

Research paper

The South American retroarc foreland system: The development of the Bauru Basin in the back-bulge province

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ARTICLE INFO

Article history:

Received 16 October 2015

Received in revised form

21 February 2016

Accepted 23 February 2016

Available online 27 February 2016

Keywords:

Andean orogeny

Back-bulge

Late Cretaceous

Retroarc foreland system

Subsidence mechanisms

ABSTRACT

The aim of this research is to understand the tectonic setting of the Bauru Basin. This basin in central-eastern South America has been classified as intracratonic, but the basin-fill geometry, the involved subsidence mechanisms and the age of the deposits are poorly understood. In this work, the ranges of the fossil taxa are analyzed and ages are proposed for the lithostratigraphic units. Isopach maps were used to reconstruct the stratigraphic intervals of the basin fill. The stratigraphy of the Bauru Basin is compared with that of the adjacent basins, and the data are integrated with the available information on South American geodynamics. The fossil record indicates that sediment accumulated from the Cenomanian to early Paleocene, beginning after the Mochica Phase of the Andean orogeny. The basin-fill geometry demonstrates migration of the depocenter through time, which occurred simultaneously with migration of the Andean Basin and immediately after the orogenic events of the Peruvian Phase. We propose that the Bauru Basin is a component of a retroarc foreland system developed during the early stages of the Andean evolution and that it was developed in the back-bulge province of this system. The Andean Basin constitutes the foredeep depozone of this foreland system (including the Potosí, Oriente, Acre and Marañón basins). In addition, the Upper Cretaceous of the Parecis and Solimões basins were likely also developed in the back-bulge province. The thickness of the Bauru accumulation indicates that other mechanisms might have overlapped the flexural subsidence in this back-bulge province.

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1. Introduction

Classification of sedimentary basins situated in intraplate settings far from plate boundaries is not obvious, as in certain other tectonic settings. To describe these basins as “intracratonics” is an oversimplification, especially if the subsidence mechanisms are not completely clear. The Bauru Basin in central-eastern South America is a case in point (Fig. 1). This basin has been classified as intracratonic, but its origin and stratigraphy can be explained with reference to plate-margin processes, and therefore, it is not intracratonic in a tectonic sense.

The Bauru Basin covers an area of approximately 379,362 km² located almost exclusively in Brazil (Figs. 1 and 2), with selected outcrops in Northeastern Paraguay (Fúlfaro, 1996). This Cretaceous

sedimentary succession reflects changing nonmarine environments, such as eolian, lacustrine, fluvial and alluvial fans.

The aim of this research is to analyze the mechanisms responsible for creation and development of the Bauru Basin. The stratigraphic range of its paleobiota is examined in detail, accompanied by age determinations of the lithostratigraphic units. The current work improves the understanding of large-scale stratigraphic patterns, subsidence mechanisms, the role of tectonics in control of the overall geometry of the Bauru basin-fill, and the geodynamics of the South American Plate and also makes comparisons between the Bauru Basin and other Cretaceous basins in South America. Isopach maps were constructed using published well log information (SIAGAS, n.d.; Sallun Filho et al., 2009) and unpublished data provided by DAEE (Departamento de Águas e Energia Elétrica, São Paulo State Government), which were, both calibrated with cores. These isopach maps illustrate the basin-fill geometry and the migration of its depocenter through time.

A deeper understanding of the tectonic setting and sedimentation controls of the Bauru Basin supports the construction of

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Fig. 1. Tectonic elements of the South American Plate. Terranes of the western margin and cratons are from Ramos (2009) and Ramos et al. (2010). Sedimentary basins are from Ramos (1999) and Milani et al. (2007b). Brazilian cratons, provinces and belts are from Cordani and Teixeira (2007). Paranapanema craton is from Mantovani et al. (2005). Transbrasilian Lineament is adapted from Cordani and Teixeira (2007), Ramos et al. (2010) and Curto et al. (2014).

stratigraphic models that might improve characterization and exploitation of the Bauru aquifer, which is located in one of the most populous regions of Brazil. In addition, the models will support biogeographic and evolutionary studies of its paleobiota. The Bauru Basin analysis aggregates information to understand the South American geodynamics in the Late Cretaceous and the development of several contemporaneous basins, which might be useful for predicting the presence of key elements of petroleum systems in frontier basins and could provide a model of investigation for similar older basins in South America.

1.1. Geological background and tectonic setting

From the Late Triassic to Early Cretaceous, an extensional regime prevailed in the South American Plate related to the Gondwana breakup and opening of the South Atlantic Ocean (Mégard, 1984; Sempere et al., 1997; Zeffass et al., 2004, 2005; Ramos, 2009; Bockhout et al., 2012). After the breakup in the Pelotas and Santos basins, in the late Early Cretaceous (Chang et al., 1992; Zalán and Oliveira, 2005; Moreira et al., 2007; Stica et al., 2014), the South American Plate changed the absolute motion to the west increasing the convergence rate with the Farallon Plate (Ramos, 1999, 2009; Ramos and Alemán, 2000; Maloney et al., 2013); and the stress regime in South America changed from extensional to compressional (Chang and Kowsmann, 1996; Riccomini et al., 2005; Ramos, 2009, 2010; Folguera et al., 2011). Therefore, the western margin of South America was first subjected to compression and began uplift of the early Andean orogen (Mégard, 1984; Jaillard and Soler, 1996; Jaillard et al., 2000; Ramos, 2009, 2010; Folguera et al., 2011; Vásquez and Altenberger, 2005; Maloney et al., 2013; Pfiffner and Gonzalez, 2013; Fennel et al., 2015). Consequently, in the early Late Cretaceous, a series of foreland basins developed parallel to the Andean thrust front, with the Andean Basin in the northern and central regions of the Central Andes (Jaillard and Sempere, 1991; Rouchy et al., 1993; Martinez and Mamani, 1995; Sempere et al., 1997; Jaillard et al., 2000; Mpodozis et al., 2005; Fennel et al., 2015).

In the interior continent, Late Cretaceous successions were deposited over Paleozoic basins parallel to the Andean Basin. The Bauru Basin, which is the focus of this study, is superimposed on the northern Paraná Basin and primarily overlaps the volcanic rocks of the Serra Geral Formation (Fig. 1). In this area, is the crustal thickness averages 40 km (Assumpção et al., 2013; van der Meijde et al., 2013; Mariani et al., 2013) with high flexural rigidity (Pérez-Gussinté et al., 2007) and is largely of the Paleoproterozoic Paranapanema Craton (Mantovani et al., 2005; Ramos et al., 2010).

The mechanisms responsible for the Bauru Basin subsidence are controversial. It has been argued that reactivation of tectonic elements of the basement resulted in an extensional sag as a consequence of opening of the South Atlantic Ocean, westward migration of the South American Plate and its interaction with adjacent plates (Fernandes and Coimbra, 1992; Fúlfaro and Perinotto, 1996; Paula e Silva et al., 2009). Suguio et al. (1977) and Fúlfaro and Barcelos (1993) proposed grabens and horsts. Certain authors noted load and thermal subsidence related to cooling of the subjacent volcanic rocks (Zalán et al., 1990; Fernandes and Coimbra, 1996, 2000; Riccomini, 1997b; Milani, 2003; Milani et al., 1994, 2007; Milani and De Wit, 2008; Fernandes and Ribeiro, 2015). From the point of view of Chang and Kowsmann (1996), the region of the Bauru deposits were downwarped by the action of intraplate stresses related to the inversion of the South American stress regime. Mariani et al. (2013) suggested that crustal overloading and cooling of magmatic underplating generated the accommodation for the Bauru deposits.

The subjacent Paraná Basin was developed from the Ordovician to Early Cretaceous (Milani et al., 1994, 2003), and its sedimentation

has been associated with cycles of terrane accretions in the pre-Andean margin of the Gondwana (Zalán et al., 1990; Milani et al., 1994, 2003; Milani et al., 2007; Catuneanu et al., 1998). The Paraná Basin resulted from flexural foreland and intracratonic subsidence mechanisms (Milani and De Wit, 2008) and can be understood in terms of different basins superposed in time and space (Zalán et al., 1990) or a sedimentary pack subdivided into sequences related to different subsidence mechanisms (Milani et al., 1994, 2003; Milani et al., 2007). In this context, the Bauru deposits have been classified as a sequence of the Paraná Basin (Soares et al., 1980; Zalán et al., 1990; Chang and Kowsmann, 1996; Milani, 2003; Milani et al., 1994, 2007; Milani and De Wit, 2008; Paula e Silva et al., 2009) or assumed as a distinct tectonic unit (Fernandes and Coimbra, 1992; Fúlfaro and Perinotto, 1996; Fernandes and Coimbra, 1996, 2000; Riccomini, 1997b).

There is a lack of consensus on the beginning of the sedimentation in the Bauru Basin and the age of each lithostratigraphic unit, which have been precluded correlations with other basins and geodynamic events. Soares et al. (1980) attributed Aptian age to the Caiuá and Santo Anastácio formations, Cenomanian–Santonian to the Adamantina Formation, and Santonian–Maastrichtian to the Marília Formation. The palynological studies of Lima et al. (1986) established certain mudstones strata as Coniacian, and posteriorly, these strata were assumed as the Araçatuba Formation. Gobbo-Rodrigues et al. (1999a, 1999b) stated that the Adamantina and Araçatuba formations are Campanian–Maastrichtian and the Marília Formation is Maastrichtian, considering the record of ostracods and charophytes. Santucci and Bertini (2001) indicated Campanian–Maastrichtian for the Adamantina and Marília formations, in view of the Sauropodomorpha record. Dias-Brito et al. (2001) noted Turonian–Santonian for the Adamantina Formation (not discriminated from the Araçatuba Formation), and Maastrichtian age for the Marília Formation by focusing on ostracods and charophytes. Furthermore, Dias-Brito et al. (2001) estimated an Aptian–Cenomanian age for the Caiuá and Santo Anastácio formations, and Coniacian–Santonian for the Uberaba Formation, considering their stratigraphic positions.

Certain recent fossil discoveries, not only in Bauru but also in other basins, have brought new elements to the discussion of the age of the Bauru Basin. In addition, the detailed geological mapping realized in the 1990s and the latest lithostratigraphical studies of the Bauru Basin (e.g. Fernandes and Coimbra, 1996, 2000; Paula e Silva et al., 1994, 2009) resulted in redefinition of a subset of the Bauru Basin fossil localities. Herein, the detailed exam of the biostratigraphic information reveals that these new discoveries extend the temporal range of selected taxa, consequently clarifying the age of the sedimentation, and indicate ages of units previously thought to be without fossils.

The definition of the ages of the events of creation of accommodation space in the Bauru Basin, allowed the establishment of correlations with other basins and coeval geodynamic events. Then, we propose that the Bauru basin-fill corresponds to a first order sequence, that is, the product of sedimentation within a particular tectonic setting. This tectonic setting conditioned the development of a new basin in an area that presents older sedimentary deposits of distinct nature. Therefore, this unit is herein viewed as a basin.

2. Bauru Basin

2.1. Lithostratigraphy

Although a large discrepancy exists between the current lithostratigraphic proposals for the Bauru Basin (Soares et al., 1980; Barcelos et al., 1983; Fúlfaro and Perinotto, 1996; Fernandes and Coimbra, 2000; Paula e Silva et al., 2005, 2009; Fernandes and

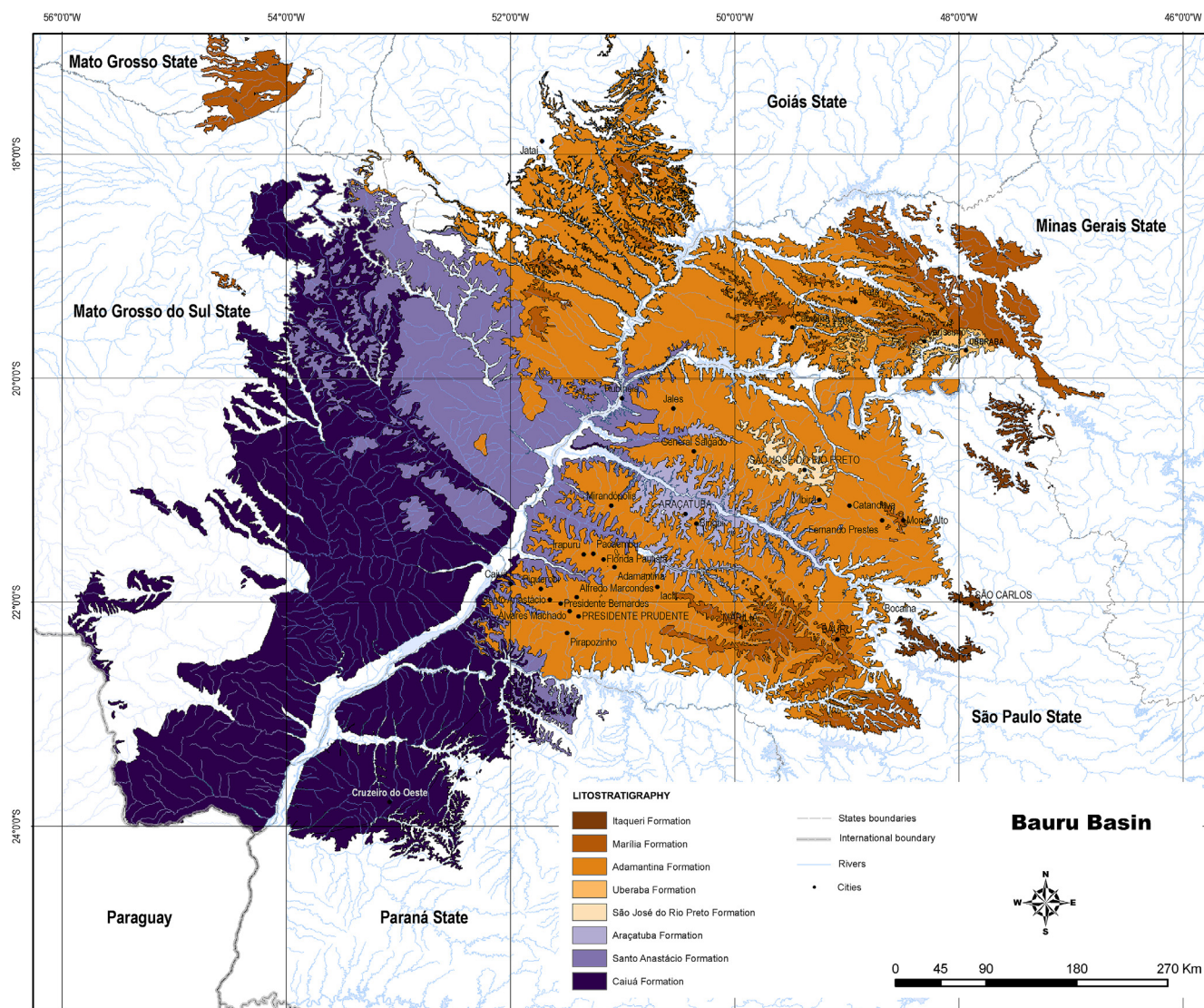


Fig. 2. Geological map of the Bauru Basin in Brazil. Modified from [Fernandes and Coimbra \(2000\)](#) and [Perrotta et al. \(2005\)](#). Selected cities indicate fossiliferous localities discriminated in [Supplementary Material](#).

[Ribeiro, 2015](#)), this succession can be characterized by ten units, namely, the Caiuá, Pirapozinho, Araçatuba, Santo Anastácio, Birigui, São José do Rio Preto, Uberaba, Adamantina, Marília and Itaqueri formations ([Fig. 2](#)).

The Caiuá and Pirapozinho formations represent the base of the basin. The Pirapozinho Formation partially overlays the Caiuá Formation, and both are superimposed on the Serra Geral Formation. The Caiuá Formation consists of well-sorted and well-rounded sandstones, with large-scale cross-bedded sets. The cross-beds commonly show coarse layers with reverse and normal grading alternating with finer beds. Additionally present are fining-upward cycles with intraformational conglomerates, fine to medium sandstones with cross-stratification or ripples, and mudstone layers. These deposits have been interpreted as eolian ([Soares et al., 1980](#)) and fluvial ([Paula e Silva et al., 2005, 2009](#)). According to [Fernandes and Coimbra \(2000\)](#), this unit can be subdivided into the Goio Erê and Rio Paraná formations, which is not adopted in this research because was not found distinction between these deposits. As reported by [Fúlfaro \(1996\)](#) and [Fariña \(2009\)](#), the Acaray Formation is the equivalent of the Caiuá Formation in Paraguay.

The lower portion of the Pirapozinho Formation is composed of

massive and thinly laminated mudstones interbedded with thin notably fine sandstone strata with ripple cross-lamination. Multiple coarsening-upward cycles of mudstones to rippled fine sandstones predominate in the upper portion, but fine to medium sandstones with planar-to trough-cross stratification also occur as well as conglomerates with intraclasts of mudstones. Bioturbation and root marks are frequent. These deposits result from lacustrine and fluvial depositional environments ([Paula e Silva et al., 2005, 2009](#)).

The Santo Anastácio, Araçatuba and Birigui formations occur in an intermediate position. The Santo Anastácio Formation overlies the units below, including the Serra Geral Formation. The Araçatuba Formation partially covers the Santo Anastácio Formation and is in contact with the basalts of the Serra Geral Formation, the basement of the basin. The Birigui Formation is interdigitated with the Araçatuba, and in certain well logs, the Birigui Formation appears above the Santo Anastácio Formation.

The Santo Anastácio Formation consists of amalgamated fine to medium sandstones beds and can be massive or display planar cross-stratification. Certain beds have primary stratifications that are obscured or destroyed, with a mottled appearance. Carbonate nodules and root marks are quite common. From the point of view

of Fernandes and Coimbra (2000), these deposits correspond to sand sheets in marginal deserts plains. Soares et al. (1980) and Paula e Silva et al. (2009) noted a braided fluvial depositional environment. In addition, Fúlfaro et al. (1999) emphasized the occurrence of paleosols.

The Araçatuba Formation is primarily composed of massive and laminated mudstones, with notably small carbonates nodules. Secondly, very fine to fine sandstones occur with tabular stratification and ripple cross-lamination, wave ripples, heterolithic facies (flaser, lenticular and wavy bedding), mudcracks and eventually micro-hummocky lenses. Ichnofossils and small root marks are frequent. The Araçatuba Formation has been interpreted as lacustrine deposits (Fernandes and Coimbra, 2000; Batezelli et al., 2003; Paula e Silva et al., 2009; Fernandes and Ribeiro, 2015).

The Birigui Formation is composed of fine to medium sandstones with trough and planar-cross stratifications or ripple cross-lamination and heterolithic intervals with thickening-upward patterns. Ichnofossils, root marks and mudcracks are also present. This formation is interpreted as deposits of a braided fluvial system (Paula e Silva et al., 2009).

Finally, the upper portion of the Bauru Basin corresponds to the Adamantina, São José do Rio Preto, Uberaba, Marília and Itaqueri formations. The Adamantina Formation overlies the Santo Anastácio and Araçatuba formations, extends over the basalts of the Serra Geral Formation on the North and East of the basin, and laterally contacts the São José do Rio Preto and Uberaba formations. The São José do Rio Preto is superimposed on the Santo Anastácio Formation, whereas the Uberaba Formation occurs in a restricted area covering the Serra Geral Formation (Fig. 2). The Marília Formation overlies the Adamantina and Uberaba Formations, and the Itaqueri Formation on the eastern area of the basin covers the Serra Geral and Botucatu formations and is partially in contact with the Marília Formation.

The Adamantina Formation is characterized as fine sandstones with ripple cross-lamination or planar-to trough-cross stratification, eventually with intraclasts of mudstones at the base of the troughs, interbedded with heterolithic facies, mudstones with mudcracks, root marks and ichnofossils, and matrix-supported intraformational conglomerates. According to Soares et al. (1980) and Paula e Silva et al. (2009), these deposits resulted from a meandering fluvial depositional environment. From the point of view of Fernandes and Coimbra (2000), the Adamantina Formation can be partitioned into the Vale do Rio do Peixe, Presidente Prudente and São José do Rio Preto formations, the former resulting from eolian deposition and the others from fluvial environments. The São José do Rio Preto Formation has distinctive characteristics, making it easy to differentiate it from the Adamantina Formation. However, the Presidente Prudente and Vale do Rio do Peixe formations are not easily distinguished, and these designations are not adopted in this research.

The São José do Rio Preto Formation consists of very fine to medium sandstones with planar-to trough-cross stratification as well as ripple cross-lamination and cross-bedded intraformational conglomerates, which are massive or show grading of clasts and/or matrix. The conglomerates are rather frequent, varying from clast-supported to matrix-supported, eventually with angular clasts. Suguio (1981) and Fernandes and Coimbra (2000) agree that this unit is the result of a braided fluvial depositional system.

The Uberaba Formation is similar to the São José do Rio Preto Formation but differs in compositional immaturity and color. It has been argued that this unit is rich in fragments of volcanic rocks from the Paranaíba High, and it is associated with a braided fluvial depositional environment (Hasui, 1968; Gravina et al., 2002; Fernandes and Coimbra, 2000).

The Marília Formation consists of fining-upward cycles,

including matrix-supported conglomerates with intra and extra-formational clasts, fine to very coarse sandstones (massive, cross-stratified or ripple cross-laminated) and rare mudstones. Calcrete structures are quite common (laminar, prismatic or massive horizons), with abundant carbonate nodules (vertically elongated, branched or with irregular shape), horizontal cracks and rhizoliths. The Itaqueri Formation is quite similar to the Marília Formation but differs in a less expressive amount of carbonate cement and occurrence of layers with silica cementation. The Marília and Itaqueri formations are interpreted as alluvial fan deposits (Mezzalana, 1974; Soares et al., 1980; Riccomini, 1997a; Ladeira and Santos, 2005).

2.2. Analysis of the stratigraphic ranges of selected fossils

The plentiful biota of the Bauru Basin, despite the absence of any corroborative radiometric dating, allowed previous works to agree on a Cretaceous age for almost all of its deposits, except for the Itaqueri Formation that is usually assumed to have reached the early Paleocene (Riccomini, 1997a; Ladeira and Santos, 2005; Perrotta et al., 2005). This fossil assemblage is composed of conchostacans, ostracods, gastropods, bivalves, charophytes, pelynomorphs, frogs, sauropods, birds, theropods, mammals, lizards, fishes, crocodylomorphs, turtles and pterosaurs. The stratigraphic ranges of the main identified fossil taxa present in Bauru Basin are summarized in Fig. 3, and their provenance and temporal range dataset are presented in the Supplementary Material.

With respect to vertebrates, although the species are endemic, almost all of the groups identified in the Bauru Basin indicate Cretaceous ages. For example, considering the oldest record of frogs and lizards in South America (Baéz et al., 2012; Estes and Price, 1973; Candeiro et al., 2009; Nava and Martinelli, 2011; Simões et al., 2015), the presence of these groups in the Marília and Adamantina formations indicates an age younger than Aptian. In addition, in view of the Campanian–Maastrichtian occurrences in South America (Marshall et al., 1983; Bonaparte, 1990; Gayet et al., 2001), a mammal specimen from the Adamantina Formation tentatively identified as placental by Bertini et al. (1993), could indicate a late Late Cretaceous age.

The record of fish has been revised as well. Alves et al. (2013) assigned the genus *Asiatoceratodus* to the specimen that was previously identified as *Neoceratodus* (Bertini et al., 1993), found in the Adamantina Formation. However, the authors noted that this specimen was found in a reworked clast and might not indicate the age for the Adamantina Formation. Most importantly, the presence of Siluriformes indicate that certain strata of the Adamantina and Marília formations are younger than late Santonian.

The age range of the Testudines taxa is estimated as Late Cretaceous (Gaffney et al., 2011; Romano et al., 2013). Menegazzo et al. (2015) noted the closer relationship among the Santo Anastácio form (from the Santo Anastácio Formation), "*Podocnemis*" *brasiliensis* (from the São José do Rio Preto and Araçatuba formations) and *Portezueloemys patagonica* (from the Portezuelo Formation, upper Turonian/lower Coniacian of Argentina). Moreover, *P. mezzalirai* and *P. caiera* from the Marília Formation are sister taxa of *Lapparentemys vilavilensis* from the Santa Lucía Formation, dated to the Paleocene of Bolivia (Gaffney et al., 2011; Menegazzo et al., 2015). Hence, the record of turtles indicates that the Santo Anastácio, Araçatuba and São José do Rio Preto formations are relatively older than the Marília Formation.

The Theropoda assemblage is indicative of Late Cretaceous age for the Adamantina, Marília, Uberaba and São José do Rio Preto formations. In Argentina, Megaraptora is recorded from Cenomanian to Santonian (Novas et al., 2008; Benson et al., 2010; Méndez et al., 2012; Martinelli et al., 2013), according to the

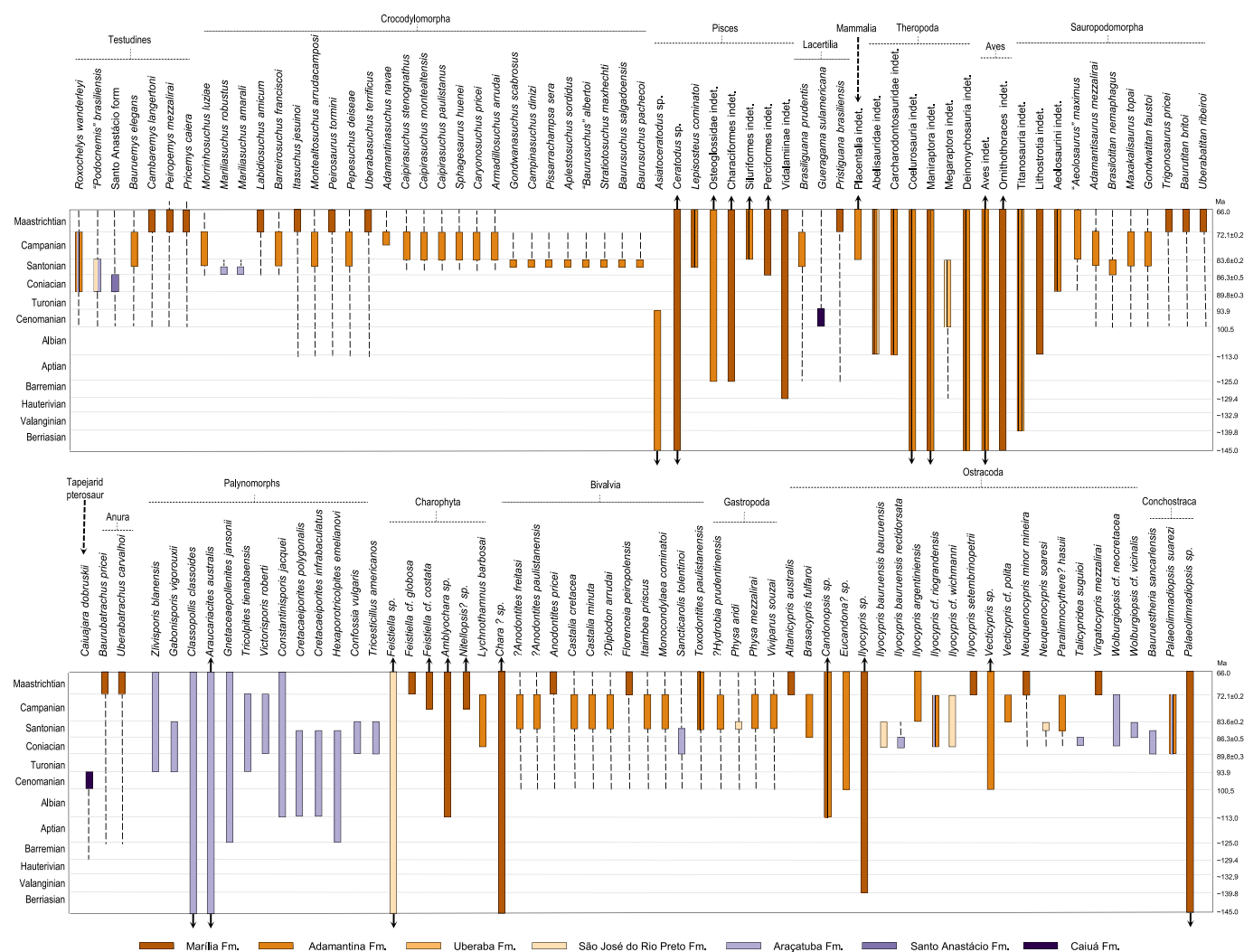


Fig. 3. Stratigraphic range chart of selected fossil taxa of the Bauru Basin. Dashed lines express possible fossil range of some endemic taxa. Colors indicate the occurrences in the lithostratigraphic units. See [Supplementary Material](#) for details and references. Time scale from [Cohen et al. \(2013\)](#).

specimens found in the Mata Amarilla Formation, which were previously dated as Maastrichtian and redefined as Cenomanian ([Varela et al., 2012; Novas et al., 2013](#)). Consequently, the Megaraptora presence in the São José do Rio Preto and Uberaba formations ([Méndez et al., 2012; Martinelli et al., 2013](#)) might be indicative of an age older than the Santonian–Campanian boundary.

The new taxonomic assignments of the sauropods from the Adamantina, Marília and Uberaba formations also indicate the Late Cretaceous. The genus *Aeolosaurus* has been used to attribute Campanian–Maastrichtian to the Adamantina and Marília formations ([Bertini et al., 1999; Santucci and Bertini, 2001](#)), based on its Argentinian record. However, from the point of view of [Martinelli et al. \(2011\)](#) and [Filippi et al. \(2013\)](#), the evidence of this genus in the Bauru Basin is inconclusive. Certain specimens anteriorly identified as *Aeolosaurus* ([Santucci and Bertini, 2001; Franco-Rosas et al., 2004; Marinho and Candeiro, 2005; Lopes and Buchmann, 2008; Santucci and Arruda Campos, 2011](#)) have been assigned to Titanosauria ([Martinelli et al., 2011](#)) and Aeolosaurini indet ([Martinelli et al., 2011; Filippi et al., 2013](#)). Considering that Aeolosaurini is recorded in Argentina from the lower Coniacian to Maastrichtian ([Filippi et al., 2013](#)), it is possible to assume that the Adamantina, Marília and Uberaba formations were deposited

within this interval. In addition, [Machado et al. \(2013a, b\)](#) noted that *Brasilotitan nemaphagus* from the Adamantina Formation appears to be closely related to *Antarctosaurus wichmannianus* (Upper Cretaceous, Argentina) and *Bonitasaura salgadoi* (Santonian, Argentina), which might also indicate a Santonian age for at least a portion of the Adamantina Formation.

The rich record of Crocodylomorpha in the Araçatuba, São José do Rio Preto, Adamantina and Marília formations also indicates the Late Cretaceous. In particular, *Mariliasuchus* is suggestive of Santonian in the upper portion of the Araçatuba Formation (or lower portion of the Adamantina Formation) because several studies ([Andrade and Bertini, 2008; Turner and Sertich, 2010; Godoy et al., 2014; Pol et al., 2014](#)) noted its closer relationship with *Notosuchus* from the Santonian of Argentina. The specimens of *M. amarali* and *Mariliasuchus robustus* are from an outcrop originally identified as the Araçatuba Formation ([Carvalho and Bertini, 1999](#)), and posteriorly assigned to the Adamantina Formation ([Carvalho and Bertini, 2000; Zaher et al., 2006; Nobre et al., 2007](#)). However, considering the descriptions of [Carvalho and Bertini \(2000\)](#) and [Zaher et al. \(2006\)](#), this outcrop is most likely to be the Araçatuba Formation, probably near the contact with the overlaying Adamantina Formation. Phylogenetic analyses have shown that *Pissarachampsia* is a sister taxon of *Wargosuchus*, from the Santonian of Argentina

(Montefeltro et al., 2011; Godoy et al., 2014), which corroborates a Santonian age to at least a portion of the Adamantina Formation. In contrast, *Adamantinasuchus* might suggest a younger age for certain strata of the Adamantina Formation because it has been considered a sister taxon of *Yacarerani* (Novas et al., 2009; Iori and Carvalho, 2011; Pol et al., 2014), from the Maastrichtian Cajones Formation in Bolivia. Nevertheless, doubts exist related to the association of *Adamantinasuchus* with the Adamantina Formation because lithological descriptions were not published, and the indicated provenance is in the outcrop area of the Marília Formation. Finally, Montefeltro et al. (2009) indicated the affinity of isolated crocodyliform teeth from the São José do Rio Preto Formation with middle Cretaceous taxa. This classification might indicate a relatively older age for the São José do Rio Preto Formation than that of the Adamantina and Marília formations.

The recent description of the Tapejaridae pterosaur *Caiuajara dobruskii* (Manzig et al., 2014) and the acrodontan lizard *Gueragama sulamericana* (Simões et al., 2015) brought clarification for the age of the poorly fossiliferous Caiuá Formation. This unit previously been dated as Late Cretaceous based on footprints tentatively associated to small theropods, dinosaurs and mammals (Leonardi 1977, 1989; Fernandes et al., 2013; Fernandes et al., 2008; Fernandes and Ribeiro, 2015). The record of *G. sulamericana* is also indicative of the Late Cretaceous (Simões et al., 2015), but according to Manzig et al. (2014), tapejarids are usually present in deposits ranging from Barremian to Cenomanian, as in the northeastern of Brazil (Kellner and Campos, 2007), China, Morocco (Wang and Zhou, 2003) and Spain. Therefore, Cenomanian is the youngest age that can be assigned to the Caiuá Formation.

For invertebrate fossils, the endemism and paucity of information on bivalves, gastropods and conchostracans inhibit their use in biostratigraphy. However, the records of *Physa aridi* in the São José do Rio Preto Formation and *Bauruetheria sancarlensis* in the Araçatuba Formation offer additional information. According to Mezzalana (1974), *P. aridi* is quite similar to *P. wichmanni*, which has been recognized in the Aptian–Maastrichtian of Argentina, Lohan Cura, Allen, Los Alamitos and Augostura Colorado formations (Martinelli et al., 2007). Rohn et al. (2005) demonstrated the similarity between *B. sancarlensis* and selected species of the late Cenomanian–Santonian of Asia.

However, the noteworthy record of ostracods has been used to assign relative ages to the uppermost units of the Bauru Basin. These records also indicate a Late Cretaceous age, considering that Dias-Brito et al. (2001) express serious reservations about the identification of certain Early Cretaceous ostracods in the Adamantina and Marília formations (*Paracyprina*? sp. 1, *Paracyprina*? sp. 2, *Reconcovona* ?ultima, “*Salvadoriella*” sp., “*Lycoperocypris*” cf. *angulate*, *Mantelliana* sp., and “*Hourcya*” sp.) and no further works confirmed the presence of these taxa. Thus, no conclusive evidence exists of Early Cretaceous ostracods in the Bauru Basin.

Recent studies extend the geographic distribution and temporal range of certain ostracod species in South America (Musacchio and Vallati, 2007; Carignano and Varela, 2011), and selected outcrops where these ostracods were collected by Dias-Brito et al. (2001) were redefined. In particular, certain outcrops in the city of Ibirá (São Paulo State) that were previously identified as the Adamantina Formation are assumed as related to the São José do Rio Preto Formation (Fig. 2), following the work of Fernandes and Coimbra (2000). Thus, the record of *Ilyocypris wichmanni*, *Ilyocypris bauruensis rectidorsata* and *Neuquenocypris soaresi* is associated with the São José do Rio Preto Formation. Moreover, certain localities with lacustrine facies association previously recognized as the Adamantina Formation in the region of Presidente Prudente and Álvares Machado cities (São Paulo State) likely represent the Araçatuba Formation, considering the studies of Paula e Silva et al.

(1994, 2006) and Gobbo-Rodrigues et al. (1999b). Consequently, the record of *I. bauruensis rectidorsata*, *Talicypridea suguioi*, *Wolburgiopsis* cf. *neocretacea* and *W. vicinalis* are associated with the Araçatuba Formation.

The Ostracoda taxa *Altanicypris australis*, *Brasacypris fulfaroi*, *I. bauruensis*, *I. bauruensis rectidorsata*, *Ilyocypris setembrinopetriti*, *Neuquenocypris minor mineira*, *N. soaresi*, *Paralimnocythere*? *hasuii*, *T. suguioi*, and *Virgatocypris mezzalirai* are endemic in Bauru Basin (Dias-Brito et al., 2001). However, Musacchio and Vallati (2007) recognize *I. bauruensis* and *Talicyprideinae* sp. in the upper Coniacian of Argentina (Plottier Formation), considering these specimens respectively similar to *I. bauruensis rectidorsata* and *T. suguioi* from the Araçatuba Formation. The subspecies *N. minor mineira* is found only in Marília Formation but is highly similar to *N. minor* (Dias-Brito et al., 2001), present in the Allen and Plottier formations of the upper Coniacian and Maastrichtian in Argentina (Musacchio and Simeoni, 1991; Musacchio and Vallati, 2007). In addition, Dias-Brito et al. (2001) consider *A. australis* from the Marília Formation to be closer to the *Antarctosaurus szzechuriae* that is present in the Campanian and Maastrichtian of Asia.

Other ostracod species from the Bauru Basin have a more widespread distribution in South America. For example, *Ilyocypris argentinensis* from the Adamantina Formation (Gobbo-Rodrigues et al., 1999a) is also found in the Loncoche Formation of the Campanian–Maastrichtian in Argentina (Uliana and Musacchio, 1978; Parras and Griffin, 2013). *Ilyocypris riograndensis* is found in several localities in Adamantina (Dias-Brito et al., 2001) and Araçatuba formations (Gobbo-Rodrigues et al., 1999b) and is also described in the Plottier, Bajo de la Carpa and Anacleto formations of the upper Coniacian–Campanian in Argentina (Musacchio and Vallati, 2007; Carignano and Varela, 2011). *I. wichmanni* from the São José do Rio Preto Formation in the Ibirá city region (Dias-Brito et al., 2001) is found in the Plottier (Musacchio and Vallati, 2007) and Anacleto formations (Musacchio, 2000). *Wolburgiopsis* cf. *neocretacea* present in Araçatuba Formation (Gobbo-Rodrigues et al., 1999b; Dias-Brito et al., 2001) are also recorded in the Plottier, Bajo de la Carpa and Anacleto formations of the upper Coniacian–Campanian in Argentina (Bertels, 1972; Uliana and Musacchio, 1978; Carignano and Varela, 2011). Similarly, *W. vicinalis* from the Santonian Bajo de la Carpa Formation (Uliana and Musacchio, 1978) is identified in the Araçatuba Formation (Gobbo-Rodrigues et al., 1999b; Dias-Brito et al., 2001). According to Colin et al. (2000), the genus *Metacypris* previously identified in the Adamantina Formation (Dias-Brito et al., 2001) was redefined as *Vecticypris* and is recorded from Cenomanian to Paleogene (Colin et al., 2000; Altınışık et al., 2004). In addition, Carignano and Varela (2011) stated that the specimens of *Vecticypris* sp. from the Allen Formation of the upper Campanian and Maastrichtian in Argentina are highly similar to *Vecticypris* cf. *polita* described in the Adamantina Formation (Dias-Brito et al., 2001).

With respect to the record of charophytes, several specimens originally described as *Chara barbosai* (Petri, 1955) of the Adamantina Formation (Dias-Brito et al., 2001) were included in the new combination *Lychnothamnus barbosai* (Musacchio, 2006, 2010), which is also identified in the upper Coniacian of Argentina. From the point of view of Dias-Brito et al. (2001), selected gyrogonites from the Marília Formation might be related to *Globichara* (*Pseudoharrisichara*), but the subspecies *Pseudoharrisichara* (Musacchio, 1973) were included in *Lychnothamnus* (Musacchio, 2010), and thus, the classification of these specimens remains inconclusive. More important are the charophytes taxa indicative of Maastrichtian age for the Marília Formation. According to Dias-Brito et al. (2001), the gyrogonites of *Feistella globosa* are comparable to those from Maastrichtian of Spain (Grambast and Gutteriez, 1977), and the gyrogonites of *F. costata* are similar to

those described in the Maastrichtian and Paleocene of Peru and Ecuador (Jaillard et al., 1993, 1994). The gyrogonites of *Amblyochara* sp. are quite similar to those described by Musacchio (1973) in the Yacoraite Formation of the Maastrichtian in Argentina (Dias-Brito et al., 2001). This genus is also recorded in the Campanian and Maastrichtian of Peru and Bolivia (Jaillard et al., 1993, 1994). The genus *Nitellopsis*, tentatively identified in the Marília Formation (Dias-Brito et al., 2001), is documented from the upper Campanian to Paleocene of Peru and the Acre Basin in Brazil (Jaillard et al., 1993, 1994; Musacchio, 2000).

The record of palynomorphs is restricted to the Araçatuba Formation from one locality in São Carlos City, São Paulo State (Fig. 2). According to Lima et al. (1986), the palynological assemblage of these strata indicates a Coniacian age. For the authors, this locality corresponds to the “Itaqueri Lithofacies”, the upper portion of the Bauru Basin. Conversely, Fernandes and Coimbra (1996) and Gobbo-Rodrigues et al. (1999b) consider that these strata are located in a basal position. Castro et al. (2002) described this outcrop as the contact of fluvial and lacustrine facies, denominated these strata as the São Carlos Formation and concluded a Coniacian–Santonian age. The description of these outcrops corresponds closely to observations of the contact between the Adamantina (fluvial) and Araçatuba (lacustrine) formations in other regions, and the water wells in the region of São Carlos confirm the occurrence of Bauru deposits in this area (Sallun Filho et al., 2009). Consequently, this locality is understood as the Araçatuba Formation. Thus, the record of the angiosperm pollen grains of *Victorisporis roberti*, *Confossia vulgaris* and *Tricestocillus americanus* (Lima et al., 1986; Vallati, 2010) limit the base of the temporal range of the Araçatuba Formation to Coniacian. Additionally, *Creataceiporites polygonalis*, *C. infrabaculatus* and *Hexaporotricolpites emelianovi* limit the top of this range to the early Santonian.

2.3. Age of the lithostratigraphic units

Based on the range of the fossils discussed previously, the age of the Bauru Basin deposits can be estimated from the Cenomanian to early Paleocene (Fig. 3). However, the evidence in each formation is not strictly accurate. Fossils are not yet recorded in certain units, and in others, the record is still poor. In addition, the paleontological data are not uniform throughout the entire basin, and the level of knowledge differs depending on the location and density of studies, which are generally concentrated on the east side of the basin.

The association of *C. dobruskii*, *G. sulamericana* and theropod footprints is suggestive of Cenomanian age for the Caiuá Formation. The Pirapozinho Formation can be considered partially chronocorrelated to the Caiuá Formation; these formations are in lateral contact, and the Pirapozinho Formation partially overlies the Caiuá Formation. Therefore, the age of the Pirapozinho Formation is estimated as Cenomanian–early Turonian?, although fossils have not yet been recorded in this unit.

The palynological assemblage assigns a Coniacian–early Santonian age to the Araçatuba Formation, which is consistent with the presence of the ostracods *I. bauruensis rectidorsata*, *I. riograndensis*, *T. suguioi*, *Wolburgiopsis neocretacea*, *W. vicinalis*, the conchostracan *B. sancarlensis* and the crocodyliform *Mariliasuchus*. Taking into account the stratigraphic relationship of the Santo Anastácio Formation with the adjacent units, it is assumed to be younger than the Caiuá and Pirapozinho formations, partially chronocorrelated with the Araçatuba Formation, and older than the other formations, and with the additional consideration of the record of the fossil turtle in the Santo Anastácio form, the age of the Santo Anastácio Formation is estimated as Coniacian. Additionally, although no record of fossils exists in the Birigui Formation, its age can be

estimated as late Santonian because it is interdigitated with the Araçatuba Formation and occurs in a position upper to the Santo Anastácio Formation.

The age of the Adamantina Formation is late Santonian–Campanian, taking into account the record of *I. argentinensis*, *I. riograndensis*, *V. polita*, *P. hasui*, *Aeolosaurini*, *B. nemaphagus*, *Placentalia*, *Siluriformes*, *P. sera*, *Campinasuchus dinizi* and *Adamantinasuchus navae*.

The age of the São José do Rio Preto Formation is estimated as Santonian, considering its stratigraphic relationship, i.e., partially correlated with the Araçatuba and Adamantina formations and overlying the Santo Anastácio Formation as well as the association of “*P.*” *brasiliensis*, *Megaraptora*, *I. bauruensis*, *I. wichmanni* and isolated teeth of crocodyliformes.

Certain studies have noted that magmatic rocks of the Paranaíba High constituted a source of sediments for the Uberaba Formation (Hasui, 1968; Fúlfaro and Barcelos, 1991; Gravina et al., 2002; Batezelli et al., 2005). Considering that different events of alkaline magmatism occurred in this area between 100 and 79 (Hasui and Haralyi, 1991; Brod et al., 2005; Riccomini et al., 2005) and the record of *Megaraptora* in the Uberaba Formation, the deposition of this unit is assumed older than the Santonian–Campanian boundary, and its age can be considered Santonian.

Taking into account the occurrence of *P. mezzalirai*, *P. caiera*, *Siluriformes*, *Perciformes*, *F. globosa*, *F. costata*, *Nitellopsis*, *A. australis* and *N. minor mineira*, Maastrichtian age is indicated for the Marília Formation. Although the range of certain of these taxa crosses the K–T boundary, the presence of Theropoda and Sauripodomorpha limits the deposition of the Marília Formation to the Cretaceous.

The Itaqueri Formation has been correlated with the base of the Marília Formation or considered younger (Fúlfaro and Perinotto, 1996; Riccomini, 1997a; Ladeira and Santos, 2005), although there is no fossil record. Riccomini (1997a) noted the Paleocene–Eocene age of Itaqueri Formation, considering that its silicified sandstones resulted from hydrothermalism related to alkaline magmatism in the Jaboticabal region, as proposed by Coimbra et al. (1981). However, the age of this magmatic event is controversial (Coimbra et al., 1981; Gomes and Valarelli, 1970; Coutinho et al., 1982) and the K/Ar dating might be masked by deuterio alteration (Valarelli et al., 1985). In addition, Ladeira and Santos (2005) consider this silicification a result of pedogenic processes but also indicate Cenozoic age for the Itaqueri Formation. In agreement with Perrotta et al. (2005), in this study, the Itaqueri Formation is considered Maastrichtian? –early Paleocene until fossil discoveries can clarify its age.

2.4. Sedimentation

As demonstrated by the paleontological data, the sedimentation of the Bauru Basin began in the Cenomanian with the deposition of the Caiuá and Pirapozinho formations, superimposed on the basalts of the Serra Geral Formation (Fig. 4A). The isopach map of the preserved entire basin-fill demonstrates that its geometry is lightly asymmetric (Fig. 4B) with two major depocenters. The isopach pattern is concentric, lacking evidence of synsedimentary faulting. In addition, the isopach map of the Cenomanian–Turonian sequence makes it evident that its deposition was not uniform throughout the entire basin and was limited to the southern area (Fig. 4C), where the maximum preserved thickness reaches 271 m (well 2-AN-0001-BR, SIAGAS n.d.).

After an unconformity that separates the Santo Anastácio Formation from the Pirapozinho and Caiuá formations, previously recognized by Paula and Silva et al. (2009), a new depositional cycle began in the Coniacian in the Bauru Basin, initiated with the

Araçatuba and Santo Anastácio formations and followed by the Birigui, São José do Rio Preto, Uberaba and Adamantina formations (Fig. 5). This Coniacian–Campanian depositional cycle overlaps only partially with the Cenomanian–Turonian succession. Indeed, the isopach map of this stratigraphic interval demonstrates the significant migration of its depocenter, from northern Paraná State to north-northwestern São Paulo State (Fig. 4D), representing a cratonward shift of approximately 250 km. The maximum preserved thickness of the Coniacian–Campanian sequence is 226 m in the São José do Rio Preto region.

Immediately after the Late Campanian tectonic phase, accommodation was created in the Bauru Basin, enabling the sedimentation of alluvial fan deposits (Marília and Itaqueri formations) (Fig. 2). These deposits partially recover the Santonian–Campanian Adamantina Formation onlapping the north and east borders of the basin. Apparently, this new cycle expresses a shift of the basin's depocenter, but insufficient information exists to reconstruct the basin-fill geometry of this stratigraphic interval. The sedimentation in the Bauru Basin ended with the Maastrichtian–lower Paleocene sequence.

The Bauru Basin is primarily bounded by Paleozoic–Mesozoic

sedimentary rocks of the Paraná Basin and the basalts of the Serra Geral Formation (Fig. 1). Along the northwestern, northeastern and southwestern it is also bounded by Cretaceous alkaline magmatic rocks and Paleoproterozoic rocks of the basement, exposed on the Asunción Arch, Rondonópolis Anticline and Paranaíba High. This areas supplied sedimentary, magmatic and metamorphic detritus to the basin. For example, the Asunción Arch was established in the Early Paleozoic (Almeida, 1983; Gomes et al., 2013), where deposition and erosion alternated during the Chaco–Paraná Basin evolution (Zalán et al., 1990; Milani et al., 1994, 2007). From the Valanginian to Aptian, alkaline magmatism occurred in this region, and the Asunción Arch was uplifted during the Cretaceous (Comin-Chiaramonti et al., 1999; Comin-Chiaramonti et al., 2014; Gomes et al., 2013); therefore, it might have been a source area of sediments for the Bauru Basin. Indeed, Coimbra (1976) recognizes a pre-existent sedimentary and volcanic (basalt) provenance for the lower interval of the basin, corresponding to the Cenomanian–Turonian sequence.

In the following sequences, from Coniacian to Paleocene, occurred a relatively more cratonal derivation of the sediments. The Paranaíba and Rondonópolis structures, that separate the Bauru

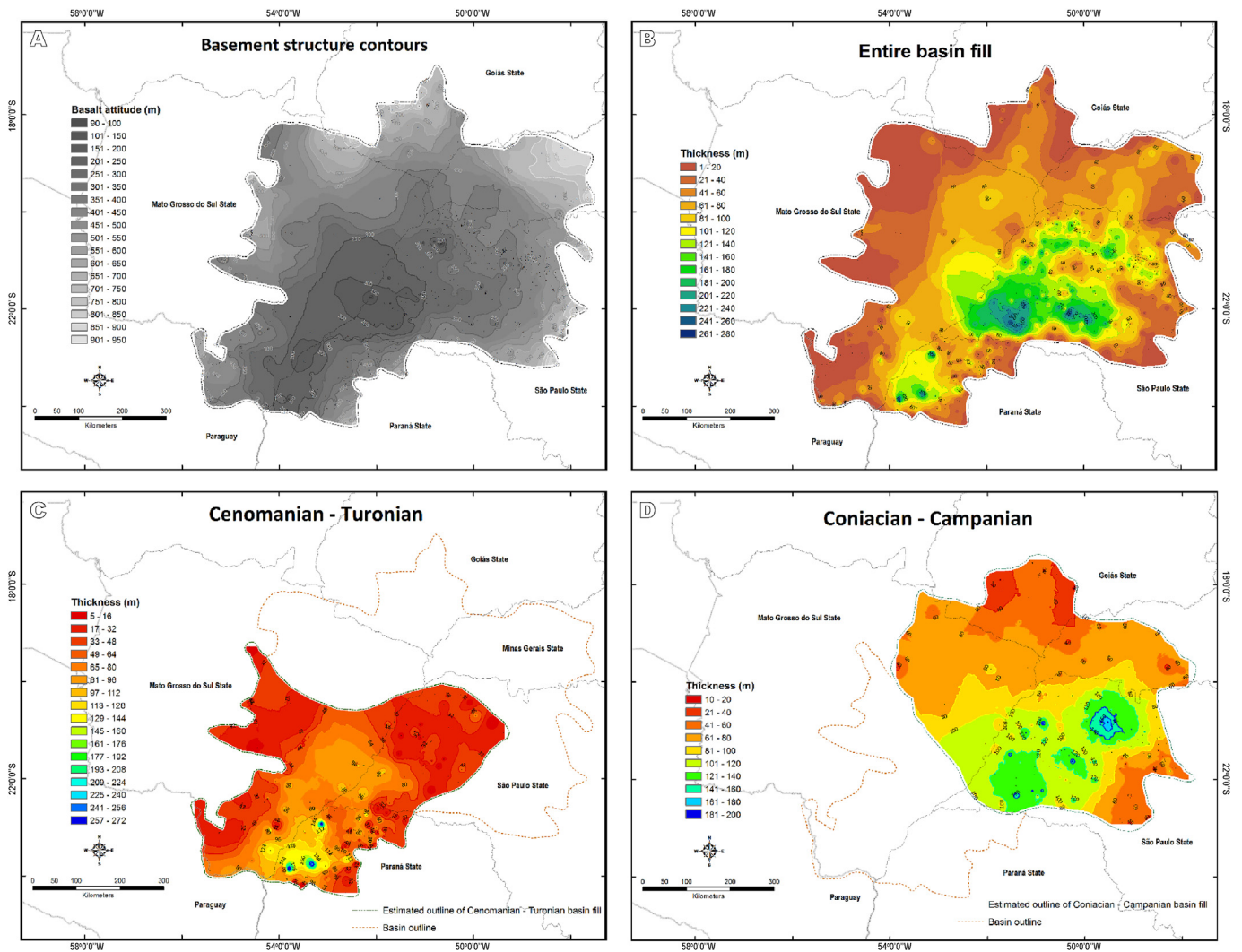
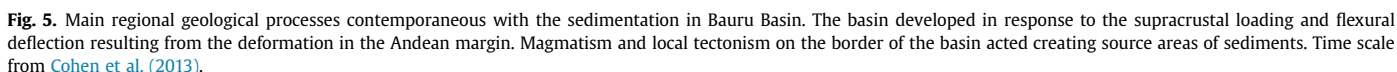


Fig. 4. A. Basement structure contours of the Bauru Basin based on the contact with subjacent basalts of the Serra Geral Formation. B. Isopach map of the preserved entire basin fill. The basin-fill is asymmetric, with two major depocenters (Northwest Paraná State and West São Paulo State). C. Cenomanian–Turonian stratigraphic interval based on the thickness of the Caiuá and Pirapozinho formations. D. Coniacian–Campanian interval, taking in account the thickness of the Santo Anastácio, Araçatuba, Birigui, São José do Rio Preto, Uberaba e Adamantina formations.



Finally, the Maastrichtian-Paleocene sequence is rich in basaltic and sedimentary rock clasts of the Paraná Basin, mostly from east and north adjacent source areas (Riccomini, 1997b; Gravina et al., 2002).

The opening of the South Atlantic Ocean and the westward migration of the South American Plate resulted in a profound change in the tectonic setting of the South American Plate. Sedimentation in Bauru Basin occurred simultaneously with the early stages of the Andean uplift during the Late Cretaceous (Mochica

Herein, we propose that the Bauru Basin developed in response to supracrustal load and flexural deflection resulting from the beginning of the early Andean orogeny, during the westward absolute motion of the South American Plate and subduction of the Farallon Plate. The Bauru Basin is in the back-bulge province of a retroarc foreland system developed adjacent to the western margin of South America (Fig. 6). The deposits of the Bauru Basin did not develop uniformly throughout the entire basin, with evidence of migration of its depocenter. The Andean foredeep basins also migrated eastward during the Peruvian pulses of the Andean orogeny (Sempere et al., 1997). This event occurred after the migration of the orogenic front during the pulses of orogeny.

Before the development of the Bauru Basin, the South America Plate was subjected to an extensional regime during most of the Triassic to Early Cretaceous, which was conditioned by the breakup of Gondwana (Sempere, 1995; Ramos, 2010). This tectonic process created regional tectonism, magmatism and creation of accommodation in sedimentary basins. In the west margin, grabens and back-arc basins have been established associated with reactivated basement structures (Jailard et al., 2000; Ramos, 2009; Zeffass et al., 2004, 2005; Sempere, 1995; Sempere et al., 1997; Maloney

et al., 2013). In the Solimões Basin, an echelon arrangement of folds and faults developed during the Late Jurassic (Caputo (1991). In the region where the Andean Basin later developed, were deposited the eolian deposits of the Ichoa and Yantara formations in Bolivia (Moretti et al., 1996) and the upper portion of the Juruá Mirim Formation in the Acre Basin (Fig. 7) (Feijó and Souza, 1994; Cunha, 2007). The fluvio-eolian Triassic or Jurassic deposits of the Pirambóia Formation and the eolian Late Jurassic–Early Cretaceous deposits of the Botucatu Formation were deposited in the Paraná Basin (Caetano-Chang, 1997; Assine et al., 2003; Milani et al., 2007a, b; Soares et al., 2008). The Parecis Basin was also submitted to extensional tectonics that generated accommodation for the eolian Jurassic deposits of the Rio Ávila Formation (Bahia et al., 2006). In addition, this Early Cretaceous extensional regime resulted in the development of a the Salta rift system in northern Argentina (Salfity and Marquillas, 1994; Marquillas et al., 2005; Ramos, 2009; Siks and Horton, 2011; Maloney et al., 2013), whose distribution has been associated with the triple junction involving the Amazonian, Arequipa and Pampia cratons (Comínguez and Ramos, 1995).

During this extensional regime, magmatism occurred in most sedimentary basins of South America (Fig. 7), including the Anari (198.0 ± 0.8 Ma) and Tapirapuã (196.6 ± 1.8 Ma) magmatic sub-provinces of the Parecis Basin and the Penatecaua magmatism in the Solimões and Amazonas basins (Marzoli et al., 1999; De Min et al., 2003; Wanderley Filho et al., 2007; Cunha et al., 2007), all belonging to the extensive magmatism of the Central Atlantic Magmatic Province (CAMP) (Marzoli et al., 1999; De Min et al., 2003; Bahia et al., 2006). In Southeast South America and West Africa, the Paraná-Etendeka Large Igneous Province (LIP) extruded over 1.2 m.y. at 134.6 ± 0.6 Ma (Thiede and Vasconcelos, 2010). In Brazil, the LIP is characterized by the Serra Geral Formation, covering most areas of the sedimentary rocks of the Paraná Basin, which are thicker than 1500 m in the Bauru Basin area (Zalán et al., 1990; Milani, 2003). These large continental flood basalts have been assumed as a result of decompressional melting of abnormally hot mantle brought to the base of the lithosphere by plumes (White and McKenzie, 1995) and are usually related to the Tristan da Cunha plume (Bizzi and Vidotti, 2003). However, paleomagnetic reconstructions and geochemical and isotopic data do not support this hypothesis (Ernesto, 2005; Ernesto et al., 2002; Rocha-Júnior et al., 2012), and the LIP has been linked to local hotter mantle conditions due to the combined effects of edge-driven convection and large-scale mantle warnings (Rocha-Júnior et al., 2013).

In this setting, during the final breakup of western Gondwana,

the South Atlantic Ocean evolved as a branch of a Jurassic–Cretaceous intraplate rift between the African and South American plates (Heine et al., 2013). The breakup of the southeastern Brazilian continental margin occurred in steps during the Early Cretaceous. In different segments of the Pelotas Basin (Fig. 1), this process occurred from the early Barremian to late Aptian (Stica et al., 2014). In the Santos and Campos basins, this process occurred near 113–115 Ma (Moreira et al., 2007; Winter et al., 2007; Stica et al., 2014). The final breakup between South America and Africa occurred in the equatorial Atlantic domain at 103 Ma (Heine et al., 2013).

3.2. Development of the retroarc foreland system

3.2.1. Initial stage - Cenomanian to Turonian

The breakup between South America and Africa in the late Albian began the westward absolute motion of the South American Plate (Jaillard and Soler, 1996; Ramos, 1999, 2000, 2009; Jaillard et al., 2000; Folguera et al., 2011; Boekhout et al., 2012; Maloney et al., 2013) and increased the velocity of the subduction of the oceanic lithosphere beneath the Andean margin (Ramos, 2009; Folguera et al., 2011). The subduction began at a steep angle, but in some segments of the Andes, evidence exists of shallow subduction by the Late Cretaceous (Jaillard et al., 2000; Ramos, 2009; Folguera et al., 2011; Spagnuolo et al., 2012).

The start of compressional deformation in Central Andes occurred at the base of the Late Cretaceous, with the beginning of the Andean uplift younger than 98.6 Ma (Tunik et al., 2010), characterizing the Mochica orogenic phase in the Central segment (Mégard, 1984; Jaillard et al., 2000). This tectonic phase is the earliest major compressional Cretaceous event in the Central Andes and has been interpreted as responsible for significant crustal shortening (Jaillard, 1994). Several studies developed in different areas of the Andean orogen have also described the initial uplift of this subduction-related orogen at Cenomanian (Mégard, 1984; Jaillard, 1994; Jaillard et al., 2000; Mosquera and Ramos, 2006; Jaimes and Freitas, 2006; Zamora Valcarce et al., 2006; Giulio et al., 2012; Pffiffer and Gonzalez, 2013; Fennel et al., 2015).

The Mochica orogeny is responsible for the initial development of the foredeep Andean Basin from northern Peru to south of Bolivia, parallel to the Andean margin (Fig. 8) (Reyes, 1972; Jaillard and Sempere, 1991; Sempere et al., 1997). In the south, a set of arches formed a boundary between the Andean Basin and the Salta Basin in northern Argentina. It is widely accepted that the Salta

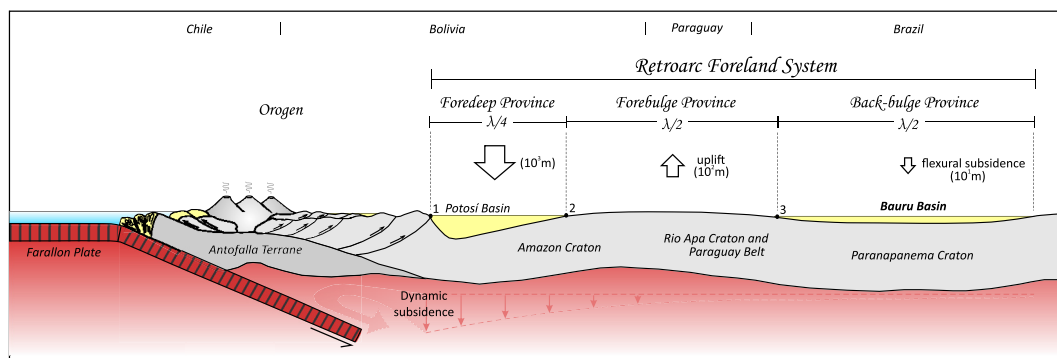


Fig. 6. The South American retroarc foreland system developed in Late Cretaceous. The flexural profile is partitioned into the provinces foredeep, forebulge and back-bulge, being the Bauru Basin developed in the back-bulge province. The amplitude of deflections is expected to decrease by three orders of magnitude (10^3 m– 10^1 m) with distance from the orogen, and the rate of dynamic subsidence exponentially decay with distance (Catuneanu et al., 1997a, b; Catuneanu, 2004). The boundary between the Antofalla Terrane and the Amazon Craton constitutes a paleosuture and crustal weakness zone inherited from Mesoproterozoic evolution of the Greenville Orogen (Jaillard et al., 2000). Subduction model of the Farallon Plate, faults and previous suture of the Antofalla Terrane from Ramos (2009). λ = flexural wavelength; 1 = orogenic front; 2 and 3 = flexural hinge lines. No vertical scale.

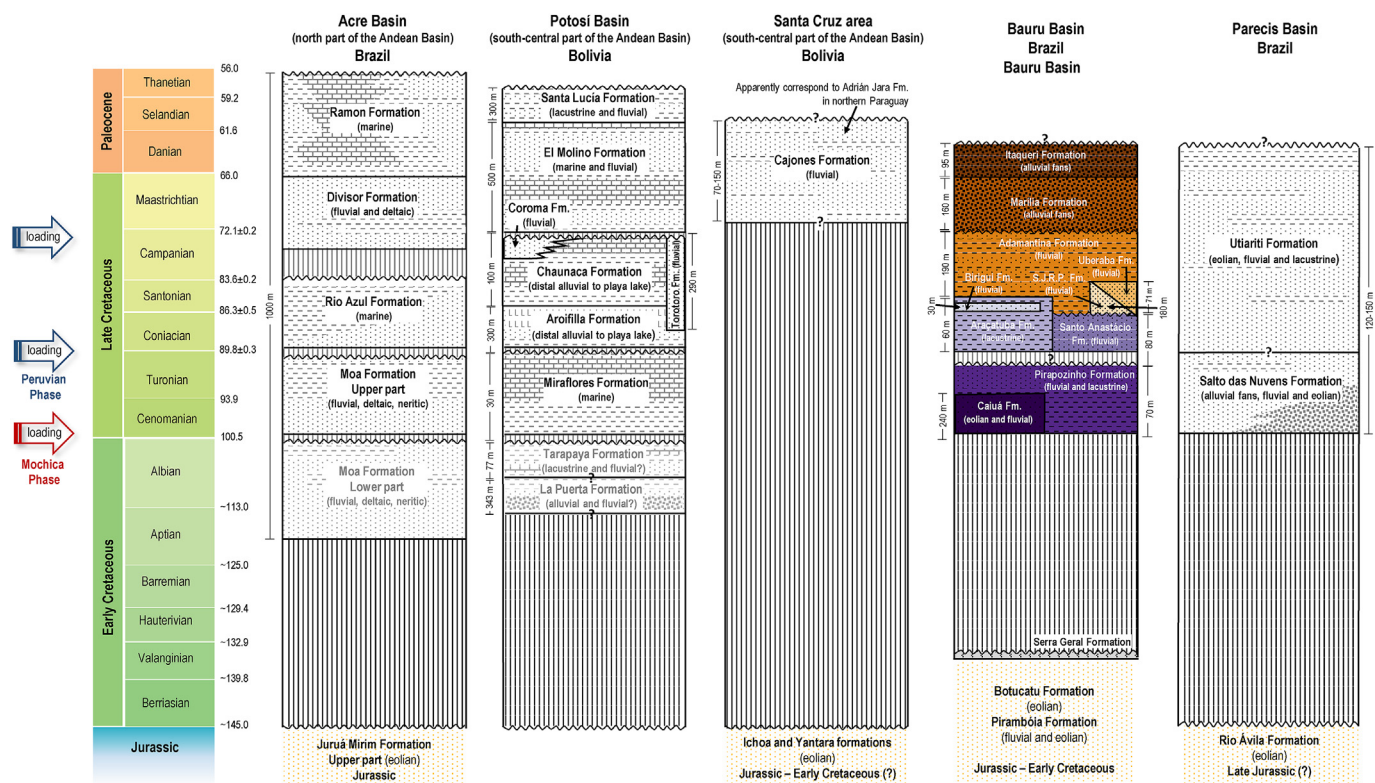


Fig. 7. Generalized stratigraphic table for the retroarc foreland system developed in South America during the Late Cretaceous. Acre and Potosí basins illustrate the foredeep province. Cajones Formation were deposited in the forebulge. Bauru and Parecis basins exemplify the back-bulge province. Andean Basin stratigraphy and maximum thickness in Bolivia compiled from Reyes (1972), Rouchy et al. (1993), Welsink et al. (1995), Moretti et al. (1996, 2002), Camoin et al. (1997), Jaillard and Sempere (1991), Sempere et al. (1997), Deconinck et al. (2000), Jaillard et al. (2000), Gayet et al. (2001), Kusiak and Zubieta Rossetti (2003) and Jimenez et al. (2009). Acre Basin from Feijó and Souza (1994) and Cunha (2007). Parecis Basin from Bahia et al. (2006) and Batezelli et al. (2014). Time scale from Cohen et al. (2013).

Basin resulted from the Early Cretaceous extensional regime, with sedimentation in a series of fault-bounded basins related either to the opening of the South Atlantic or back-arc extension (Salfity and Marquillas, 1994; Siks and Horton, 2011; Becker et al., 2015). As a result, from the Valanginian to Campanian, the sedimentation in the Salta Basin has been assumed as a rift phase (Hernandez et al., 1999; Masferro et al., 2003; Siks and Horton, 2011), and the Maastrichtian–Paleocene sedimentation has been attributed to post-rift thermal subsidence, which began with the Yacoraite Formation (Salfity and Marquillas, 1994; Comínguez and Ramos, 1995; Masferro et al., 2003; Siks and Horton, 2011; Marquillas et al., 2011). Usually, only the Cenozoic sequence of the Salta Basin has been interpreted as a foreland basin (Masferro et al., 2003; DeCelles et al., 2011; Siks and Horton, 2011; Becker et al., 2015).

A marine transgression in the Andean foredeep basin, from the north toward the south, occurred in the Cenomanian (Jaillard, 1994; Camoin et al., 1997; Martinez and Mamani, 1995; Jaillard and Sempere, 1991; Sempere, 1995; Sempere et al., 1997) and is represented (for example) by the upper portion of the Moa Formation in the Acre Basin (Feijó and Souza, 1994; Cunha, 2007) and the Miraflores Formation in the Potosí Basin (Fig. 7) (Jaillard and Sempere, 1991; Sempere, 1995; Sempere et al., 1997). The Cenomanian age of the Miraflores Formation is well established by the record of *Neobolites vibrayeanus* (Reyes, 1972; Jaillard and Sempere, 1991; Wiese and Schulze, 2005). However, the ages of the subjacent non-fossiliferous La Puerta and Tarapaya formations are controversial (Reyes, 1972; Jaillard et al., 2000), raising doubts related to the relationship of these units with the foreland system or the previous extensional regime. In southern central Bolivia, the total Cretaceous succession is approximately 1200 m thick and is composed of

marine and nonmarine deposits (Sempere et al., 1997).

Late Cretaceous sedimentation also occurred in cratonward areas parallel to the Andean Basin characterizing the Bauru, Parecis and Solimões basins (Figs. 7 and 8). The Cretaceous of the Parecis Basin is composed of nonmarine deposits of the Salto das Nuvens and Utariti formations (Parecis Group), occupying a relative wide area and with a maximum preserved thickness of approximately 150 m (Bahia et al., 2006). The fossil record provides evidence of Late Cretaceous age for the undifferentiated Parecis Group consisting of theropods (Kellner and Campos, 2002; Bittencourt and Langer, 2011), the sauropod *Gondwanatitan* (Franco-Rosas et al., 2004), and crocodyliformes similar to *Mariliasuchus* and *Noto-suchus* (Marconato, 2006). The underlying eolian deposits of the Rio Ávila Formation are assumed to be Late Jurassic (Batezelli et al., 2014), although no fossil record exists to support this hypothesis. The sedimentation of the thick Alter do Chão Formation in the Solimões (1250 m) and Amazonas (1000 m) basins began at the same time. According to Silva et al. (2003), the preservation of these fluvial and alluvial fan deposits in the Solimões Basin is related to the Andean orogeny events. However, the age assignment of the Alter do Chão Formation is controversial, i.e., early Cenomanian–Maastrichtian (Daemon and Contreiras, 1971) or late Aptian for the lower portion and Cenomanian for the middle portion (Dino et al., 1999), with a lack of biostratigraphic definition for its upper portion (Dino et al., 2012; Mendes et al., 2012).

There is no record of Cenomanian–Turonian sedimentation in the region between the Potosí Basin in Bolivia (south-central portion of the Andean Basin) and the Bauru and Parecis basins in Brazil (Figs. 7 and 8), which indicates the existence of an uplifted area that was not received or did not preserve sediments at this

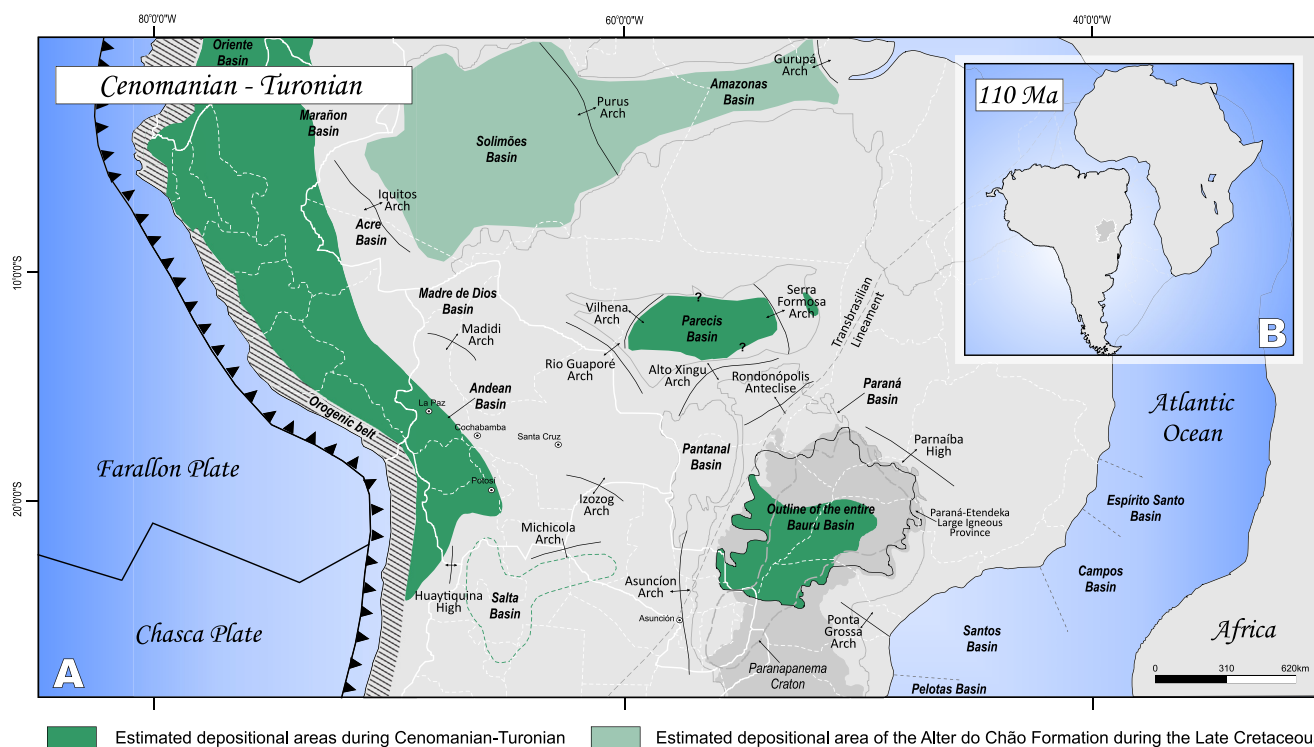


Fig. 8. A. The retroarc foreland system at Cenomanian-Turonian, with South America fixed in present-day coordinates. Orogenic belt and estimated extent of the Andean Basin from Jaillard and Arnaud-Vanneau (1993), Pindell and Tabbutt (1995), Jaillard et al. (2000) and Mpodozis et al. (2005). Plate configuration from Maloney et al. (2013). Outcrop area of the Salto das Nuvens Formation (Parecis Basin) from Bahia et al. (2006). Extent of the Alter do Chão Formation from Caputo (1991) and Mendes et al. (2012). Deposition in the Bauru Basin following the isopach maps presented in this work (Fig. 4C). Position of arches and highs compiled from Comínguez and Ramos (1995), Riccomini et al. (1997), Mpodozis et al. (2005), Bahia et al. (2006), Milani et al. (2007b), Marquillas et al. (2011), Curto et al. (2014) and Lisboa et al. (2013). B. Relative position of South America and Africa at 110 Ma, with Africa fixed in present-day orientation (Heine et al., 2013).

time.

3.2.2. Migration of the system – Coniacian to Campanian

By the late Turonian to earlier Coniacian, a new event of Andean orogeny occurred in the Central Andes and portions of the North Andes, which characterize the first event of the Peruvian Phase (Noblet et al., 1996; Jaillard and Soler, 1996; Sempere et al., 1997; Jaillard et al., 2000). The Peruvian Phase extended from the Coniacian to early Paleocene and was characterized by relatively short deformation events: the Turonian-Coniacian boundary (~88 Ma), Santonian (~85 Ma), late Campanian (~80–75 Ma) and late Maestrichtian (~70–65 Ma) alternating with longer intervals of tectonic quiescence (Jaillard, 1994; Jaillard and Soler, 1996; Sempere et al., 1997; Jaillard et al., 2000). The subduction that began near the Aptian-Albian boundary in the west margin of South America accelerated after 90–95 Ma, simultaneously with the acceleration of the absolute motion of the South American plate during the opening of the South Atlantic Ocean (Folguera et al., 2011; Ramos, 1999, 2000, 2009; Maloney et al., 2013).

For Jaillard et al. (2000), the Turonian-Coniacian boundary is marked by the end of carbonate sedimentation in the Andean Basin and is replaced by marine shales in the north portion and red shales and siltstones in the south with abundant evaporites layers (Fig. 7). A significant tectonic shortening occurred in the western areas of northern Chile, and consequently, a remarkable increase in subsidence occurred in the Andean foredeep basin (Sempere et al., 1997; Jaillard et al., 2000). The sudden increase in subsidence is defined in the Potosí Basin by the accumulation of a thick sequence initiated with the deposition of the Aroifilla Formation in the Coniacian and followed by the Chaunaca, Torotoro and Coroma formations

(Jaillard and Sempere, 1991; Sempere, 1995; Sempere et al., 1997; Jaillard et al., 2000). At the same time, the deposition of the marine deposits of the Rio Azul Formation began in the Acre Basin (Cunha, 2007). In both basins, unconformities separate the Cenomanian-Turonian and Coniacian-Campanian successions (Sempere, 1995; Cunha, 2007). In the Coniacian, the arc migrated eastward in Peru together with the depocenter of the Andean Basin (Fig. 9) (Jaillard, 1994; Jaillard et al., 2000).

For the Parecis Basin, the exposition area of the Utariti Formation might indicate that migration of its depocenter (Fig. 9), but there is insufficient data available to support this hypothesis. Similarly, there is no available information that allows reconstruction of the basin-fill geometry of the Alter do Chão Formation in the Solimões Basin. However, Caputo (1991) states that the Iquitos Arch between Acre and Solimões basins resulted from flexural uplift in response to orogenic loading of the Andes, and this structure migrated toward the Solimões Basin during its Cretaceous and Cenozoic sedimentation (Fig. 9).

In the northern Peruvian margin, an incipient thrusting and progressive emergence occurred in the early Santonian (Jaillard, 1994; Jaillard et al., 2000). However, in southern Peru and Bolivia, the stratigraphic data are insufficient to demonstrate the occurrence and duration of this event (Jaillard et al., 2000; Sempere et al., 1997), and it appears equally unrecorded in the Bauru Basin.

From the point of view of Sempere (1995), the latest Turonian-earlier Coniacian orogenic pulse of the Peruvian Phase in Central Andes was followed by a stage of orogenic quiescence in Campanian, which is viewed in the Andean Basin as the decrease in the rate of sedimentation and progradation of immature sands from the west (Sempere, 1995). In the Bauru Basin, the fluvial

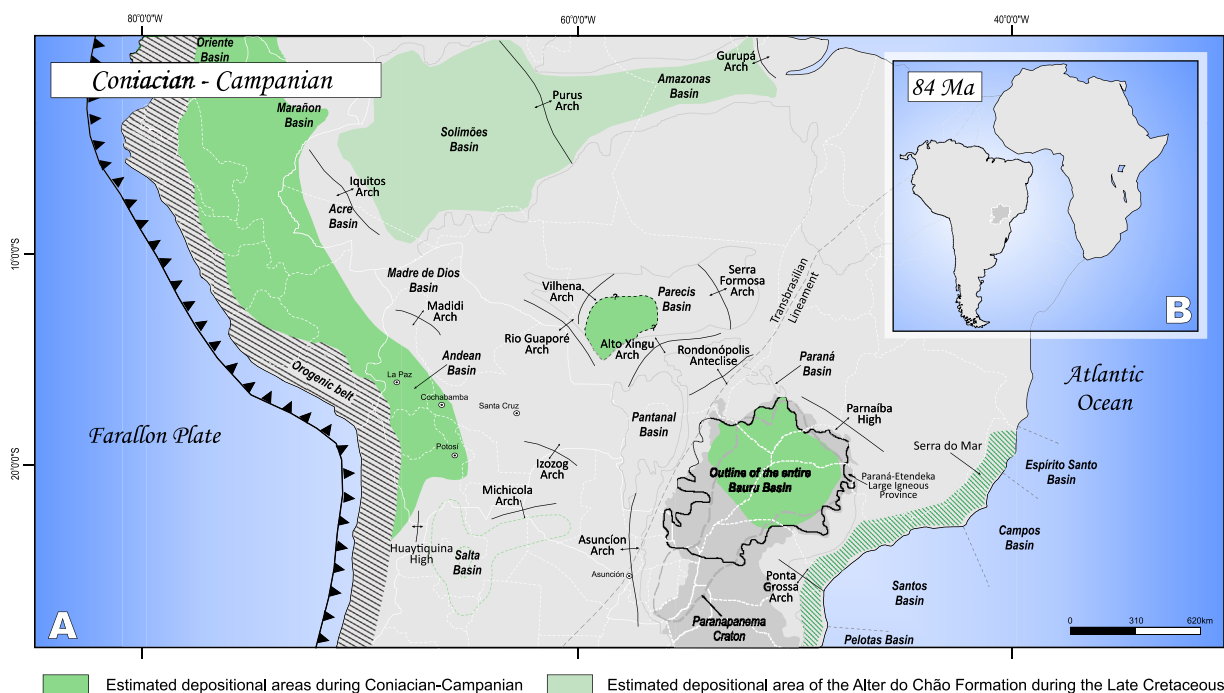


Fig. 9. A. The retroarc foreland system at Coniacian-Campanian, with South America fixed in present-day coordinates. Orogenic belt and estimated extent of the Andean Basin from Sempere (1995), Pindell and Tabbutt (1995), Sempere et al. (1997) and Mpodozis et al. (2005). Plate configuration from Maloney et al. (2013). Outcrop area of the Utiariti Formation (Parecis Basin) from Bahia et al. (2006). Extent of the Alter do Chão Formation from Caputo (1991) and Mendes et al. (2012). Deposition in the Bauru Basin following the isopach maps presented in this work (Fig. 4D). Serra do Mar uplift from Chang et al. (1992), Gallagher et al. (1994), Zalán and Oliveira (2005) and Karl et al. (2013). See Fig. 8 for references about structures. B. Relative position of South America and Africa at 84 Ma, with Africa fixed in present-day orientation (adapted from Nurnberg and Müller, 1991).

meandering deposits of the Adamantina Formation were deposited by the Campanian.

Similar to the Cenomanian-Turonian interval, no Coniacian-Campanian sediment is preserved in the west of Bolivia and northwest of Paraguay, between the Andean Basin and the Bauru Basin (Figs. 7 and 9).

3.2.3. End of sedimentation in the Bauru Basin – Maastrichtian to Paleogene

After a relatively long interval of tectonic quiescence, a new contractional pulse of the Peruvian orogeny occurred in the late Campanian, with thrusting and uplift recorded in the Central Andes (Jaillard, 1994; Jaillard and Soler, 1996; Sempere et al., 1997; McQuarrie et al., 2005; Ramos, 2009). In the North Andes, accretion of oceanic crust occurred at 75–65 Ma along the Ecuadorian and Colombian margins, interpreted as multi-episodic collisions of the leading edge of the Caribbean Plateau with the South American Plate (Ramos, 1999, 2009; Ramos and Alemán, 2000). Moreover, in the Late Campanian, the location of the magmatic arc in northern Chile significantly shifted eastward, which has been interpreted as a decrease of the slab subduction angle (Jaillard and Soler, 1996; Jaillard et al., 2000). This latest Late Cretaceous shallowing of the subduction is also interpreted in the southern Central Andes and the South Andes (Ramos, 2009; Folguera et al., 2011; Spagnuolo et al., 2012).

The crustal loading event is represented in the Andean Basin by a high rate of subsidence near the Campanian-Maastrichtian boundary, which declined in the late Maastrichtian when the Maastrichtian–lower Paleocene sequence was deposited (Sempere, 1995; Jaillard and Soler, 1996; Sempere et al., 1997). In addition, the depocenter of the Andean Basin shifted eastward (Fig. 10) (Jaillard et al., 2000).

According to Sempere (1995), the sedimentation rate increased

in the Potosí Basin near the Campanian-Maastrichtian boundary, and an influx of sand occurred from the west (Brazilian Shield), attributed to thrusting in the Andes accompanied by uplift of a forebulge in the west of Bolivia. A new cycle of marine transgression simultaneously occurred in the Andean Basin, extending southward to the Salta Basin. This transgression is characterized by the base of the El Molino Formation in the Potosí Basin and its equivalent Yacoraite Formation in the Salta Basin (Sempere, 1995; Gayet et al., 2001). In the Acre Basin, the Maastrichtian-Paleocene sequence is represented by the basal fluvial deposits of the Divisor Formation, the marine shales of the Rio Azul Formation (Fig. 7), and the marine transgressive facies of the Ramon Formation (Cunha, 2007). From the point of view of DeCelles and Horton (2003) and McQuarrie et al. (2005), the El Molino and Santa Lucía formations (Potosí Basin) might have been deposited in a backbulge province of a retroarc foreland system. Conversely, Sempere et al. (1997) propose that this Upper Cretaceous-Paleocene sequence was deposited in a foredeep basin, and Mpodozis et al. (2005) noted the absence of a forebulge separating the Bolivian Potosí Basin from the deposits of the Purilactis Group close to the zone of active deformation.

During the Maastrichtian to Paleocene, the sedimentation in the studied area was markedly different from that of the preceding periods, with the preservation of the Cajones Formation in the area between the Andean Basin and the Bauru Basin (Figs. 7 and 10). An unconformity separates the Cajones Formation from the Jurassic extensional deposits below (Moretti et al., 1996). In the region of Santa Cruz in Bolivia, the Cajones Formation is interpreted as the product of fluvio-deltaic environment with a marine influence from the northwest and is laterally equivalent to the El Molino Formation (Kusiak and Zubieta Rossetti, 2003). The Cajones Formation is dated to the Maastrichtian based on fish fossils that also occur in the El Molino (Maastrichtian–Danian) and Yacoraite (Maastrichtian)

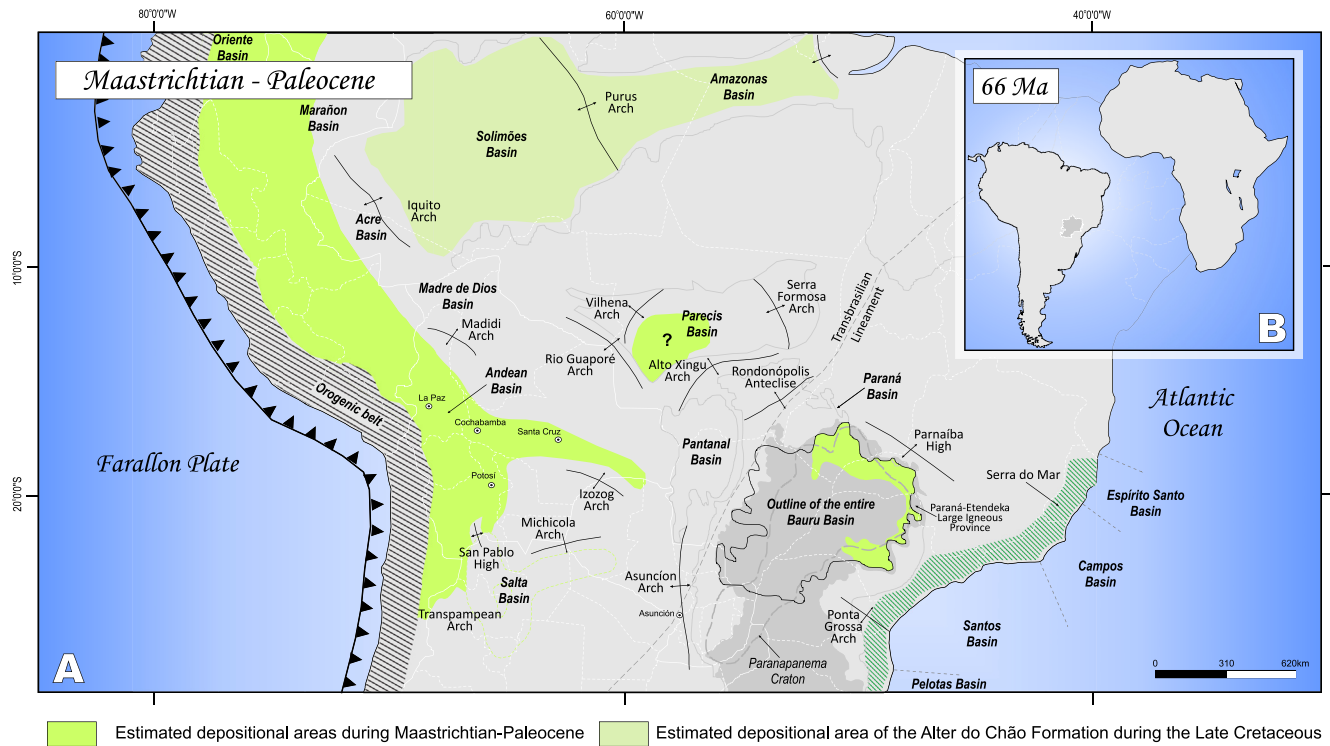


Fig. 10. A. The retroarc foreland system at Maastrichtian-Paleocene, with South America fixed in present-day coordinates. Orogenic belt and estimated extent of the Andean Basin from Reyes (1972), Gayet et al. (1993), Sempere (1995), Pindell and Tabbutt (1995), Sempere et al. (1997), Jaillard et al. (2000), Gayet et al. (2001), Kusiak and Zubieta Rossetti (2003), and Mpodozis et al. (2005). Plate configuration from Maloney et al. (2013). Outcrop area of the Utiariti Formation (Parecis Basin) from Bahia et al. (2006). Extent of the Alter do Chão Formation from Caputo (1991) and Mendes et al. (2012). Deposition in the Bauru Basin following outcrop area of the Marília and Itaqueri formations (Fig. 2). See Fig. 8 and 9 for references about structures and Serra do Mar uplift. B. Relative position of South America and Africa at 66 Ma, with Africa fixed in present-day orientation (Scotese, 2001).

formations (Aguilera et al., 1989; Gayet et al., 2001; Kusiak and Zubieta Rossetti, 2003; Cónsole Gonella et al., 2009). The Adrián Jara Formation in northwest Paraguay corresponds to the Bolivian Cajones Formation (Gómez Duarte, 1986).

In the Parecis Basin, the age range of the Utiariti Formation extends to the early Paleocene (Bahia et al., 2006); but in the Solimões Basin, the upper limit of the Alter do Chão Formation is controversial (Dino et al., 1999).

According to Jaillard et al. (2000), widespread unconformities are recorded near the Maastrichtian-Paleocene boundary, with coarse sedimentation and deformation in the Central Andes suggesting that a tectonic event occurred by this time, but this possible event is still poorly known.

3.2.4. The system during the Cenozoic

The retroarc foreland system evolved during the Late Eocene to Miocene into the Chaco Basin (Fig. 1), with a Cenozoic sequence that is approximately 2.5 km thick, predominantly consisting of fluvial and lacustrine deposits (DeCelles and Horton, 2003). Its evolution is related to a clockwise rotation of the South American Plate in the Eocene and the orogenic pulses of the Incaic and Quechua tectonic phases (Mégard et al., 1984; Noblet et al., 1996; Jaillard and Soler, 1996).

A full discussion of the development of the system during the Cenozoic is beyond the scope of this contribution, which is focused on the depositional time of the Bauru Basin. The Cenozoic evolution has been intensively studied by many authors (i.e. Lamb et al., 1997; Horton and DeCelles, 1997; Jaillard et al., 2000; DeCelles and Horton, 2003; McQuarrie et al., 2005; Uba et al., 2006; Sempere et al., 1997; Jiménez et al., 2009; Prezzi et al., 2009; DeCelles et al., 2011; Maloney et al., 2013; Cohen et al., 2015; Engelder and

Pelletier, 2015; Quade et al., 2015). Certain studies have demonstrated that the system also migrated toward the craton during the Cenozoic, coupled with the eastward shift of the orogenic front and the development of the Eastern Cordillera (Horton and DeCelles, 1997; Jaillard et al., 2000; DeCelles and Horton, 2003; McQuarrie et al., 2005; Uba et al., 2006; Prezzi et al., 2009). Consequently, the Miocene deposits of the Pantanal Basin were deposited in the back-bulge province of the system and still constitute the back-bulge of the modern foreland system (Horton and DeCelles, 1997; McQuarrie et al., 2005).

The thickness of the Pantanal basin-fill exceeds 400 m (Assine, 2003; Assine and Soares, 2004), which is not consistent with the thickness expected in back-bulge depozones (Horton and DeCelles, 1997; DeCelles and Giles, 1996). Therefore, additional mechanisms of subsidence have been considered to explain this thickness, including the dynamic effects of the subduction and isostatic effects of high-density rocks underlying the Pantanal Basin (Horton and DeCelles, 1997).

The Cenozoic sedimentation in Acre and Solimões basins has been associated with the Andean orogeny (Feijó and Cunha, 1994; Cunha, 2007; Cunha et al., 2007) and is characterized by fluvial meandering and lacustrine deposits of the Solimões Formation that reach 2200 m of thickness in the Acre Basin (Cunha, 2007) and 1800 m in the Solimões Basin (Wanderley Filho et al., 2007).

4. Discussion

The mechanisms of sediment accumulation and preservation in the Bauru Basin have been poorly understood. Integration of its characteristics (basin-fill geometry, thickness, depositional systems and age) with the available information on the overall tectonic

setting and adjacent coexistent basins has aided in characterizing its subsidence pattern and has contradicted previous hypotheses.

The depositional systems in the Bauru Basin are nonmarine, and the caliber of sediment is predominantly fine. Local accumulations of coarse-grained sediments (alluvial fans) are limited to the Maastrichtian to early Paleocene portion of the succession in the north and east borders of the basin. As selected sedimentological studies previously noted (Soares et al., 1980; Paula e Silva et al., 2009), no evidence exists that faults controlled the initial sedimentation. Although the extensional regime that prevailed from the Jurassic to Early Cretaceous was responsible for widespread rifting in South America, the sedimentation in the Bauru Basin began in the Cenomanian with the Caiuá and Pirapozinho formations, as proposed herein. Hence, evidence was not found to support the preceding proposals of grabens and rifts.

Among the mechanisms that might potentially be important in basin development, the dynamic effects of plumes can be rejected as a basin-forming mechanism for the Bauru Basin. Previously, alkaline magmatism in structures near the border of the basin (Rondonópolis Anticline, Paranaíba High and Asunción Arch) and the uplift of Serra do Mar were linked to the impact of the Trindade and Tristan da Cunha mantle plumes (Riccomini et al., 2005; Gibson et al., 1995; Van Decar et al., 1995; Brod et al., 2005; Zalán and Oliveira, 2005). Posteriorly, certain studies demonstrated a disconnection between the isotopic and geochemical compositions of the alkaline magmatic rocks and these plumes (Ruberti et al., 2005; Comin-Chiaramonti et al., 1997, 1999; Comin-Chiaramonti et al., 2014; Gomes et al., 2013). Moreover, paleomagnetic reconstructions have demonstrated that the Tristan da Cunha plume was located too far from the Paraná-Etendeka LIP at the time of its eruption (Valanginian), and the Trindade Plume was located approximately 1000 km to the north of Serra do Mar and Paranaíba High at the moment of the Late Cretaceous alkaline magmatism (Ernesto et al., 2002; Ernesto, 2005). These alkaline magmatic events could have resulted from decompressional melting due to local lithospheric extension and reactivation of normal pre-existing faults in the basement reactivated as transgressive structures related to the opening of the Atlantic Ocean (Almeida and Carneiro, 1998; Hasui and Haralyi, 1991). Consequently, alkaline magmatism adjacent to the Bauru Basin does not indicate that the basin was subjected to an extensional regime in Late Cretaceous and/or its subsidence related to mantle plumes. In addition, the uplift of these extrabasinal arches is not responsible for the basin subsidence, in opposition to previous hypotheses (i.e. Batezelli, 2015). The alkaline magmatic events are local occurrences in regions that were uplifted simultaneously with the sedimentation in Bauru Basin and might have acted as source areas of sediments.

Similarly, although the initial pulses of the Serra do Mar uplift were almost simultaneous with the beginning of the Coniacian sedimentation in Bauru Basin (Figs. 5 and 9), they do not have a direct relationship with its subsidence. For Chang et al. (1992) and Gallagher et al. (1994), the uplift of the Serra do Mar was caused by isostatic rebound due to unloading of the lithosphere during the rifting of South Atlantic, probably associated with underplating. This continental marginal uplift is deeply associated with the opening of the South Atlantic Ocean and cannot explain the creation of the Bauru Basin. During the climax of the Serra do Mar uplift, an expressive increase occurred in the sediment supply to the Santos Basin (Pereira e Feijó, 1994) and might have contributed similarly to the sedimentation of the Bauru Basin.

The Bauru Basin is positioned over the maximum preserved thickness of the basalts of the Paraná-Etendeka LIP or Serra Geral Formation (700–1000 m; Zalán et al., 1990) (Fig. 11). Therefore, certain authors have explained the Bauru Basin development as the result of thermal subsidence related to the cooling of these

extensive and thick lava flows (Zalán et al., 1990) and/or a flexural depression created by the load of the basalts (Riccomini, 1997b; Fernandes and Coimbra, 1996, 2000; Milani, 2003; Milani and De Wit, 2008; Batezelli, 2015). Leng and Zhong (2010) demonstrated that surface subsidence occurs over tens of millions of years before food basalt eruptions in LIPs, and during the basalt eruption, the loading from erupted rocks causes subsidence at the periphery of the eruption area that affects the environment for subsequent episodes of basalt eruption.

In contrast, Mariani et al. (2013) suggested that underplated material might have been the mechanism that drove the subsidence of the Bauru Basin due to subcrustal loading and cooling. Indeed, magmatic underplating appears to be common in LIPs (MacLennan and Lovell, 2002; Thybo and Artemieva, 2013), but the existence of underplating beneath the Bauru Basin is still disputed (Molina et al., 1999; An and Assumpção, 2006; Mariani et al., 2013). MacLennan and Lovell (2002) have shown that the addition of mafic magma to the lithosphere causes surface uplift of approximately 10% of the thickness of the magma sill. Due to solidification of the magma, rapid subsidence occurs at an order of magnitude of approximately half of the original uplift, and additional relatively slow subsidence results from cooling of the solid lithosphere after magma injection (MacLennan and Lovell, 2002). According to MacLennan and Lovell (2002), the time scale of solidification is controlled by the sill thickness, but even considering the thickness of 10 km of the underplated material suggested by Mariani et al. (2013), the overall subsidence expected should be on the time scale of a few million years.

In both cases, loading from the erupted rocks or subcrustal loading and cooling of underplated material, these hypothesis should take in account that subsidence is expected to occur during or immediately after these events. Because the LIP extruded in the Valanginian (Thiede and Vasconcelos, 2010) and sedimentation in Bauru Basin is assumed to have started in Cenomanian, there is a time span of approximately 34 Ma between these hypothesized basin-forming mechanisms and the beginning of sedimentation.

The key aspects of the Bauru Basin are that the isopach pattern of the preserved entire basin-fill shows regional closure surrounding central thick zones (Fig. 4B), with accumulation in a shallow and broad region suggesting that the sediment accommodation could have involved a component of flexural subsidence. Furthermore, the isopach map of the interval from the Cenomanian to Turonian shows a depocenter in northwestern Paraná State (Fig. 4C), whereas the interval from the Coniacian to Campanian shows a depocenter in northwestern São Paulo State (Fig. 4D). This pattern can be explained as a migration of the depocenter through time and suggests an episodic history of creation of accommodation.

The stratigraphic architecture of the Bauru and Potosí basins are in phase, and both are out of phase with the region between them (Fig. 7). The tectonic load event at early Cenomanian led to subsidence and the establishment of marine and distal alluvial environments in the Andean Basin, e.g., as characterized by the upper portion of the Moa Formation in Acre Basin and the Miraflores Formation in Potosí Basin (Jaillard, 1994; Camoin et al., 1997; Martinez and Mamani, 1995; Jaillard and Sempere, 1991; Sempere, 1995; Sempere et al., 1997). The accumulation in the Bauru Basin began concomitantly with this Cenomanian transgression in the Andean Basin with the lacustrine, fluvial and eolic deposits of the Pirapozinho and Caiuá formations. A similar situation is observed during the second loading event in the early Coniacian (Peruvian Phase), starting with the deposition of the Rio Azul and Aroifilla formations, respectively, in Acre and Potosí basins, and the Araçatuba and Santo Anastácio formations in Bauru Basin. In both the Andean and Bauru basins, eastward shifts of their

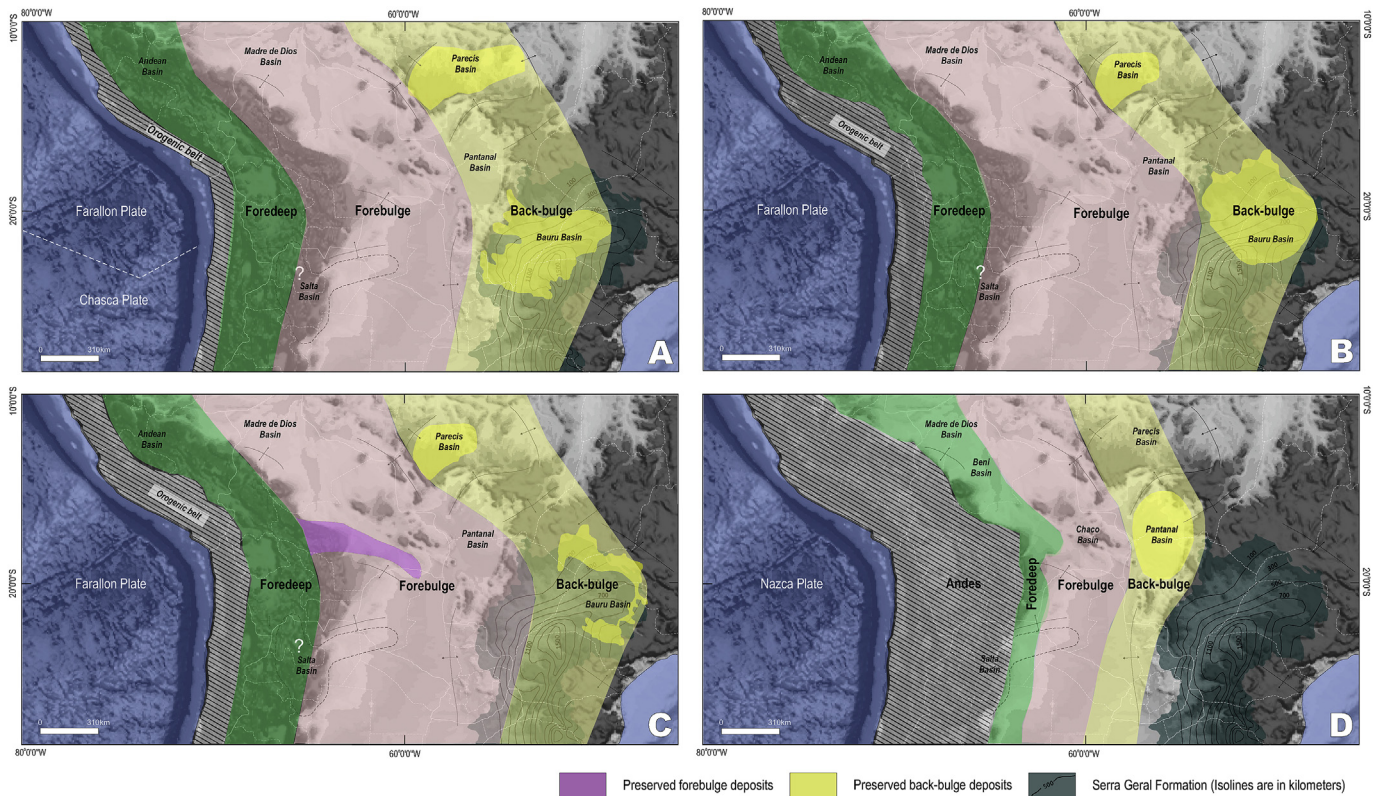


Fig. 11. Evolution of the retroarc foreland system. **A.** Initial development of the system. **B.** Cratonward migration of the whole system. **C.** Sedimentation in the forebulge resultant of the increased dynamic subsidence. **D.** Modern retroarc foreland system, after its eastward migration during the Cenozoic (following Horton and DeCelles, 1997; DeCelles and Horton, 2003; McQuarrie et al., 2005). Estimated position of the provinces based on preserved deposits of the Andean Basin and back-bulge basins (see references in Figs. 8–10). Isopach map of the Serra Geral Formation from Zalán et al. (1990).

depocenters are observed (Fig. 9). Again, the load event in Late Campanian is responsible for the creation of accommodation in the Andean Basin (i.e., the Divisor and El Molino formations) following a new cratonward migration of the orogenic front and the associated shift of the basin depocenter (Sempere, 1995; Sempere et al., 1997). At the same time, there is an apparent shift of the Bauru depositional area, with the northwest onlap pattern of the Marília and Itaqueri formations onto the cratonic margin (Fig. 10). The simultaneous shifting of the Andean and Bauru basins in the same direction after orogenic events and their reciprocal stratigraphy are characteristics indicating that supracrustal loading was also responsible for the subsidence in Bauru Basin. From the shape, thickness and cratonward position of the Bauru Basin related to the foredeep Andean Basin and the Late Cretaceous orogen, it is deducible that the Bauru Basin has evolved in the back-bulge province of a retroarc foreland system (Figs. 2 and 11). Therefore, the Bauru sedimentary fill corresponds to a first-order sequence, with basal contact changing the tectonic setting, i.e., from the Mesozoic extensional deposits of the Paraná Basin below to the foreland system above.

The forebulge in retroarc foreland systems is an elevated feature, and therefore, it is generally considered to be a zone of nondeposition or erosion, and a resulting unconformity has been used to track its position through time (DeCelles and Giles, 1996; Catuneanu, 2004). Consequently, the region in the west of Bolivia and northwest of Paraguay characterizes the forebulge depozone of this system, which has not received or preserved sediment from the Cenomanian to Campanian (Figs. 7–9).

Taking into account the extent of the foredeep Andean Basin, other contemporaneous deposits should be expected to exist along

the cratonic side of the system, similar to the back-bulge Bauru Basin. Indeed, the position of the Upper Cretaceous deposits of the Parecis and Solimões basins appear to indicate that these basins evolved in the back-bulge province of this retroarc foreland system (Figs. 7 and 11). For Caputo (1991), the Iquitos Arch between the Acre and Solimões basins resulted from flexural uplift in response to the orogenic loading of the Andean belt during the Cretaceous. The progressive cratonward onlap of the foredeep strata of the Acre Basin onto an unconformity in the Iquitos Arch indicates that this structure underwent eastward migration coeval with the Cretaceous and Cenozoic sedimentation (Caputo, 1991). Therefore, the Iquitos Arch might be a forebulge between the foredeep Acre Basin and the back-bulge Solimões Basin. Still, there is no accurate biostratigraphy for the Upper Cretaceous deposits of the Parecis and Solimões basins due to the scarcity of fossil records, and there are insufficient data to reconstruct their basin-fill geometry. Therefore, the evidence is inconclusive for demonstration that they are back-bulge basins.

Certain authors interpreted the Cretaceous accumulation in Potosí Basin as resulting from thermal subsidence related to a rift system (Welsink et al., 1995; Camoin et al., 1997; Viramonte et al., 1999; Deconinck et al., 2000; Jiménez et al., 2009), similar to the nearby Salta rift system in northern Argentina. However, several authors have shown that at least a portion of the upper Puca Group (Aroifilla, Chaunaca, Torotoro, Coroma, El Molino and Santa Lucía formations) might represent a foreland basin related to initial shortening of the Andes (Sempere et al., 1997; Horton and DeCelles, 1997; DeCelles and Horton, 2003; McQuarrie et al., 2005). These authors suggested that the high rate of sediment accumulation, the variable rates of sedimentation through time and the abrupt

cessation of sedimentation after the deposition of the Santa Lucía Formation are characteristics that differs from the expected gradual decrease of the subsidence rates predicted for rift models. In addition, the influence of the early Andean orogen on the Salta Group sedimentation in the Salta rift system remains unclear; it might not be initially controlled directly by the geodynamic process related to the orogenic belt (Salfity and Marquillas, 1994; Cominquez and Ramos, 1995; Hernandez et al., 1999; Masferro et al., 2003; Marquillas et al., 2011; DeCelles et al., 2011; Siks and Horton, 2011; Becker et al., 2015). This research is consistent with the previous interpretations in which the Potosí Basin is a component of the foredeep Andean Basin.

The Cenomanian–early Paleocene South American retroarc foreland system is comparable with other systems in flexural profile, size, preserved thickness and pattern of depositional shift through time, especially the Karoo and Western Canada basins, which are also formed on thick and Precambrian crusts with high flexural rigidity (Catuneanu et al., 1997b; Catuneanu et al., 1998; Miall et al., 2008). Along the dip, the relative proportion among the flexural provinces of the South American system is compatible with extant models and the proportions observed in the Karoo and Western Canada foreland systems (DeCelles and Giles, 1996; Catuneanu, 2004). The forebulge and the back-bulge are wider than the foredeep province (Figs. 6 and 11). The foredeep measures approximately a quarter of the wavelength, the forebulge is slightly longer than half of the wavelength, and the back-bulge is somewhat narrower than half of the wavelength. The flexural profile might be modified if the crust contains pre-existing weaknesses that allow the development of fault-controlled uplifts and depocenters (DeCelles and Giles, 1996; Catuneanu et al., 1997a, b). Accordingly, the slightly asymmetric flexural profile observed in the South American foreland system might be a consequence of heterogeneities of the basement.

The South American retroarc foreland system extends approximately 2005 km from the orogenic front, considering the preserved sedimentary fill of the interval from the Maastrichtian to Paleocene of the Bauru Basin (Fig. 11). Thus, the maximum size of this system along dip is wider than the North American Western Interior Foreland Basin with 865 km on the Western Canada Basin but is comparable with the Karoo retroarc foreland system. Although the back-bulge province of the Karoo system has not been completely understood, it is known that a portion of the sediments of this province is preserved in Tuli Basin, approximately 1550 km from the orogenic front (Catuneanu et al., 1998; Catuneanu et al., 2002; Bordy and Catuneanu, 2001; Catuneanu et al., 2005). Because the maximum extent of its foredeep reaches 415 km (Catuneanu et al., 1998), and this province represents a quarter of the wavelength of the sinusoidal flexural profile, the total extent of the Karoo system might be estimated at approximately 2075 km. Wide retroarc foreland systems are developed in thicker, older or less deformed crust (high flexural rigidity or effective elastic thickness T_e) because the wavelength of the flexural profile depends primarily on the rheology and thickness of the underlying lithosphere (Beaumont, 1981; Watts, 1992; Catuneanu, 2004). Therefore, a wide system could be predictable for the South American case during the Late Cretaceous because this system was formed largely on thick and old crust with high T_e (Figs. 1 and 6), including the Proterozoic–Archean Amazonian Craton (Cordani and Teixeira, 2007), the Neoproterozoic Paraguay Belt (Godoy et al., 2010), the Paleoproterozoic Paranapanema Craton (Mantovani et al., 2005; Ramos et al., 2010), and the Paleoproterozoic Rio Apa Block (Godoy et al., 2009; Faleiros et al., 2015).

A cratonward movement of the system is likely a response to the orogenic front shifting and redistribution of the load. The simultaneous shift of the depocenters of Bauru and Andean basins during

times dominated by orogenic loading (Mochica and Peruvian phases) demonstrate that progradation of the South American retroarc foreland system occurred during its early stages in the Late Cretaceous, when the crust probably had a relatively higher rigidity. However, several authors have demonstrated that shifting occurred toward the craton during its later stages, as represented by the Chaco foreland system extending to the modern Pantanal Basin (Horton and DeCelles, 1997; DeCelles and Horton, 2003; McQuarrie et al., 2005). Flexural models predict that after successive loading events, the lithosphere relaxes stress, and with lower flexural rigidity, the system tends to be relatively smaller (Beaumont, 1981; Beaumont et al., 1988; Watts, 1992; Catuneanu, 2004). This condition explains the relatively shorter wavelength of the system observed in Cenozoic (Fig. 11), and a more significant progradation of the orogenic front occurred during this time of the Andean evolution.

The maximum preserved thickness (271 m) of the Bauru sedimentary fill is an order of magnitude greater than what would be expected if accommodation resulted only from flexural subsidence. Flexural models predict a subsidence on the order of 10 m for reasonably rigid continental lithosphere in back-bulge depozones (Beaumont, 1981; Allen and Allen, 2005; DeCelles and Giles, 1996). Nevertheless, several authors have documented accumulations of sediment in back-bulge depozones that range in thickness from tens of meters to more than 600 m (e.g. Plint et al., 1993; DeCelles and Currie, 1996; Holt and Stern, 1994; DeCelles and Giles, 1996; Leier et al., 2007). Thick accumulation have been explained by superimposed mechanisms, such as differential uplift and subsidence of basement blocks (Plint et al., 1993; Catuneanu et al., 1997a, b), dynamic subsidence and subcrustal loading (Holt and Stern, 1994). A key question is which set of variables controlled the widespread accommodation and thick sediment accumulation in the Bauru Basin.

First, mini-basins within the back-bulge province as well as the foreland system as a whole might have potentially been generated by differential subsidence and uplift associated with zones of weakness in the underlying crust, similar to the Western Canada Basin (Plint et al., 1993) and the Karoo foreland system (Catuneanu et al., 1997a, b). Therefore, in theory, the reactivation of crustal faults by the Cretaceous geodynamics could have been a mechanism overlapping the flexural subsidence of the Bauru Basin and might also have controlled its thickness.

Subcrustal loading processes can also act in the retroarc foreland system as a minor subsidence mechanism (Holt and Stern, 1994; Ingersoll, 2011; Allen et al., 2015). Thus, the subcrustal loading and cooling of underplating material suggested by Mariani et al. (2013) could be taken into account as an addition control on the Bauru accommodation. However, as discussed previously, the addition of mafic magma to the lower crust and uppermost mantle associated with the LIP is still uncertain, and the time span between this process and the beginning of the sedimentation should be better investigated.

Apparently, the forebulge was elevated above the base level during the Cenomanian to Campanian because there is no sediment with this age preserved in the west of Bolivia and northwest of Paraguay (Figs. 7 and 11). In the Maastrichtian, the dynamic subsidence might have been sufficiently high to outpace the flexural uplift of the forebulge and led to the preservation of fluvial-deltaic deposits in the forebulge province, the Cajones Formation in Bolivia and the Adrián Jara in Paraguay (Fig. 7). In the Santa Cruz region, a nonconformity separates the Jurassic extensional deposits below from the foreland deposits of the Cajones Formation above, and a northwest marine influence is recorded in the Cajones Formation (Kusiak and Zubieta Rossetti, 2003). However, the northwest marine incursions did not flood the entire foreland system because

there is no evidence of marine paleodepositional environments in the back-bulge province. In fact, the rate of dynamic subsidence decreases exponentially from the subduction zone to the back-bulge province (Fig. 6), tilting the entire flexural profile toward the orogen (Catuneanu, 2004; DeCelles et al., 2011), which explains the extent of the marine incursions only in the areas undergoing higher subsidence and preservation of the forebulge deposits in an area closer to the foredeep province. The active role of dynamic subsidence in the late Cretaceous is consistent with interpretations of shallowing of the subduction angle of the Farallon Plate at this time (i.e. Jaillard and Soler, 1996; Jaillard et al., 2000; Ramos, 2009; Folguera et al., 2011; Spagnuolo et al., 2012) because certain studies demonstrated that the dynamic loading by viscous mantle corner flow coupled to the subducting slab is more expressive if there is a shallow dipping slab as a result increasing dynamic subsidence (Mitrovica et al., 1989; Gurnis, 1992; Holt and Stern, 1994; Catuneanu et al., 1997a). Therefore, evidence exists of the action of dynamic subsidence in this foreland system, which also might have influenced the accommodation in Bauru Basin.

The South American plate shifted westward during the opening of the Atlantic Ocean (Figs. 8–10). The plate shows horizontal stress orientations nearly parallel to the direction of the absolute plate motion, indicating that the forces related to its motion are also responsible for the regional stress orientations (Allen and Allen, 1990; Assumpção, 1992). According to Cloetingh (1992), plates deform as they change latitude, and temporal fluctuations in stress are a natural consequence of the horizontal motions of the plates. It has been demonstrated that horizontal (in-plane) stress modifies the effects of existing vertical stresses on sedimentary basins, enlarging or reducing the amplitude of the resulting flexural deformation and simultaneously affecting several basins within a plate (Cloetingh et al., 1985; Miall, 2000). Indeed, in-plane stress is a minor mechanism in retroarc foreland basins (Allen and Allen, 1990; Ingersoll, 2011; Allen et al., 2015), and its superposition might amplify the primary driving mechanism; therefore, it might potentially act as another variable controlling the thickness of the Bauru deposits. This hypothesis is consistent with the idea addressed by Chang and Kowsmann (1996), who proposed that the region occupied by the Bauru Basin and other Cretaceous deposits (the Codó Formation in the Paraíba Basin and the Urucua Group in the São Francisco Basin) were downwarped by the action of in-plane stress associated with the change in stress regime in South America.

5. Conclusions

The Bauru Basin developed from the Cenomanian to early Paleocene in the back-bulge province of a retroarc foreland system that developed in response to Andean orogenic events. The characteristics discussed demonstrated that a supracrustal load was the main process acting in the Bauru Basin, but other subsidence mechanisms might have overlapped and accounted for its thickness.

The development of this foreland system should have influenced the evolution and geographic distribution of the coeval biota. Close similarities are noted among the paleobiota of the Bauru, Parecis and Potosí basins. Thus, the geodynamic events discussed in this contribution can be applied in future paleobiogeographic studies involving cycles of vicariance and geo-dispersal.

A differential orogenic loading along the strike might have generated a tilt in the direction of increased loading on the northwest of the system, which could have controlled the direction of the shoreline transgressions in the foredeep basin from north-western South America and also might have controlled the differential thicknesses of the Andean Basin deposits along the strike.

The recognition of this Upper Cretaceous back-bulge deposit as a component of a retroarc foreland system in South America provides additional evidence that the Andean orogeny began in the Late Cretaceous. Moreover, the migration of the Bauru and the Andean basins can be viewed as an indicator of thrust-belt activity. This case study demonstrates that the increase in the rate of dynamic loading with time might be associated with the decrease of the subduction angle below the Andean margin, and the interaction between the supracrustal load and dynamic subsidence led to the lowering of the forebulge below the base level, generating accommodation at continental scales.

Acknowledgments

Financial support for this research was provided by Petrobras (Petróleo Brasileiro S.A.), CNPq (Science without Borders Program, process 242946/2013-5) and LEBAC (Laboratório de Estudo de Bacias, Universidade Estadual Paulista). Additional support was provided by the University of Alberta. We thanks DAEE (Departamento de Águas e Energia Elétrica of the São Paulo State Government) in the person of José Luiz Galvão de Mendonça for providing most of the well-log information employed in this research. Thanks to Rodrigo Prudente de Melo for the construction of the database and support to draw the isopach maps. We are grateful to Caio Fabrício Cezar Geroto for discussions on the record of Crocodylomorpha. In addition, we owe a debt of gratitude to Jefferson Cassu Manzano and Liliane Maia T. Manzano who has helped immensely in fieldwork, and Aparecido Andrade Buchner who immeasurably helped preparing the core samples.

Appendix A. Supplementary data

Supplementary data related to this article can be found at <http://dx.doi.org/10.1016/j.marpetgeo.2016.02.027>.

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