water drilling in the Central sector of the Campos Basin (Mizusaki et al., 1992). These authors demonstrated that Cabiûnas Formation tholeiitic basalts have similar bulk-rock chemistry and ages (K–Ar ranging 134–122 Ma) slightly younger than basalts of the Paraná–Etendeka LIP (135–131 Ma). However, Mizusaki et al. (1992) speculate that such younger ages found in the Cabiûnas Formation could be related to K2O addition during alteration, affecting the K–Ar ages.

**FIGURE 2** Compilation of the stratigraphic framework of the Campos Basin (not to scale). Ages and lithostratigraphy according to Winter et al. (2007) (1). Ages according to (Pietzsch et al., 2020; Sanjinés et al., 2022; Tedeschi et al., 2017, 2019) (2), where 128 ma is the age for the top of the Coqueiros formation. Horiz.: mapped horizons in this work: tSalt = Top salt, SBU = Salt Base Unconformity, IPRU = Internal Post-Rift Unconformity, PRU = Post-Rift Unconformity, tRift = Top Rift, tSRII = Top Syn-Rift II, tI/A = top Itabapoana plus Atafona formations, tBas = top Basement. T.S. (tectonic stages) are relative to the whole Campos Basin (Winter et al., 2007) (3) and to the deepwater domain of the North Campos Basin (Olivito & Souza, 2020) (4). Lithologies were based on Winter et al. (2007), Lima and De Ros (2019) and Olivito and Souza (2020). The 134 ma value is the oldest age obtained in the basalts of the Cabiûnas formation (Mizusaki et al., 1992).
The Lagoa Feia Group is stratigraphically above the Cabiúnas Formation and extends to the top of the evaporites of the Retiro Formation, embracing both rift and post-rift stages (Figure 2). It comprises five formations separated by regional and local unconformities, thus representing the primary seismic-stratigraphic markers (Winter et al., 2007) (Figure 2). The study area is located in the outer (distal) tectonic-sedimentary domains of the rift section of the Lagoa Feia Group, where isolated grabens were initially filled with siliciclastics to mixed carbonate deposits that evolved to isolated carbonate platforms built over structural highs, continuously in lacustrine conditions (Dias et al., 1988; Muniz & Bosence, 2018).

The rift stage continued with the deposition of coarse siliciclastics to shales under strong structural control, formally described as the Itabapoana and Atafona Formations, respectively. Towards the end of the rift stage, bioclastic rudstones and hybrid laminites represent the Coqueiros Formation (Winter et al., 2007). Recently, its facies associations were reinterpreted in a 3D context and the lower section of the Coqueiros formations was placed in the rift stage, while the upper section was classified as a transition rift stage (Olivito & Souza, 2020).

During the post-rift phase, the onset of thermal subsidence led to the environment of deep lakes with internal topographic highs being replaced by an extensive shallow, highly alkaline evaporative lacustrine system in which the Macabu Formation was deposited (Chang et al., 1992; Pietzsch et al., 2018). This shallow lake was in turn replaced by a highly evaporative shallow marine setting, leading to the deposition of up to 2000 m of evaporites of the Retiro Formation. Traditionally, the age of evaporites is ca. 113 Ma, at the top of Aptian (Winter et al., 2007) (Figure 2).

From the Albian, the deposition of the drift supersequence triggered remarkable thin-skinned gravitational gliding of salt and related salt compression in the outer regions of the Campos Basin (Davison, 2007; Szatmari et al., 2021).

### 2.1 Revised age dating of the pre-salt sequence

The Lagoa Feia Group ranges in age from the Barremian to the Late Aptian (Bertani & Carozzi, 1985a, 1985b; Thompson et al., 2015; Winter et al., 2007). Because of the endemic character of its fauna, it is challenging to make correlations with basins outside the Central Segment of the South Atlantic. Therefore, the depositional age of the pre-salt sequence had been constrained by radiometric data of underlying basalts of the Cabiúnas Formation (Mizusaki et al., 1992) and at its top by the age of the onset of evaporite deposition of the Retiro Formation (ca. 115 Ma) (Szatmari et al., 2021).

However, the assumption of a late Aptian age for the end of deposition of the Macabu Formation and the onset of salt deposition has recently been challenged by several authors (Lima et al., 2018; Pietzsch et al., 2020; Tedeschi et al., 2017, 2019) (Figure 2). New high-resolution carbon isotope ($\delta^{13}C$) profiles from shallow- and deepwater drillings in the Eastern Brazilian Margin (Sergipe-Alagoas, Campos and Santos Basins) were correlated with well-calibrated Tethyan sections. These new correlations constrain the end of major evaporite deposition in the South Atlantic to the early Aptian Oceanic Anoxic Event (OAE) 1a interval (124–125 Ma), thus the age of PRU as being 128 Ma (Tedeschi et al., 2019). Also, an increase in atmospheric $\mathrm{CO}_2$, predating the OAE 1a event, resulted in a change in the mode of the carbonate factory, recorded in the last tens of metres of the Barra Velha Formation in the Santos Basin (equivalent to the Macabu Formation in the Campos Basin). Hence, it puts further time constraints on the age of deposition of the top of the pre-salt sequence as not younger than Barremian (Pietzsch et al., 2020). Additional evidence came from new biostratigraphic records from wells drilled in the deepwater portion of the Campos Basin, where Aptian sections of up to 800 m thick occur above the salt (Lima et al., 2018; Sanjinés et al., 2022).

However, the new constraints for the age of salt thereof have been challenged by a review paper (Szatmari et al., 2021). Their results kept the age of salt deposition within the usual time frame (i.e., 116–110 Ma), based on $^{39}$Ar–$^{40}$Ar ages from evaporites in the Sergipe-Alagoas basin and a pre-salt volcano in the southern Santos Basin. The main argument of these authors is that the Tethyan section was located in a different realm than the Brazilian margin during the Early Cretaceous, thus making the correlation problematic.

### 3 METHODS

This study is based on commercial 2D (PSTM) and 3D (PSDM) seismic surveys and well data. Eight 2D semi-regional, dip-oriented deep seismic lines recorded from 8 to 12 s TWT provided control of the crustal structure and contextualized the 3D survey. This 3D survey, corresponding to the study area of this work, is a 3D streamer Pre-Stack Depth Migration (PSDM) seismic data consisting of a merge of two seismic surveys (see Strugale et al., 2021 for details). Well-logs and geological markers from 34 exploration and production wells named W01–W34 were calibrated with the horizon interpretation in the 3D survey. Sidewall cores from the few wells used to calibrate the
rift phase sedimentary units were invaluable in correlating seismic markers with the regional lithostratigraphic framework.

Six seismic horizons (BSU = Base-Salt Unconformity, IPRU = Internal Post-Rift Unconformity, PRU = Post-Rift Unconformity, tRift = Top Rift, tI/A = Top Itabapoana plus Atafona Formations and tBas = Top Basement) that crudely correspond to the formal lithostratigraphic units described in Winter et al. (2007) were extensively mapped throughout the 3D survey and represent unconformities or their correlative surfaces in the depocentres (Figure 2). They subdivide the pre-salt sequence into six (seismic) tectono-stratigraphic sequences (TSSs) comprising the syn-rift (SRI, SRII, SRIII), rift transition (RTR) and post-rift (PRI, PRII) tectonic stages. Local surfaces were also mapped between these horizons to identify changes in the seismic architecture. Faults, fold axes and erosional scarps were also mapped throughout.

The interpreted 3D framework of horizons and faults was converted to a 3D geological model using the Structural Model workflow in Petrel™, based on a 50 × 50 m mesh resolution. Depth, dip and dip direction attributes were extracted from every fault node to construct cross-plots and equal area lower-hemisphere stereographic diagrams (stereograms). True Vertical Thickness (TVT) isopach maps were generated from the model to evaluate thickness variations and evaluate fault-controlled accommodation. Edge effects due to lack of correlation at the margins of the 3D survey were considered during interpretation.

Faults were divided into subpopulations grouped by strike direction and structural domains (Fossen & Rørnes, 1996). Faults were also geographically grouped to represent genetically related structures and hence fault branches (high strain) and intra-fault zone domains (low strain). The 3D model was used to perform fault length versus displacement (L × D) analyses and automatic extraction of horizontal separation (heave) for each modelled horizon.

3D kinematic analysis of more prominent basement faults in rift settings has generally been applied where hanging wall depocentres are thick (typically >2 km), and a large number of syn-rift reflectors occurs (e.g., Collanega et al., 2019; Duffy et al., 2015; Lenhart et al., 2019). Ideally, mapped horizons must not represent erosive or highly condensed/non-depositional events since comparing depositional thickness across the faults is critical for understanding the evolving fault activity. However, underfilled half-grabens are challenging to assess since they often exhibit erosion of the footwall crest (McLeod et al., 2000). Also, prominent footwall highs, local and regional unconformities often merge into a few tuned reflectors making correlation challenging.

In the study area, fault scarps, folds, syn-kinematic and prograding wedges and local structural complexities pose further challenges when using these techniques. Taking account of these limitations, we opted for a more geometrical approach, where the hanging wall and footwall cut-offs of each seismic horizon and their respective seismic-depositional sequence were displayed in the map view, resulting in coloured polygons representing the heave for each stratigraphic unit (Figure 3a). Therefore, the heaves were analysed for either a single or a combination of horizons, resulting in tectono-stratigraphic maps on which the gross cumulative strain and spatio-temporal strain variations can be analysed (Figure 3b). However, this indeed takes no account of the possibility of underfilling in particular intervals, despite fault scarps occurring only in a handful of fault segments. Hence, for complex faults with significant across-fault correlation uncertainties, the cut-off lines were modified by additional seismic observation, like stratal terminations (onlap, erosional truncation and downlap), divergent syn-kinematic reflector and thickness variations, providing better time constraints of the fault activity (see faults ‘d’ and ‘e’ in Figure 3a).

4 | RESULTS

4.1 | Structural framework

The integration of 2D and 3D seismic structural interpretations from this study were correlated with existing structural maps of the Central Campos Basin (Guardado et al., 1989a, 1989b), to extend the documented structural framework to the northern limit of the basin. In the northern Campos Basin, normal faults striking crudely NNE-SSW, dip both east and west and constitute a similar rift-related fault framework to that mapped towards the south. Therefore, major structural features (i.e., the São João da Barra and Corvina-Parati Lows and the Badejo high) extend northwards into the study area (Figure 4a).

From a cross-section perspective, the pre-salt depocentres are younger towards the offshore, indicating the progressive rejuvenation of the rifting towards the spreading axis (Figure 4b). Another notable feature is the shallowing of the seismic Moho below the present-day shelf break where there are thicker rift deposits. This is interpreted as an early stage upper-crustal bondin of an originally thicker upper crust (c.f. Clerc et al., 2018). Also notable is the inverse correlation between the upper crust and the pre-salt sedimentary thicknesses (Figure 4b).

Over 170 normal faults were mapped within the 3D survey. The strike of faults has an average value of N40 with a significant population at N-S (Figure 5a). Dip directions are evenly distributed to E and W, although the faults with major displacements (up to 1250 m) are antithetic to
FIGURE 3  (a) Explanatory cartoon using six randomly picked faults (Fa to Ff) to explain the production of the tectono-stratigraphic maps. Seismic profiles across the structures are also shown to present the fault–horizon relationships and across-fault thickness variations. HW/FW cut-off lines were not entirely followed in faults ‘Fd’ and ‘Fe’. Fd is likely to represent more than one fault segment. In this figure, stratal terminations and dip indicators are merely illustrative. Coloured polygons represent the TSSs (see text for description). (b) Strain vs time distribution map according to the methodology shown in ‘a’ for a selected region of the studied area
FIGURE 4  (a) Semi-regional fault framework at top basement level, mapped from 2D seismic lines (red dotted lines) and 3D seismic survey (red polygon). The location map is based on Figure 1. The framework is correlated with the fault interpretation of Guardado et al. (1989a). The Vitória-Colatina lineament is projected from the onshore. (b) Interpreted semi-regional two-way time 2D seismic line showing the shallow- to deepwater transition. The crustal domains of the Campos rifted margin (c.f. Stanton et al., 2019) are shown, where NZ is the necking zone and HED is a distal highly-extended domain. The continental-oceanic crustal boundary is ca. 75 km to the east of the section limit. Note the shallowing Moho and the significant thickness of the pre-salt section in the shallow water. As depicted in the study area, the Guriri fault zone is discussed in this work. For the uninterpreted seismic section, see Figure A1.
the regional basin dip. The dip angle tends to increase towards the top of the pre-salt sequence (Figure A2b).

Fault distribution and patterns in map view were used to subdivide the area into three major structural domains. These correspond to a central, roughly NE-SW-oriented horst, flanked by two domains that are structurally lower at the basement level (Figure 5a). This horst is 10 km wide and is delimited by the major faults that are the strain loci of the area (Figure 4a). The central horst is here named Guriri Fault System (GFS). To the west of GFS, the structural low is denominated the Inner Guriri Low and to the east is the Outer Guriri Low. These three main structural domains exhibit internal deformation by subsidiary faulting, denoting internal structural highs and lows (Figure 5b).

The fault networks that focused the strain, thus acting to delimit the margins of the GFS are subdivided into three fault branches (GFSN, GFSE and GFSS) (Figure 5a). The GFSN and GFSE have fault segments dipping in a single direction (west and east respectively), while the GFSS exhibits structures dipping in both directions. Elsewhere, a footwall complex (FWC) represents the major positive structural element, flanked to the SSW by a depocentre (Low C). Within the Inner Guriri Low, two internal depocentres are observed (A and B), and the Outer Guriri Low has two internal lows (Low D and Low E) and one internal high (Well 33 High—W33H). L×Dmax analysis demonstrates that GFS branches focused the strain (Figure A3), and stereographic diagrams show that faults within the GFS branches have the main strike direction (NE-SW) that differs from the average direction of the faults elsewhere (NNE-SSW) (Figure A4).

The structural contours of BSU (base salt) and the isopach map for the pre-salt sequence are shown in Figure 5c,d respectively. Interestingly, the latter barely records the rift-related structural framework of Figure 5a, attesting to the deposition of a significant succession (i.e., the post-rift units) in a different tectonic setting.

The GFSN hosts the major faults in the 3D survey area, with >1000 m maximum displacements. The fault pattern in the map view is highly segmented with curvilinear fault traces whose strikes range between N-S and NE-SW, and dip systematically to NW. Also, dip angles decrease northwards along the strike, from 45° up to 25°, and splay into two segments towards the north (Figure 5a,b). The GFSN has the Low A as the hanging wall block, hosting the thickest depocentre in the area with up to 2.6 km of pre-salt sediments. Low A consists of three subsidiary hanging wall depocentres that are adjacent to zones of maximum displacement of the main segments of the fault zone (c.f. McLeod et al., 2000). The footwall of the GFSN (i.e., FWC) is internally faulted by NNE-SSW striking, oblique to GFSN and GFSE, mostly E-dipping faults with displacements <250 m (Figure 5b).

The GFSE is the eastward limit of the GFS and constitutes an array of quasi-parallel fault segments that are either hard- (breached relays) or soft-linked (Figure 5b). Segments dip systematically to the E with angles ranging from 45° to 70° and maximum displacements of up to 600 m. To the east, the Outer GFS Low has subsidiary depocentres in the hanging wall of the major segments of GFSE (Lows D and E) and an internal high (W33H), which has the thinnest pre-salt of the 3D survey (Figure 5d). Faults within Outer GFS Low have maximum displacements ranging from 200 to 400 m and are organized as a conjugate set whose intersection branch line is N25 oriented.

The GFSS is an apparent direct south-westward continuation of the GFSN branch but has a contrasting polarity of its main constituent faults, dipping predominantly SE. The east-dipping faults strike N-S to NNE and represent the structures with the largest maximum displacement values, of up to 500 m, that in turn delimit the half-graben of Low C (Figure 5a,b). During the course of the rifting, new west-dipping NE-SW striking faults nucleated westwards in an apparent southern extension of the GFSN branch.

We interpret a roughly NW-SE striking transverse structure to the GFS as an accommodation zone (c.f. Morley et al., 1990) (Figure 5a). This interpretation is based on two main lines of evidence. Firstly, an alignment of a number of breached relays and fault tips (Figure 5a,b), and secondly, a demarcation in changes in thickness distribution of the pre-salt tectono-stratigraphic units.

### 4.2 Seismic-stratigraphic framework

The primary hierarchy of the subdivision of the pre-salt sequence in the study area is two tectono-stratigraphic supersequences: syn-rift (SR) and post-rift (PR), separated by a rift transition (RTR) sequence. The terminology and interpretational diagnostics leading to this subdivision into supersequences were adapted from previous regional and local scales works (Cainelli & Mohriak, 1999; Olivito & Souza, 2020, respectively). In addition, internal unconformities and their correlative conformities allowed the subdivision of the supersequences into units with their own internal organization and termination patterns of seismic reflectors, lithological content and thickness distribution. They represent units that are lower in the stratigraphic hierarchy and therefore denominated as (seismic) tectono-stratigraphic sequences (TSS) (Figure 6).

Six TSSs were recognized; three representing the syn-rift supersequence (syn-rift I, II and III—SRI, SRII and SRIII), one is equivalent to the rift transition (RTR).
FIGURE 5  (a) Scheme of the major NE-SW trending structural domains showing the GFS (grey), their high strain branches (yellow) and the adjacent structural lows (inner and outer GFS lows). Simplified traces of the major fault segments within the GFS branches (black) and a stereographic diagram of all the mapped faults are also shown. (b) Top basement (tBas) structural map showing the local structural lows (A–E), their internal subdivisions (A1–3) and highs (FWC, W33H). Dotted lines are the external boundaries of the GFS branches as shown in ‘a’. (c) Base salt (bSalt) structural map. (d) Thickness map of the pre-salt sequence.
and two represent the post-rift supersequence: post-rift I (PRI) and post-rift II (PRII). The syn-rift TSSs show the diachronous distribution of divergent seismic reflectors (wedges) against faults and particular terminations of seismic reflectors (onlapses and truncations) against the delimiting horizons (tI/A, tSRII), justifying this subdivision. Also, the horizon between PRI and PRII sequences (IPRU) shows truncations and contrasting gamma-ray signatures between these TSSs (Herlinger Jr. et al., 2017).

The gross stratigraphy is well calibrated by well markers and sidewall cores. The following interpretations are based on geological cross-sections and thickness maps (Figures 6 and 7 respectively) that evidence the spatio-temporal changes in the distribution and internal organization of each of the TSSs. To illustrate the main lithotypes, sidewall cores from selected wells are displayed in Figure A5.

4.2.1 | The syn-rift supersequence

Syn-rift I TSS (SRI)
This unit is delimited at its base by a major unconformity (tBas) that separates the Precambrian basement from the sedimentary section. The top surface is also unconformable (tI/A), occurring as an erosive truncation in the structural highs and either as correlative conformity or onlapped surface in the depocentres (Figure 6a,c). In Low C, this unit defines classical divergent wedge geometry that is diagnostic of syn-depositional faulting (Figure 6c). Elsewhere, SRI commonly occurs as parallel layered internal reflectors, including in the hanging wall of some major GFSN faults (Figure 6a). Parallel configurations are commonly inferred as being pre-rift in a rift context, but this is not considered likely here. Instead, the parallelism is interpreted as low strain syn-rift, where distributed faulting provides the accommodation space, rather than a single major fault leading to pronounced fault block rotation and divergent geometry.

The SRI isopach map shows three NNE- SSE-trending depocentres in the centre of the 3D survey of which Low C is the westernmost and thickest, with >900 m of sediments (Figures 6c and 7a). Interestingly, the map distribution of the SRI thickness shows little correlation with the GFS branches (compare with Figure 5a).

Five wells penetrated to the tBas horizon (W01, 2, 4, 5, 34). The recognition and seismic interpretation of the tBas horizon in the study area is not straightforward and was investigated in detail by Strugale et al. (2021). Three wells found SRI basalts that sit on top of the basement when it is present (W04, 33, 34). Wells 33 and 34 drilled coarse siliciclastics over basalts, while in W04 bioclastic rudstones of the syn-rift III (SRIII) sequence overly SRI basalts. Wells W12, 14 and 25 only reached the top of the SRI and drilled medium to coarse siliciclastics with high feldspar content.

SRI is equivalent to three lithostratigraphic units, the Cabiúnas, Itabapoana and Atafona Formations (Figure 2). The basalt thicknesses in the wells are below the vertical seismic resolution (~90 m) and, therefore, the top of the Cabiúnas Formation was not mapped. Also, the seismic signal of SRI and available well data made any differentiation between the Itabapoana and Atafona Formations impossible.

Syn-rift II TSS (SRII)
The syn-rift II (SRII) sequence is delimited at its base by the tI/A unconformity and at its top by the tSRII horizon. SRII internal reflectors are parallel or onlap onto the tI/A near the active divergent wedges (Figure 6a,b), except for a localized truncation over the Low A1 (Figure 6d). The upper section of SRII is often conformable to tSRII in dip sections (Figure 6b). However, the strike cross-section shows that tSRII pinches out and is truncated towards the south (Figure 6d) and against the inverted Low C (Figure 6f). Deposition of SRII was coeval with substantial footwall erosion in the GFSN, thus constituting a fault scarp that was onlapped during SRIII (Figure 6a).

The main characteristic of the thickness distribution of SRII is its occurrence as a sub-basin (Low A) within the Inner GFS Low in the hanging wall of the GFSN (Figure 7b). SRII is absent by non-deposition or erosion elsewhere. Up to 1 km of sediments accumulate in this sub-basin, and they show remarkable divergent wedges within Low A2 (Figure 6a), where the GFSN has its maximum displacement. Within Low 2, a strike cross-section shows that the upper SRII thickens then onlaps onto the tI/A horizon towards Low A1 (Figure 6d). Elsewhere, parallel to tSRII, internal seismic reflectors are predominant (Figure 6a,b,d).

There is no well calibration for the SRII; therefore, its lithology will be interpreted in the context of the following SRIII TSS. The limited areal occurrence of SRII implies that the tI/A and tSRII horizons merge elsewhere as tSRII, consequently representing a significant gap in the sedimentary record (Figure 6c,e,f). However, it is not clear if the absence of SRII is the consequence of non-deposition, pre-SRIII erosion or both.

Syn-rift III TSS (SRIII)
The syn-rift III (SRIII) sequence is bounded at its base by both tI/A and tSRII surfaces and at its top by the tRift horizon. The tRift is widely distributed in the 3D survey as a mostly conformable surface and rarely shows evidence of erosion (e.g., Figure 6d). The base of SRIII onlaps the
FIGURE 6  Sets of dip (a–c) and strike (d–f) seismic-geological cross-sections showing selected stratigraphic and structural features. A simplified stratigraphic key is also shown (int. S., internal surface; T.S, Tectonic stage of Olivito and Souza (2020); TSS, Tectono-stratigraphic sequences; UNC., unconformities). White arrows indicate onlap, toplap, truncation and downlap terminations
tSRII, onto internal transverse faults in the hanging wall of GFSN (Figure 6a,d) and over the tI/A in Low C (Figure 6f). SRIII internal reflectors parallel to the base are rare and can be observed locally in strike sections (Figure 6d).

The SRIII package resembles the typical pattern of residual relief filling, progressively onlapping onto fault scarps and covering structural highs (Figure 6). Therefore, seismic reflectors are often parallel to the top. Syn-tectonic thickening also occurs related to selected segments of the GFSN (Figure 6b) and, more widely, along the GFSE and GFSS branches (Figure 7c). Elsewhere, a parallel internal organization is predominant.

The thickness distribution of SRIII follows the main Guriri Fault System (GFS) structural compartments. The
isopach map shows the configuration of thicker depocentres on both sides of the GFS horst, ranging from 400 m thick in the outer and up to 800 m thick in the Inner GFS Low. Interestingly, thicker deposits of SRIII in the Outer GFS Low occur near the GFS East (GFSE) branch, while it thins considerably elsewhere (Figure 7c). Along the strike of the GFS North (GFSN), thicker depocentres of SRI and SRII are located in the northern and central sectors of the GFSN branch, respectively, while SRIII thickens southwards, following newly nucleated NE-SW striking faults west of GFS South (GFSS) branch (Figures 6d and 7c). It demonstrates the overall southward strain migration along the strike of the GFSN and subsequently, GFSS, thus the overprinting of NE-SW segments over SRI N-S faults, that bounds Low C. Consequently, Low B was established and therefore configuring the horst geometry of the GFS (Figures 6c,f and 7c).

Another notable feature is the post-SRII erosion nucleation of subsidiary faults towards the footwall of the GFSN, which are parallel in the strike but steeper compared to the earlier segments of the GFSN. The timing of formation is endorsed by the thickening of SRIII (and rift transition—RTR) in the hanging wall block (Figure 6a,b). Major rift faults often exhibit erosion in the footwall that is usually related to footwall uplift and the formation of new faults as the consequence of the rotation of the faulted block, in a process known as footwall degradation, where faulting occurs either towards the hanging or the footwall blocks (Hesthammer & Fossen, 1999; Morley et al., 2007).

Wellbore calibration of SRIII is given by the nine penetrations (see Section 4.2.1.1 and W17, Figure 7c). Bioclastic rudstone (coquinas) is the most common lithotype in the structural highs, while low energy hybrid siltstones occur in depocentres (Olivito & Souza, 2020). The SRIII sequence is therefore correlated to the Coq A unit, the basal unit of the Coqueiros Formation according to those authors.

From the amplitude of the constituent reflectors and similar isopach distribution of SRII (seismic) tectono-sedimentary sequence (TSS) in the Inner GFS Low compared to the SRIII TSS, it is interpreted that both sequences are part of a single phase of rifting and probably share the lithological content. The GFSN branch exerts a fundamental role in controlling the creation of depositional space for both units (Figure 6a). The section equivalent to SRII has been included in the combined Itapaboa and Cabiúnas Formations, while SRIII represents the base of the Coqueiros Formation (Olivito & Souza, 2020). Because of the similarities thereof, SRII plus SRIII are here interpreted as belonging to the basal unit of the Coqueiros Formation and record syn-rift deposits (Figure 6b). GFSN activity prevails during the SRII and SRIII deposition and contrasts significantly with the SRI structure (Figure 7a–c). In addition, fault scarp formation during SRII denotes an active sediment-starved graben, which is typical in the early stages of rift filling, followed by generalized fill of the structural relief during SRIII (e.g., Prosser, 1993) (Figure 6a).

### 4.2.2 The rift transition sequence

The rift transition (RTR) TSS is delimited by tRift at the base and PRU (post-rift unconformity) at the top. The tRift is widely recognized as a conformable surface that is only occasionally onlapped by internal reflectors of RTR. The PRU has an erosive character towards the NE region of the survey that leads to top truncation and eventually the complete absence of the RTR in the easternmost portions of the study area (Figure 7d). Otherwise, PRU is also a conformable surface (e.g., Figure 6d).

Internally, the organization of the seismic reflectors are predominantly parallel except when submitted to erosion related to PRU, where the modest thickness compromised the resolution of the seismic signal. However, the proximal hanging wall block of the southern sector of GFSN shows a quite distinct pair of mound-shaped reflectors (Figure 7d). They have a circular to NE-trending ellipsoidal planform, with 3–5 km average length and exhibit downlapping terminations in the base and are onlapped by the post-rift units (Figure 6e). The mounds are not contemporaneous because the southern mound downlaps in the internal seismic reflector equivalent to the top of the northern mound (internal surface IRTR1—Figure 6e). However, a dip section shows that the mounds are comprised of wedges that prograde from the footwall complex onto the hanging wall of the GFSN (Figure 6b). This figure also shows that the depositional space occupied by the prograding wedge was created by newly formed NNE-SSW to NE-SW striking faults towards the footwall of the fault zone. These structures are amongst the few that were active during the RTR and are probably related to the ongoing footwall degradation observed during the SRIII. Finally, the geometry of RTR mounds is similar to the clinoforms observed in deposits in Santos Basin of similar age, except for the absence of syn-depositional tectonics (Barnett et al., 2021).

The isopach map of the RTR shows an overall thickening to the SW and absence of erosion in the eastern region as major features (Figure 7d). It also shows the imprint of the main structural compartments of Figure 5a, although the W33H internal high has a thinner RTR than the GFS horst. The two mounds are quite distinctive, hosting the thicker RTR with over 700 m. Tilting to the SW along the GFS exerts the primary control on the first-order thickness variations, so the RTR thickens towards the SW despite the GFS and their counterpart structural lows. The southward stepping of the mounds...
and erosion towards the NE supports the interpretation of such widespread tilting.

Several wells (34) reached or drilled through the RTR sequence, mainly in the footwall complex. The RTR is interpreted as equivalent to the upper section of the Coqueiros Formation, described by Olivotto and Souza (2020) as a ramp-margin shallowing-up sequence in modest water depths. These authors subdivided the upper section of the Coqueiros Formation into four sub-units (Cq B–E) and proposed their inclusion as a transition rift stage. The mounds are interpreted as berm and shoreface facies associations.

The relatively homogeneous isopach distribution and depositional style could be interpreted as resembling a post-rift architecture. However, the role of syn-depositional faulting remains significant on both sides of the GFS horst (Figures 6b and 7d). Therefore, the RTR is better explained as a transition sequence between clear syn-rift sequences (SRII and SRIII) and clear post-rift sequences (PRI and PRII).

4.2.3 The post-rift supersequence

The post-rift tectono-stratigraphic sequence has aggradational (parallel) to mound-shaped seismic facies and is characterized by a lack of any significant syn-depositional fault activity. It corresponds to the Macabu Formation that extends throughout the Campos and Santos basin, where it is named Barra Velha Formation, both deposited in an extensive sag-like basin geometry (Pietzsch et al., 2018). The Macabu Formation comprises shallowing-up cyclothems of mud-grade laminated carbonates, mm-diameter spherulitic carbonates and calcite shrub-like crusts, all deposited in a highly alkaline evaporative lake (Lima & De Ros, 2019; Wright & Barnett, 2015). Using well data (logs and cores) as a proxy for depositional facies, clay content, and therefore water depth, the overall stacking pattern is a deepening (post-rift I—PRI—tectono-stratigraphic sequence—TSS) followed by a shallowing-up (post-rift II—PRII—TSS) sequence (Herlinger Jr. et al., 2017; Muniz & Bosence, 2015); the IPRU (Internal Post-Rift unconformity) separates the two TSSs and, according to Herlinger Jr. et al. (2017), is equivalent to a third-order sequence boundary.

Post-rift II (PRII) TSS

The uppermost unit of the pre-salt sequence is delimited at the base by the IPRU and at the top by the base-salt unconformity (BSU—Figure 6). The PRII internal reflectors are usually parallel to the IPRU, except when it is downlapped by mounds (Figure 6a). The BSU is quite distinctive and reveals significant erosion in the hinge of fault-propagation folds and footwall blocks (Figure 6b,c,e) and within the Inner GFS Low (Figure 6d). The internal reflectors of PRII are usually parallel, but often onlapping onto the IPRU as tilting to SW remains active just after the PRI, and are commonly truncated at the top by the BSU. Internal surfaces PRII-3 are particularly abundant in the Inner GFS Low and suggest a cut-and-fill geometry (Figure 6d,e,f).

Mounded shape packages occur extensively over the upward projection of the GFSN branch and are apical to its faults related to footwall degradation (Figures 6a and 7f). Therefore, these mounds are symmetrical and pro-grade onto both footwalls and hanging walls of the fault segment that anchors them.

Apart from differences in terms of average and maximum values, the thickness distribution is similar to that of the PRI, i.e. it is divided into two major domains separated by an NNW-SSE accommodation zone and does not follow the major structural domains of the rift stage. The main difference between PRI and PRII thicknesses is the lack of broad post-depositional erosion towards the east in the latter. The relatively homogeneous thickness of the PRII shows an increasingly ‘sag’ depositional geometry in the form of...
wider depocentres compared to syn-rift and RTR sequences, preceding evaporite deposition (Figure 4). However, the present-day configuration of the pre-Salt, far from being horizontal and undisturbed (Figure 5c), points towards significant post-salt tectonism affecting the pre-salt sequence.

4.3 | Tectono-stratigraphic evolution

Having the gross structural and stratigraphic framework established, it is integrated into the tectono-stratigraphic maps (Figure 3) for each of the (seismic) tectono-stratigraphic sequences (TSSs—Figure 8). These maps focused on an area that includes the Guriri Fault System (GFS) and the near adjacent Inner and Outer GFS Lows. Fault displacement analysis and fold descriptions provided further input for the kinematic analysis that underpins this summary of the tectono-stratigraphic evolution.

The tectono-stratigraphic maps show that the rift-related subsidence was highly diachronous along the NE-SW striking GFS (Figure 8). They also show that the final horst geometry of the GFS had only developed towards the end of rifting. Thickness maps and the alignment of fault tips and changes in the strike of rift faults suggest the underlying and sometimes cryptic influence of NW-SE transverse structures (Figures 3b and 5b,d).

A comparison of the three TSSs of the syn-rift stage reveals similarities between the maps of SRI to PRII that make them distinct from SRI. Amongst the differences, we note the lithological contrast (volcanics plus coarse siliciclastics of SRI to carbonates of SRII plus SRIII TSSs—Figure 2) and, mostly, the contrasting structural pattern and therefore depocentre distribution (Figure 7a–c). Also, the significant time gap related to the tSRII unconformity since it represents the merge of two unconformable surfaces (tI/A and tSRII) in most areas should be considered. Finally, the inversion of the most significant depocentre of SRI (i.e., Low C) clear predates the SRII plus SRIII deposition (Figure 6c,f).

It is important here to distinguish between the inversion that predates the SRII plus SRIII, related to stress redistribution during the rift-related extension and best exemplified in the Low C, and the inversion represented by selective reactivation of rift faults during the post-rift stage as a result of compressional far-field stress. Interpretation of inversion was based on the occurrence of depocentres in a higher structural position than the neighbour blocks with respect to the newest stratigraphic unit and the observation of displacement variations along a fault on which normal displacement occurs in the deeper segment and reverse displacement occurs in the shallower segment.

Based on the contrasting strain and depocentre distribution, filling patterns and lithologies, we interpret the syn-rift supersequence as constituted of two distinct rift phases, where the rift phase 1 (RP1) comprises SRI, and rift stage 2 (RP2) comprises SRII combined with SRIII. Our subdivision of RTR, PRI and PRII closely follows the general classifications of rift transition and post-rift stages proposed by Olivito and Souza (2020) and Winter et al. (2007) respectively (compare Figures 2 and 6).

One of the main differences between our interpretation of the rift evolution in the study area from the previous study by Olivito and Souza (2020) is that these authors assigned a climax rift stage to the whole interval below the rift transition, in agreement with most of the previous publications (e.g., Winter et al., 2007) (Figure 2). Our interpretation is to replace the climax rift stage of Olivito and Souza (2020) with a syn-rift supersequence. On the other hand, we agreed with the term rift transition proposed by Olivito and Souza (2020), where the RTR mimics characteristics of both syn-rift and post-rift stages.

By the way of summary, we present our interpretation of the tectono-stratigraphic evolution of the pre-salt sequence in the North of the Campos Basin as a series of evolutionary schematic sections in Figure 9. It presents two cross-sections located north and south of the accommodation zone (Figure 5a), equivalent in location to the geological cross-sections of Figure 6b,c, respectively. Figure 9 emphasizes the general features of the sediment fill geometries and bounding surfaces. This figure demonstrates that the subsidence is often diachronous and heterogeneous from SRI to PRI, along and across the GFS horst, leading to sedimentary thickening contemporaneous to aggradational depositional patterns, and sub-aerial exposure thus erosion in each of the TSSs.

Based on the tectono-stratigraphic maps (Figure 8) and the evolutionary schemes of Figure 9, Figure 10 summarizes in map view the strain distribution and depocentre creation of each of the rift phases (RP1 and RP2) during the deposition of the syn-rift supersequence. The extension directions for each of the rift phases were established as being roughly orthogonal to the major active fault segments thus depocentre distribution. Figure 10 also shows the effects of the roughly NW-SE structure transverse to the GFS in the structural style and strain evolution during the syn-rift stage. While acting as a WNW-ESE-oriented kinematic buffer during RP1, the NW-SE structure is interpreted as an accommodation zone during the RP2 with close spatial relation with the Vitória-Colatina Lineament (Figure 4a).

4.3.1 | Rift phase 1 (RP1)

Rift phase 1 (RP1) took place over a crustal framework constituted of two distinct Precambrian basement
domains (Strugale et al., 2021). In Figure 11, a cluster of faults with minor displacements (~200 m) can be seen to the east (i.e., the Guriri Fault System East branch—GFSE), in contrast with sparse faults with major displacements (~400 m up to 900 m) to the west (GFSN-S), reflecting the duality in basement geology,

**FIGURE 8** Tectono-stratigraphic maps of the syn-rift sequences (SRI—a, SRII—b, SRIII—c) and rift transition (RTR—d) built according to Figure 3. Polygons represent the active faults for each tectonic stage of which the corresponding thickness map is shown.
FIGURE 9 Schematic geological cross-sections summarizing the tectono-stratigraphic evolution of the pre-salt sequence in the study area in the north (left) and south (right) of the NW-SE accommodation zone. The column on the right shows the main tectonic stages here interpreted. Black arrows represent local subsidence and uplift based on depositional geometries and thicknesses of each stage. The sections are not to scale and are vertically exaggerated. SBU: Sub-aerial unconformity/exposure (Discussion in the text)
whose implications for rift-related strain are discussed in Strugale et al. (2021).

Most of the early available accommodation space during syn-rift I (SRI) was concentrated in N-S to NNE-SSW elongated half-grabens filled with syn-tectonic sediments (Figure 10). Graben asymmetry is given by west-bounding faults, whose fault tips are often delimited by WNW–ESE lineaments (Figure 8a). Low C exemplifies these N-S striking half-grabens, with its divergent fill, two internal truncation surfaces and footwall erosion (Figures 6c and 9). Otherwise, most of the SRI internal geometry of seismic reflectors is parallel and therefore does not suggest syn-depositional GFSN faulting (Figures 6a,b and 7a).

Earlier deposition of the SRI is represented by the syn-tectonic volcaniclastic sediments of the Cabiúnas Formation. The lack of high-amplitude anomalies suggests a low volcanic/clastic ratio for this interval. This contrasts with the thicker Cabiúnas Formation and other correlated magmatic intrusions in the Central and Southern Campos Basin (Stanton et al., 2019). Once the grabens were filled, sedimentation spread more widely resulting in an aggradational filling pattern that tapered even the most prominent footwalls in the GFSS
This pattern suggests that an early phase of sediment-starved grabens was followed by the widespread creation of accommodation space and sedimentation. Following the terminology of Prosser (1993), the overall filling pattern of rift phase 1 suggests a cycle of syn-rift to late post-rift half-graben fill, in which the rift initiation system is not recognized. This interpretation, however, should be considered preliminary due to the low seismic resolution and lack of wellbore calibration in the depocentres.

Thickness and paleostress analyses in the SRI are somewhat difficult because the extensive erosion and localized inversions that preceded the RP2. The geometry of N-S to NNE-SSW striking half-grabens delimited by WNW-ESE lineaments suggests a WNW-ESE extension (Figure 10a) that either reactivated basement structures or nucleated new faults depending on the basement domains (Strugale et al., 2021) (Figure 11). In turn, the gentle obliquity between NE-SW basement structure and NNE-SSW rift faults in the FWC could suggest some degree of left-lateral kinematics on basement structures (Figure 10). The left-stepping geometry of the depocentres and fault tips suggests that extension was partially accommodated as right-lateral kinematics on the WNW-ESE transverse structure (Figure 8a). The contrasting thicknesses across this structure suggest that it acted as a hinge for local subsidence.

4.3.2 | Rift phase 2 (RP2)

After a period of relative tectonic quiescence and erosion/non-deposition, another pulse of fault activity occurred, initially restricted to the Guriri Fault System North branch (GFSN), followed by widespread faulting that was either newly formed or reactivated rift phase 1 (RP1) faults. Consequently, the GFSE was fully established, new W-dipping were formed on the GFSS and major faults.

FIGURE 11 Fault framework of early syn-rift (SRI) superposed on the basement geology map (modified from Strugale et al., 2021) comparing the fault distribution and the main basement units. Notice the obliquity between RP1 faults and basement structures in the FWC. The thickness map of the SRI sequence is also shown, on the left, along with the main boundaries of basement geology superimposed. CFTD, Cabo Frio tectonic domain; NProt, neoproterozoic; PProt, paleoproterozoic.
developed in the GFSN, resulting in major overprinting of rift phase 2 (RP2) over RP1 structures (Figures 8 and 10).

The first stage of RP2 deposition occurred during the syn-rift II (SRII) sequence in a mini basin (Low A) in the hanging wall block of the GFSN branch. Fault growth and linkage inherited from the previously E-W to NE-SW trending RP1 faults resulted in a fault zone whose main segments were sigmoidal (Figure 8b). Towards the syn-rift III (SRIII) sequence, thicker depocentres migrated southwards along the GFSN, from Low A2 to Low A3 away from the thicker SRII (A1 Low) depocentre (see Figure 7a–c). Such southward strain migration kept active during the rift transition sequence (RTR). This along strike strain migration suggests that the GFSN was part of a longer fault zone that extended northwards (out of seismic coverage) on which the strain had been propagating southwards until the rift transition phase (Figure 10). Also, the increasingly lower fault dip angles (<30°) northwards suggest that rotation of fault planes during extension led to the ‘locking’ of the fault (c.f. Reston, 2020) and therefore strain had to migrate southwards.

The filling pattern of the early RP2 (i.e., SRII) is characterized by classic sediment-starved/underfilled half-graben accompanied by divergent wedges, footwall uplift and erosion (i.e., fault scarp) in the position of maximum displacement of GFSN (Figure 6a). Such a pattern is reminiscent of the rift climax or through-going faulting stages of, respectively, Prosser (1993) and Gawthorpe and Leeder (2008) (Figure 9). Rift initiation or pre-rift layers are represented by RP1 deposits (Figure 6a,b,d). Discrete onlap of SRII onto the tI/A horizon (Figure 6a) suggests that fault rotation occurred prior to the onset of deposition, consistent with the unconformable character of the tI/A plus tSRII unconformity (Figure 6c,f).

A period of non-deposition within the Inner GFS Low occurred before the onset of SRIII deposition, as evidenced by the onlaps of SRIII onto the tSRII surface (Figure 6b,d). Sub-aerial exposure and erosion occurred elsewhere, tuning the seismic response of the tSRII and tI/A surfaces. These suggest that another episode of tectonic activity could have rearranged the depositional systems, leading to a long enough period of sediment starvation that allowed the deformation of the substrate that was subsequently onlapped by the SRIII (Figure 6d).

The depositional pattern of SRIII completed the tapering of the structural relief of SRII, exemplified by the wide occurrence of SRIII (Figure 7c) and its onlapping onto syn-SRII fault scarps and the footwall complex (FWC—Figure 9). Divergent wedges occurring against segments in the south of GFSN, in the GFSE, and west of GFSS (Figure 6a,b,c) imply that most of the GFS structures were still active in this interval.

The overall strain distribution during RP2 is relatively homogeneous, defining the present-day rift-related structural compartmentalization of the study area (Figure 5a). The tectono-stratigraphic maps shown in Figure 8b,c shows that the two major fault faults, GFSN and GFSE (southern part), are collinear but have opposite dip directions. This change in polarity occurs across the NW-SW accommodation zone (Figure 10).

The dominant NE-SW strike of the normal faults with major displacement and length is argued to be the result of an NW-SE to NNW-ENE extension during the rift phase 2 (RP2—Figure 10). Besides this rotation of the extension direction with respect to rift phase 1 (RP1), it was still slightly oblique to the basement structural grain, hence some left-lateral reactivation could still occur. However, a few but remarkable N-S faults were active during the syn-rift III sequence (SRIII) deposition, exemplified in Figure 6b and by an N-S graben within the NW-SE accommodation zone (Figure 8c). Therefore, we argue for a slight anticlockwise rotation of the extension direction toNNW-ENE during SRIII and a secondary right-lateral kinematics along the NW–SE trend (Figure 10). Figure 12 exemplifies the potential effects of such change in the extension direction in an area where two non-parallel major segments of the GFSN interact. Notice how the post-syn-rift II sequence (SRII) deformation by amplification of hanging wall folding as a consequence of localized shortening preceded and therefore is onlapped by SRIIII. This shortening is also noticeable during the rift transition (RTR) sequence as evidenced by the onlap onto folded tRift (Figure 12a).

4.3.3 | The rift transition phase

The rift transition phase is equivalent to the RTR sequence and marks the onset of thermal subsidence coeval with regional tilting to SW and localized faulting related to the ongoing displacement of rift phase 2 (RP2) faults. The isopach map demonstrates the interaction of the processes above, leading to the accumulation of remarkable carbonate mounds in the hanging wall of the Guriri Fault System North (GFSN) branch and an overall thickening of the RTR towards the SW (see Figure 8).

Active deformation during the RTR shown in Figure 8d is restricted to fault segments located within the GFS branches, most notably in the southern sector of GFSN (Figure 12). Strike and dip attributes resemble the later RP2 structures. The onset of SW tilting resulted in localized erosion of later RP2 deposits (i.e., SRIII) in structural highs, mainly in the Well 33 High (W33H) within the Outer GFS Low (Figure 7d). The extensive erosion that predates the post-rift deposition suggests increasing tilting during the transition to the post-rift and therefore explains the gentle onlap of both RTR and Post-Rift I (PRI) sequences onto the tRift and Post-rift Unconformity...
The prominent mounds that developed during RTR are anomalous against this background of uniform sedimentation and gently varying thickness. The geometry and spatial organization of the westward prograding mounded build-ups are the result of (1) southward along strike migration of strain locus in GFSN that (2) enhance the faulting related to footwall degradation, thus creating depositional space in the hanging wall demonstrated by prograding wedges (Figure 6b) that are (3) fed with sediments as the result of increasing substrate gradient by footwall (FWC) uplift. In the FWC, the deposition occurred in shallow conditions with events of sub-aerial exposure, thus working as a bypass zone (Olivito & Souza, 2020) (Figure 9).

We interpret that paleostress during the RTR resembles that observed in the latter part of RP2 (i.e., WNW-ESE extension and right-lateral kinematics in the NW–SE lineaments), which explains both NE-SW and N-S normal faulting (Figure 8d). Thicker RTR sediments south of the NW-SE trend could suggest a normal fault component with downthrowing block to the South (Figures 7d and 8d). Finally, the diminution of active faulting and the onset of tilting suggests that thermal subsidence had superseded the rift-related extension (Figure 9). The overall thickening to SW observed in RTR and in the post-rift I sequence is part of a wider depocentre that extends towards the Corvina-Parati Low (Figure 4a).

4.3.4 | The post-rift phase

The post-rift supersequence was deposited under a thermal subsidence regime, leading to an aggradational filling pattern that healed residual relief on the rift-related structures. Subsidence during Post-Rift I (PRI) was asymmetric and thickens westwards, while in the Post-Rift II (PRII)
is uniform, followed by sub-aerial exposure and erosion preceding the evaporite deposition (Figure 9).

There was some minor syn-depositional faulting during the PRI, but this was restricted to the active segments during the rift transition phase. The group of carbonate mounds are also anchored over the upward projection of underlying faults but shows no thickness variations across the structures (Figure 13). Towards the top of the PRII sequence and throughout evaporite deposition, inversion by selective reactivation of rift-related faults resulted in a set of fault-propagation folds (Figure 14). We interpret most of the fault-propagation folds in the study area as related to the inversion of normal faults through compression thought to be orthogonal to the faults. The main evidence is the hanging wall of the previously normal faults is higher than the footwalls (Figure 13c,d) (Konstantinovskaya et al., 2007; Letouzey et al., 1990; Withjack et al., 2002).

There were significant changes in the intensity of reactivation along the strike of GFSN. In the south, inversion is restricted to the hanging wall (Figures 9 and 13c), while inversion folds involved both hanging and footwall blocks towards the north, leading to widespread uplift of both GFSN and the adjacent footwall (FWC) (Figure 13b). SE dipping normal faults were apparently inverted in the FWC, which suggests the contribution of reactivation of Precambrian metamorphic fabric that resembles the dip and strikes direction of the rift faults in this location (Strugale et al., 2021). It also resulted in a monocline structure in the hanging wall block and the formation of up to 500 m of structural relief. The presence of a relict pod of salt on the western side suggests that the structural relief was active during salt tectonics (Figure 13b). Notice that the inversion-related folds have no correlation with salt thickness, therefore are not the result of errors during the time-depth conversion due to velocity variations related to variations in the salt thickness.

The interpretation of paleostresses related to tectonic inversion is problematic because of the thin sedimentary cover and strong coupling between inversion and the inverted faults. Therefore, newly formed structures may not directly reflect the new stress field but rather the structure that has been reactivated (c.f. Kelly et al., 1999). The orientation of the inversion folds at first suggests a simple NW-SE shortening. However, the framework of fold axes in the thicker southern half of Figure 13a shows two sets of NNE-SSW and NW-SE hinge lines, both deforming the bSalt horizon (Figure 13e).

The erosion in the hinge zones of fault-propagation folds (FPPFs) and truncations on the Post-rift II (PRII) sequence by the bSalt horizon is interpreted as the result of significant long-term sub-aerial exposure of the top of PRII, thus related to the base-salt unconformity (BSU). Also, a folded geometry is observed towards the top of the salt in some localities (Figure 13c,d), implying that folding was active for a significant time after the onset of evaporite deposition and potentially until the onset of the salt tectonics in the Albian (Cobbold & Szatmari, 1991; Fetter, 2009). In addition, assuming that the inversion leads to localized (e.g., segments of Guriri Fault system East—GFSE) or generalized (GFSN and footwall complex—FWC) uplift, this could have delayed the onset evaporite deposition over these areas. Consequently, using the basin scale correlations to estimate the age of the onset of evaporite deposition in the context of the study area could be problematic (Davison, 2007; Kukla et al., 2018; Szatmari et al., 2021). Meanwhile, it is possible that there was significant sub-aerial epigenetic karstification related to the BSU, resulting in secondary porosity in the upper section of post-rift carbonates. The occurrence of epigenetic karst on top or within carbonate sequences is well known in this type of context (e.g., Saller et al., 1994).

5 | DISCUSSION

5.1 | Strain migration during rifting

The shallowing of the Moho below the present-day shelf-break of the Campos Basin has been observed in several publications (Matos et al., 2021; Meisling et al., 2001; Mohriak et al., 2008; Stanton et al., 2019). Matos et al. (2021) show that such uplift correlates with a thicker pre-salt package, referred by Figure 4b. The location of Moho uplifts in the SE Brazilian margin is considered to be related to lateral changes in the lithospheric thickness inherited from Brasiliano orogeny assembly that allowed later along rift-axis flow of plume material (Morgan et al., 2020). The presence of an important basement tectonic limit near the study area (i.e., tectonic limit between the Oriental Terrane of Ribeira belt and the Cabo Frio Tectonic Domain—Figure 1) (Stanton et al., 2019; Strugale et al., 2021) could have worked as such a lithospheric limit. Hence, a longer residence of plume material could result in lithospheric thinning and extension thus strain localization, evidenced by the development of major rift faults and thicker depocentres such as the Corvina-Parati and São João da Barra Lows in the shallow waters of the Central and North Campos Basin (Figure 4).

Rift basins that evolve to oceanic crust are characterized by strain diachronity along dip and strike, meaning that the strain locus migrates towards the propagation centre (Lavier & Manatschal, 2006). Consequently, rift units in a distal position will correlate to post-rift units towards the
onshore (Figure 4b). The strain migration is not uniform and will depend, amongst other factors, on crustal heterogeneities. The occurrence of a shallower Moho under thick rift deposits could be interpreted as a proto spreading axis and therefore a strain locus during the rift phase 1 (RP1), located in the present-day shallowwater domain of the Campos Basin. During the rift phase 2 (RP2), the strain locus moved slightly eastwards (Figures 3b and 10) before moving rapidly offshore (‘jump’) as a consequence of a thick boudin of the upper crust (c.f. Clerc et al., 2018) formed during RP1 strain localization. This buoyant boudin of the upper crust has clear spatial correlations with thinner RP2 and post-rift deposits and the occurrence of internal unconformities in the eastern half of the study area (Figure 4b).

The role of basement-inherited NW-SE brittle fracture and dyke swarms observed in the onshore Precambrian basement on the evolution of the rifting in the offshore domain is still a matter of debate. While some publications consider they are fundamental (Alvarez et al., 2021; Fetter, 2009), others do not recognize their expression offshore (Lourenço et al., 2016; Stanton et al., 2019). The regional map shown in Figure 4a infers the projection of the Vitória-Colatina lineament transecting the study area.

FIGURE 13 Tectono-stratigraphic maps of the post-rift TSSs (PRI—a, and PRII—b). Thickness maps of each phase are superposed by polygons representing the active faults, fold axes, stratal dip and termination geometries extracted from 3D seismic. Mounds and erosive features are also shown. Fault traces of PRII (b) are shown with dotted borders since they do not displace the hSalt horizon.
FIGURE 14  (a) Map of folds distribution, showing their geometry and classification according to the age of formation. Dotted black lines represent the GFS branches and dotted grey lines are the accommodation zone (Figure 5a). Notice the SRIII hanging wall folds in the lows A1–3 related to displacement variations on the GFSN (c.f. Schlische, 1995). (b) GFSN footwall-crest fold with significant proximal hanging wall deformation and possible inversion (double arrowed displacement indicator). Notice the constant thickness of the post-rift across the GFSN, suggesting post-depositional deformation. (c) Fault-propagation folding (FPF) in the hanging wall of the GFSN related to reverse kinematics due to inversion, leading to uplift, erosion and scarp development. (d) FPF over a fault segment of the GFSE. Notice the erosive truncation in the top of the RTR and within the PRII sequence and inferred reverse faults. (e) FPF over a segment of the GFSE, a transverse fold (centre of the section) and a differential compaction fold to the west, probably amplified due to later shortening as evidenced by the deformation of the bSalt horizon.
Detailed 3D seismic interpretation identified an active NW-SE accommodation zone during the RP2 near the position of this regional lineament (Figures 8, 10, and 14). Similarly, a WNW-ESE trend worked as a kinematic buffer during RP1, which can be correlated with the Castelo Lineament (c.f. Almeida et al., 2021) (Figure 10). Detailed 3D seismic mapping of the pre-salt sequence (this work) and of the basement (Strugale et al., 2021) did not recognize any discrete NW-SE structure. Therefore, assuming the NW-SE and WNW-ESE lineaments here identified as being correlative to either Vitória-Colatina or Castelo lineaments onshore is still speculative, especially the second.

The inversion structures resulting from selective reactivation of NE-SW rift faults due to orthogonal compression observed towards the end of post-rift supersequence and onwards are best explained by intraplate stress propagation from oceanic mid-ridge spreading (c.f. Artysuhkov, 1973). Such a mechanism had been correlated with Late Cretaceous and Tertiary inversions in the Campos Basin, competing with far-field compression from the Andean Orogen (Fetter, 2009). Considering the estimated onset of oceanic crust generation in the Campos Basin as ca. 117 Ma (Heine et al., 2013), the observed inversion in Figures 13 and 14 could be correlated with the onset of oceanic spreading thus ridge pushing.

5.2 Rift and post-rift tectono-stratigraphic evolution

The filling architecture of rift basins in both natural and experimental (physical and numerical) is the subject of numerous publications. Amongst them, 2D and 3D models for tectono-sedimentary evolution of rift basins in general (Gawthorpe & Leeder, 2008; Prosser, 1993), for continental lacustrine environments (Lambiase, 1990; Morley, 2002), and the model for rift sequence stratigraphy of Martins-Neto and Catuneanu (2010) will be compared with our interpretations. In common, these publications point to the difficult task to understand the feedback between tectonic-related and purely eustatic-driven sea/lake level changes. This is especially problematic in rift basins where the fault-controlled depocentres form diachronously and the accommodation space in lacustrine settings is subject to local climate constraints, thus disregarding global eustatic curves (Catuneanu, 2006).

Although the references thereof are relevant contributions to understanding the evolution of rift basins they also proposed denominations for the stages and patterns of rift filling that are far to be homogenous. The main point is that they mostly rely on the interpretation of regional 2D seismic lines to set up the models. However, our interpretation is founded on a 3D seismic survey at the oilfield scale, which differs from the models mentioned. Hence, each geological cross-section here presented would fit (or not) in each of the models used as references (Gawthorpe & Leeder, 2008; Lambiase, 1990; Morley, 2002; Prosser, 1993). For example, in the context of the RP2, the syn-rift II sequence (SRII) could represent either the rift climax system (Prosser, 1993), interaction and linkage stage during highstand in a lacustrine or shallow water context (Gawthorpe & Leeder, 2008) and early half-graben stage (Morley, 2002). Our interpretations clearly show that there is no silver bullet in the literature and therefore is pointless to put effort into trying to fit our results to any individual or a combination of these classifications.

We tentatively compared our findings with the few publications focused on the petrological–sedimentological content and the 2D seismic-stratigraphic architecture of a single half-graben in the shallow water domain of the Central Campos Basin (Alvarenga et al., 2021; Goldberg et al., 2017). Three syn-rift system tracts represent the rift tectono-sedimentary record as interpreted by Alvarenga et al. (2021). Their basal unit (Rift Initiation) could potentially be correlated with RP1, and the upper High and uppermost Low Tectonic Activity Systems Tracts could be correlated with, respectively, SRII and syn-rift III (SRIII) (i.e., the RP2). Alternatively, the Low Tectonic Activity System Tract (LTAST) could be correlated with the Rift Transition sequence. However, this latter correlation is problematic because (1) LTAST is wedge-shaped and is similar to either SRII or SRIII, (2) the authors interpret significant LTAST syn-tectonic gravitational deposits (Goldberg et al., 2017) and (3) the lack of data and interpretation in the footwall block does not allow correlations with the graben fill and a clear understanding of how fault activity evolves towards the end of rifting.

From that, our choice is to use the strain distribution and evolution tied with a well-constrained seismic-stratigraphic framework, which is only possible if preceded by high-density 3D seismic mapping (e.g., Figure 5). A dense mapping of seismic horizons, faults and stratal terminations against them, followed by integration with isopach contours to build tectono-stratigraphic maps (Figures 3, 8, and 14), was the main criteria to distinguish two phases of rifting (RP1 and RP2) (Figures 9 and 10). The superposing of snapshots of the strain distribution along time was the best strategy to unveil the complex strain evolution. Such an approach could fill the gap originated in the biased sampling by wellbore data (i.e., wells were drilled only in the structural highs), especially regarding the lack of rock data of gravitational and reworked deposits in the regions where syn-sedimentary tectonism is expected to be high.

Our interpretations reveal that the changes in strain distribution during the rifting are the product of two
directions of extension (E-W during RP1 and roughly NW-SE during RP2) (Figure 10). PR1 is characterized by small and isolated grabens, with preserved thickness up to 900 m, while RP2 evolves to longer grabens that host up to 1800 m that often did not take advantage of RP1 faults (Figure 8a–c). In addition, RP1 depocenters were inverted while new faults nucleated or linked from initially isolated RP1 faults (e.g., GFSN), resulting in the final horst geometry of the GPS. Such an overall pattern resembles the models of rift basins where they are initially characterized by isolated grabens with small displacements, followed by fault interaction and linkage, leading to the formation of larger grabens (Gawthorpe & Leeder, 2008; Gupta et al., 1998). These models, however, do not necessarily mean a change in extension directions and do not mention different rift phases.

An additional argument for two phases of rifting is the contrasting lithologies, from volcanioclastics plus coarse siliciclastics (RP1) to carbonates and hybrid siltstones (RP2) (Figure 9). Also, the merge of two unconformities (tI/A and tSRII) into a single surface in most of the area is meaningful for a significative gap in the sedimentary record between RP1 and RP2, which is not observed for any other mapped horizon (Figure 6).

Superimposed rift phases have been observed in rifts and are related to changes in the stress field during their evolution (Bosworth, 1994; Lambiase, 1990; Morley, 2017). In examples from the North Sea and Sudan, rift phases are directly superimposed or have a ‘lag’ phase in between (McHargue et al., 1992; Phillips et al., 2019). Multiple rift phases are recognized in the NE Brazilian margin and within aulacogens in the counterpart Borborema province (Matos et al., 2021). However, the stratigraphic charts of the Campos Basin and the adjacent Santos and Espirito Santo basins (França et al., 2007; Moreira et al., 2007; Winter et al., 2007) only recognize a single rift phase. Therefore, our proposal for two rift phases for the Campos Basin is a significant turning point for the knowledge of the basin. We must recognize that such an interpretation is not novel (see Mohriak et al., 2008) but had been neglected in most publications afterwards.

5.3 | Implications for the age of the pre-salt sequence

The interpretation of two phases of rifting in the northern Campos Basin here presented shall be confronted with the new age constraints for the end of the pre-salt deposition, the onset of evaporite deposition and the post-salt sedimentary column (Lima et al., 2018; Pietzsch et al., 2020; Sanjinés et al., 2022; Tedeschi et al., 2017, 2019). Traditionally, the pre-salt sequence in the East Brazilian margin is interpreted as a single rift phase plus post-rift succession spanning 19 Myr in total, from Hauterivian (134 Ma) to Late Aptian (ca. 115 Ma) (e.g., Mohriak et al., 1990; Winter et al., 2007). Recently, the age of the end of pre-salt deposition was interpreted as not younger than Barremian (125 Ma) (Pietzsch et al., 2020) and the age of evaporite deposition as being Early Aptian (ca. 124 Ma) (Tedeschi et al., 2017, 2019).

Although the age of evaporites is still under debate (see Discussion in review paper of Szatmari et al., 2021), the problem remains: how to accommodate two phases of rifting (instead of one), a rift transition and a post-rift stages in a time interval ca. 10 Myr shorter than previously assumed?

The first option is to challenge the age of the onset of volcanism assuming it as a proxy for rift-related mechanical subsidence. In the onshore basement counterpart to the northern Campos Basin, tholeiitic dykes emplaced along NW-SE lineaments are dated as being as old as 142 Ma—Berriasian (Santiago et al., 2020) and 136 Ma—Valanginian (Pessano et al., 2021), in close geochemical and temporal relationship with the Paraná-Etendeka LIP and to early manifestations of the break-up of the South Atlantic Ocean (e.g., Matos, 2021). Also, several publications show evidence of Late Jurassic to Valanginian rift-related deposition along the proto Central Segment of the South Atlantic Ocean. For example, the Afro-Brazilian depression in Northeast Brazil (Kuchle et al., 2011; Matos et al., 2021), and Kimmeridgian (L. Jurassic) rift units at the base of the Camboriú Formation in the Santos Basin, equivalent to the Cabiúnas Formation (Alves et al., 2020).

A simple though not definitive solution would be considering the life span of the pre-salt envelope as being the previously assumed 19 Myr (Winter et al., 2007), and therefore, the age of the onset of rifting would be at ca. 144 Ma. In this assumption, we are contesting the oldest ages of the volcanic of the Cabiúnas Formation (i.e., 134 Ma). This assumption is partially supported by the mean age of emplacement of onshore dykes (ca. 142 Ma in the Castelo swarm, according to Santiago et al., 2020), which is here considered as a proxy for the onset of mechanical subsidence. The tholeiitic composition of these onshore dykes is the same as the Cabiúnas volcanioclastics (Mizusaki et al., 1992). However, although this interpretation kept the pre-salt rifting plus post-rift within the global dataset of the life span of rift basins (ca. 20 Myr, Woodcock, 2004), the assumption of dyke emplacement and volcanism as a proxy to the onset of rifting is problematic, especially in the context of a magma-poor margin (Morgan et al., 2020). Therefore, an investigation focused on obtaining further radiometric ages in the pre-salt basalts, especially in the northern Campos Basin, should be considered in future works.
A secondary possibility to adjust the ages of the pre-salt sequence in face of new age constraints would be to challenge both the recently proposed Barremian age for the end of the pre-salt deposition (Pietzsch et al., 2020; Tedeschi et al., 2019) and the Early Aptian age for evaporite deposition (Tedeschi et al., 2019); both views are contested in the review paper of Szatmari et al. (2021). From that, our understanding is that there are good reasons to consider the correctness of the 125 Ma for the end of pre-salt deposition. This assumption is reinforced by the thick Aptian post-salt sequence described in the Campos Basin (Lima et al., 2018), later studied in detail in the SE Brazilian Margin in Sanjinés et al. (2022), who concluded that the Aptian–Albian border is 10 Ma older than previously assumed (Figure 2). If the onset of evaporite deposition occurs between 116 and 110 Ma as suggested by Szatmari et al. (2021), it means an at least 9 Myr gap in the base-salt unconformity. Nevertheless, these authors mention the diachronous character of evaporite deposition in the proto-South Atlantic Ocean (see also Farias et al., 2019). In the northern Campos basin, we observed significant erosion on the top of the Macabu Formation (Figure 9). A similar observation was made in the shallow water domain of the Central Campos Basin (Alvarenga et al., 2021; Goldberg et al., 2017). Such a long sub-aerial exposure (9 Myr) could imply the occurrence of extensive epigenetic karstification on the uppermost section of the pre-salt sequence, which was never reported in the literature on the Campos and Santos Basins.

6 | CONCLUSIONS

The tectono-stratigraphic record of the pre-salt of the northern Campos Basin is subdivided into syn-rift supersequence, rift transition (RTR) sequence and post-rift supersequence. The syn-rift is composed of three (seismic) tectono-stratigraphic sequences (TSS SRI, II, III), and the post-rift supersequence encompasses two TSSs (PRI, II), with a rift transition sequence (RTR) in between. Sub-aerial unconformities and their correlative conformities separate them.

The main structural compartments are represented by an NE-SW-oriented horst that encompasses high and low strain domains, constituting the Guriri Fault System (GFS). It is constituted by three high strain branches of normal faults named north, south and east (GFSN, GFS and GFSE respectively) and a footwall complex adjacent to GFSN. The GFS horst delimits two relative structural lows, Inner GFS Low to the west and Outer GFS Low to the east.

The syn-rift supersequence records two phases of riftting (RP1 and RP2), where the latter is equivalent to syn-rift II (SRII) combined with syn-rift III sequence (SRIII) and the former with syn-rift I sequence (SRI). Both phases recorded an initial phase of sediment-starved half-grabens followed by widespread deposition that heals the rift-related structural relief. Lithostratigraphic content also differs significantly.

Rift phase (RP) 1 was deposited in isolated half-grabens typically defined by E-dipping NNE-SSW to N-S faults formed under a roughly WNW-ESE extension. Following the erosion and depocentre inversion of RP1 deposits, rift phase 2 (RP2) sediments deposited in longer grabens formed by segment linkage of roughly NE–SW fault zones dipping either east or westwards, thus defining an asymmetric horst. This horst comprises a fault system denominated Guriri Fault System (GFS), correlated southwards to a previously unnamed major structural high in the Central Campos Basin. A roughly NW-SE extension was active during early RP2 (SRII) and slightly rotated back to WNW-ESE towards the end of RP2 (SRIII).

The RTR supersequence is related to the onset of asymmetric thermal subsidence coeval with localized syn-depositional faulting. Thermal subsidence dominates throughout the post-rift supersequence. Towards the end of the post-rift stage, NW-SE tectonic shortening promoted selective reverse reactivation of rift faults, resulting in the upward propagation of fault-propagation folds. They were active at least until the onset of salt tectonics during the Albian. The inversion is interpreted as the result of intraplate stress propagation triggered by the onset of oceanic spreading in the Campos Basin during Early Albian. Also, erosion of fold hinge zones means that a significant degree of sub-aerial erosion preceded the onset of evaporite deposition.

Based on ages of dyke emplacement onshore, new age constraints for the end of the pre-salt deposition, and in order not to shorten the traditionally accepted span of the rift plus post-rift supersequences by half, we suggest that the onset of PRI occurred during the Berriasian.

The tectonic shortening and inversion during the post-rift have potential links with the localized syn-sedimentary hydrothermal venting during post-rift deposition and the widespread hydrothermal diagenesis recorded in the pre-salt succession. In addition, events of sub-aerial exposure point to potential epigenetic karstification on the top of RTR, PRI and PRII sequences. Hence, these pieces of information should be considered for future investigations and, mostly, for reservoir characterization and modelling of the Pre-Salt reservoir of the Campos Basin.

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**DATA AVAILABILITY STATEMENT**
The data used in this publication is of public domain. However, it belongs and are managed by the Brazil’s Agency of Hydrocarbons (ANP) and could be requested directly to the agency for academic purposes. Therefore, the authors are not authorized to forward the data used in this publication.

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