The role of tectonics and climate in the late Quaternary evolution of a northern Amazonian River

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ABSTRACT

The Amazon basin has most of the largest rivers of the world. However, works focusing the geological evolution of the trunk river or its tributaries have been only partly approached. The Branco River constitutes one of the main northern Amazonian tributaries. A previous work proposed that, before flowing southward into the Negro-Amazon Rivers, the Branco River had a southwest to northeast course into the Caribbean Sea. The present work aimed to establish if the proposed change in the course of this river is supported by morphological and sedimentological data. Other goals were to discuss the factors influencing river development and establish its evolution over time within the chronological framework provided by radiocarbon and optically stimulated luminescence dating. The work considered the entire course of the Branco River downstream of the Precambrian Guiana Shield, where the river presumably did not exist in ancient times. The river valley is incised into fluvial sedimentary units displaying ages between 100 and 250 ky old, which record active and abandoned channels, crevasse splay/levees, and point bars. The sedimentary deposits in the valley include two alluvial plain units as old as 18.7 ky and which intersects a Late Pleistocene residual megafan. These characteristics suggest that a long segment of the Branco River was established only a few thousand years ago. Together with several structural anomalies, these data are consistent with a mega-capture at the middle reach of this river due to tectonic reactivation in the Late Pleistocene. This integrated approach can be applied to other Amazonian tributaries to unravel how and when the Amazonian drainage basin became established.

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1. Introduction

Landforms and sediments formed by large rivers play an important role in understanding the Earth's dynamics over the Cenozoic. The majority of the existing large river systems is located in South America, particularly in the Amazon basin (Latrubesse et al., 2005). With a length of approximately 6992 km (INPE, 2010) and a discharge of 175,000 m$^3$ s$^{-1}$ (Wohl, 2007), the Amazon River is a transcontinental drainage with crucial contribution for the maintenance of an ecosystem that holds one of the highest levels of biodiversity on Earth. Despite this relevance, several aspects of this river system remain to be investigated. For instance, the geological history of the Amazon River and its tributaries is an issue open for debate. Previous publications approaching this theme have chiefly aimed at discussing hypotheses of the time when the Amazon River shifted its flow from westward into the Pacific to eastward into the Atlantic Ocean (e.g., Potter, 1978; Figueiredo et al., 2009; Shephard et al., 2010; Sacek, 2014). Earlier works provided only a preliminary discussion on the geological history of the Madeira and Negro trunk rivers (e.g., Latrubesse and Franzinelli, 1998; Latrubesse and Franzinelli, 2005; Latrubesse, 2003; Rigsby et al., 2009; Plotzki et al., 2013, 2015; Rossetti et al., 2014a).

The Branco River is the main tributary of the Negro River. An existing hypothesis claimed that before its establishment as a north-to-southflowing river over the northern Amazonian landscape, the Branco River comprised an ancient drainage basin called Proto-Berbice (Guerra, 1956; Crawford et al., 1985; Gibbs and Barron, 1993; Schaefer and Dalrymple, 1996). According to these authors, this basin drained from southwest to northeast into the Caribbean Sea during the Late Neogene or Early Quaternary. Since then, erosion of highland basement rocks of the Guiana Shield has beveled the terrain, allowing the connection between the Tacutu and Solimões Basins located to the north and south, respectively (Fig. 1). This process would have promoted the reorganization of the Proto-Berbice drainage basin and the southward reversal of its main flow in order to discharge into the Negro-Amazon...
River basins (Schaefer and Dalrymple, 1996). Although such drainage reconstruction is supported by the presence of common fish species on both the Essequibo and Branco River drainage basins (Lujan, 2008; Lujan and Armbruster, 2011), there is an overall lack of geological and geomorphological data to test this hypothesis.

Another issue of interest for investigation is the temporal relationship between the Branco River and residual megafan deposits previously mapped in the region (e.g., Rossetti et al., 2012a, 2014b). Such deposits dominate the landscape of the Negro-Branco River basins, being associated with the most expressive occurrences of open vegetation that are anomalously intermingled with the rainforest, a phytophysiognomy long under debate (e.g., Takeuchi, 1960; Anderson, 1981; Furley et al., 1992; Pessenda et al., 2001; Sanaiotti et al., 2002; Cochrane and Cochrane, 2010; Rossetti et al., 2012b). The megafan deposits also constitute the largest Amazonian wetlands, which are testimony of a distributary paleodrainage that differs significantly from modern tributary systems that form the Amazon basin. Thus, it is expected that the establishment of the temporal relationship between the Branco River and the megafan deposits may bring about new elements for determining when this river developed its course southward into the Negro River.

This work aims to test if the proposed change in the course of the Branco River is supported by geomorphological and sedimentological data. This investigation, carried out within a geochronological framework provided by radiocarbon and optically stimulated luminescence (OSL) dating, considers the entire course of the Branco River downstream of the Precambrian Guiana Shield, where this river presumably did not exist until at least the early Quaternary.

2. Physisography and geological context

The Branco River, which accounts for about one third of the Negro River basin, drains an area of approximately 193,700 km² and has a mean annual discharge at the Caracaraí station (Fig. 1) of 2875 m³ s⁻¹. This area is dominated by a tropical climate (Aw in Köppen’s classification), with an average annual temperature of 24 °C and average rainfall accumulation of 1500 mm year⁻¹. The dry season is well defined, being concentrated between October and March. Rainy seasons occur between May and July, when 55–66% of the annual rainfall is received (Radambrasil, 1975). The prevalent vegetation on non-flooded areas consists of dense rainforest, whereas seasonally-flooded areas display white-sand vegetation, locally known as campinarana (Anderson, 1981; Cordeiro and Rossetti, 2015).

The Branco River is formed by the confluence of the Uraricoera and Tacutu Rivers, both located to the north of the city of Boa Vista (Fig. 1) and extending southward into the Negro River. Along its course, the Branco River can be described in terms of three segments (Fig. 1): an upper channel along the Tacutu Basin from the confluence of the Uraricoera and Tacutu Rivers up to the Mucajáí River; a middle channel that cuts down into Precambrian rocks of the Guiana Shield and forms several rapids downstream up to the city of Caracaraí; and a lower channel that flows into sedimentary rocks of the Solimões Basin and extends...
downstream up to the confluence with the Negro River. The Branco River varies downstream from dominantly straight to sinuous and has several local branching and characteristic related to anabranching sand-bed rivers (e.g. Nicholas, 2013). Its hydrological nature is annually bimodal, with changes from clear water having only bedload sediments in dry seasons to white water with high suspended-load in wet seasons.

Along its course, the Branco River basin drains Precambrian basement rocks of the Guiana Shield, as well as sedimentary deposits of the Tacutu and Solimões Basins. The northern part of the latter basin is also referred as the Pantanal Setentrional (Santos et al., 1993; Fig. 1), which corresponds to a topographically low-lying wetland of about 100,000 km² probably established in the Pliocene or Pleistocene (Santos et al., 1993; Rossetti et al., 2014b; Rossetti et al., 2016).

In surface, most of the Branco River basin was established on sedimentary deposits assigned to the Içá Formation (e.g., Bizzi et al., 2003). This unit, primarily defined by Maia et al. (1977), includes fluvial sandstones and mudstones of an estimated Plio-Pleistocene age. However, it has been suggested that at least part of these strata is Late Pleistocene-Holocene in age and includes aeolian sand dunes (e.g. Carneiro-Filho et al., 2002), as well as several triangular-shaped deposits related to inactive megafan depositional systems (Zani and Rossetti, 2012; Cremon et al., 2014; Rossetti et al., 2012a, 2014b).

3. Materials and methods

We mapped modern and ancient fluvial landforms using remote sensing. This procedure was integrated with sedimentological and chronological data, the latter applying radiocarbon and optically stimulated luminescence (OSL) dating of quartz grains.

Remote sensing data consisted of the Advanced Land Observing Satellite (ALOS-1) Phased Array type-L band Synthetic Aperture Radar (PALSAR) images, optical images with multispectral bands from Operational Land Imager (OLI)/Landsat-8 of a dry period, and a digital elevation model (DEM) obtained during the Shuttle Radar Topography Mission (SRTM). The ALOS-PALSAR data, distributed by the Japan Aerospace Exploration Agency (JAXA), consisted of orthorectified-flight beam dual (FDB) images in HH + HV polarizations. These images have pixel sizes of 12.5 × 12.5 m and were freely acquired from the Brazilian Institute of Geography and Statistics-IBGE (ftp://geoptf.ibge.gov.br/imagens/Alos/). ALOS-PALSAR images correspond to the wet period, which favoured distinguishing alluvial (flooded) areas from surrounding non-flooded (i.e., terra firme) areas. The multispectral OLI/Landsat-8 image (http://earthexplorer.usgs.gov) was processed in composite colors. This image helped the visual interpretation of open vegetation areas. The SRTM-DEM has a pixel size of 3 arc-seconds (~90 m), derived from the C-band interferometric InSAR sensor (Rabus et al., 2003; ftp://eosp01.uec.ac.jp/srtm/). These data were processed using customized shading schemes and palettes (e.g. Hayakawa et al., 2010). Such procedure was particularly useful for highlighting morphological features of interest from on-screen observations within the environment of the Global Mapper (Global Mapper LLC, 2009) software.

Geomorphological mapping using the ALOS-PALSAR, OLI/Landsat-8 and SRTM-DEM resulted in the identification of different depositional units at the mapping scale of 1:100,000. The resulting map provided the basis for establishing the strategy to approach the fluvial deposits in the field.

The sedimentological investigation was based on field campaigns, undertaken in the summers of 2013 and 2014. We analyzed riverbanks and cores acquired with a sampling hand auger. The sedimentary facies were described and photographed, with the data being recorded on measured lithostratigraphic profiles, with indication of lithology, texture, sedimentary structure and type of contact. Whenever possible, lateral variations in facies distribution along riverbanks were also noted.

Organic sediment samples were selected for ¹⁴C dating. The bulk organic matter was dated using an accelerator mass spectrometer (AMS) at the Beta Analytic Radiocarbon Dating Laboratory, Florida, USA. A standard pre-treatment with acid-alkali-acid washes was used to extract the organic matter. Conventional ¹⁴C ages were converted to calendar years before present (cal yrs BP) using the CALIB 7.0 software and these data were calibrated according to the INTCAL13 curve (Reimer et al., 2009).

The OSL dating was based on two independent measurements: equivalent dose and dose rate. Their ratio gives the OSL age. Samples for equivalent dose measurement were collected in PVC tubes, properly sealed to avoid exposure to light during sampling, storage and laboratory processing. Additional samples were collected in the same stratigraphic position for determining dose rates by high resolution gamma spectrometry. For measurements of equivalent doses of quartz, 180–250 μm grain sizes were separated by wet sieving. Afterwards, these fractions were treated with HCl (3.75%), H₂O₂ (29%) and HF (40%) for 50 min to dissolve carbonates, organic matter, feldspar and the outer rim of quartz grains damaged by alpha particles, respectively. Quartz grains were then isolated by gravity settling using a lithium polytungstate solution with densities of 2.85 g cm⁻³ and 2.62 g cm⁻³ in order to remove heavy minerals and remaining feldspar grains. OSL measurements for equivalent doses were carried out in aliquots with 100–200 grains of quartz. This was performed on a Riso OSL/Ti model DA-20 system equipped with blue LEDs, Hoya U-340 filters and built-in 39Ar/37Ar beta source (dose rate of 0.084 Gy s⁻¹) from the Luminescence and Gamma Spectrometry Laboratory (LEGaL) at the Institute of Geosciences of University of São Paulo (USP). The analyses followed the single aliquot regenerative-dose protocol (OSL-SAR) in quartz aliquots (Murray and Wintle, 2000; Wintle and Murray, 2006). The OSL-SAR protocol used for equivalent dose estimation (Table 1) was validated by a dose recovery test in quartz aliquots bleached under a solar simulator. The calculated-to-given dose ratio was 1.01 ± 0.12, for a given dose of 30, 200, 550, 1000 and 2000 s with preheat temperature of 200 °C (Supplementary Table 1). Only aliquots with recycling ratio in the range 1.0 ± 0.1, recuperation ~10% and negligible feldspar content (tested by infrared stimulation) were used for equivalent dose calculation, which was determined by the fitting of the dose-response curve using a linear-exponential function. This procedure was based on the central age model (CAM) and the minimum age model (MAM) (Galbraith et al., 1999) for samples with equivalent dose distributions presenting over-dispersion below or above 30%, respectively (Supplementary Fig. 1). Minimum equivalent doses estimated using the MAM for multigrain aliquots may be larger than equivalent doses of well bleached grains, because possible grains with poor bleaching within the aliquots cannot be discarded (Galbraith and Roberts, 2012). Dose rates were determined by the concentrations of U, Th, ⁴⁰K radionuclides, added to the cosmic radiation dose rate. A high-purity germanium gamma-ray detector (HPGe, relative efficiency of 55% and energy resolution of 2.1 KeV) was used to obtain U, Th and K concentrations. Samples were measured after sealing during 28 days in plastic containers in
order to reestablish the $^{222}\text{Rn}$ equilibrium. The cosmic dose was calculated considering location (latitude/longitude), depth below surface and altitude of the samples (Prescott and Stephan, 1982). The content of water can affect dose rate by absorbing ionizing radiation, thus it was also measured and included in the calculation. Dose rates were calculated using conversion factors provided by Adamiec and Aitken (1998).

4. Geomorphological characterization

Six morphological units were identified in the studied segment of the Branco River valley. These units include island, lower alluvial plain, upper alluvial plain, undifferentiated alluvial plain, sedimentary basement referred as the Içá Formation in published geological maps (e.g., Bizzi et al., 2003), and inactive megafan surface. Numerous sand bars also occur within this river, but they were not mapped in this work.

The HH and HV polarizations of the ALOS-PALSAR sensor had the best performance for identifying alluvial plains in the river valley (Fig. 2a), reflecting the double-bounce backscatter from the flooded forest (alluvial). The SRTM-DEM in color palette scheme (Fig. 2b) differentiated the lower and upper alluvial plains (Fig. 2c). Alluvial islands correspond to discontinuous landforms surrounded by water. The resulting map with geomorphological units is presented in Fig. 3. The distinction between the lower and upper alluvial plains was not possible along tributaries of the Branco River due to the resolution of the remote sensing data. Thus, these units were joined in the unit undifferentiated alluvial plain.

The lower Branco River has about 140 islands. These are generally elongated and parallel to the channel margins. The islands display areas ranging in planform from 1.7 to 19 km$^2$ (mean of 1.45 km$^2$). Topographically, they stand up to 3 m higher than the water level at low stages.

The lower alluvial plain is located ~4 m higher than the low water stage. This unit is distributed asymmetrically along the river. Between the city of Caracaraí and upstream of the confluence between the Branco and Anauá Rivers, the lower alluvial plain occurs only locally. However, this unit is expressive downstream from these rivers’s confluence, continuing in both margins of the Branco River. In the left margin of this river, the lower alluvial plain increases in width from 1 to ~3 km downstream, being nearly constant in the opposite river margin, i.e., between 3 and 4 km, although it locally reaches up to 7 km.

The upper alluvial plain is located 4–6 m above the low water level. This unit is continuous in both margins of the Branco River between the city of Caracaraí and the confluence with the Anauá River. Its width is only <2.5 km at this locality, but forms a long belt as wide as 7 km in length at the right margin of the Branco River between the Anauá and Água Boa do Univini Rivers, being absent downstream until the Itapará River. Southward of this river, there are five other discontinuous occurrences of this unit in both sides of the Branco River. The most expressive is 30 km in length and 8 km in width and occurs upstream of the Xeriuini River.

The undifferentiated alluvial plain refers to all sedimentary units <2.5 km in width located up to 5 m above the low water stages of main tributaries with discharge into the lower segment of the Branco River. These tributaries have meandering channels, contrasting with the anabranching pattern of the lower Branco River. Some of these tributaries, for example the Ajarani, Anauá and Catrimani Rivers, have confluence angles ranging from only 10° to nearly orthogonal.

The unit sedimentary basement, referred to as the Içá Formation in previous geological maps (i.e., Bizzi et al., 2003), corresponds to the

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Fig. 2. Maps of a segment of the Branco River valley with geomorphological characteristics. a) ALOS-PALSAR color composite image [R(HH), G(HV), B(HH)] for the wet season. Red line = valley boundary. b) SRTM-DEM showing elevation. Black lines = boundaries of units with contrasting heights. c) Corresponding geomorphological units.
Fig. 3. Geomorphological maps of the study area in the lower Branco River. a) to c) corresponds to the sections from upstream to downstream.
substrate where the Branco River valley was established (Fig. 3). This unit has an area of approximately 22,770 km² and stands 2 to 5 m above the upper alluvial plain, being protected from seasonal flooding. The drainage pattern over this area varies from trellis and rectangular to amorphous.

The unit inactive megafan surface corresponds to several discontinuous, seasonally flooded, fan-shaped paleo-landforms developed over the sedimentary basement, which was previously related to residual megafan deposits formed by a distributary channel network (Zani and Rossetti, 2012; Rossetti et al., 2014b). This geomorphological unit, covered by open vegetation (Rossetti et al., 2012a), is only partly present in the study area (Fig. 3). One inactive megafan surface is of particular interest to this study, which corresponds to the largest megafan located in the western margin of the Branco River, in the northern part of the study area (Fig. 4). This feature corresponds to a major triangular-shaped megafan that extends in the NE-SW direction, constituting an area up to 1648 km². The megafan, named here as the Caracará megafan, is currently abandoned in the modern landscape. The sediments that built this megafan were supplied by a feeder channel up to 5 km in length, which is sharply intercepted by the Branco River valley (Fig. 4). Although larger, the Caracará megafan displays a morphology and orientation similar to the better-studied Viruá megafan observed in the eastern margin of the Branco River (Zani and Rossetti, 2012; Rossetti et al., 2012a) (Fig. 4). These two inactive megafans, as well as all other megafans in northern Amazonia, have their axes toward the Precordillera basement rocks of the Guiana Shield and spread out radially into the center of the Pantanal Setentrional Basin (see Fig. 3 of Rossetti et al., 2014b).

5. Facies description and depositional environments

The geomorphological units of the Branco River valley are sedimentologically represented by nine lithofacies (Table 2), organized into four facies associations and related to the following depositional paleoenvironments: 1) active channel; 2) abandoned channel/floodplain basin; 3) crevasse splay/levee (Fig. 5); and 4) point bar. Although lithologies in these facies associations are described as sedimentary rocks, descriptions include also unconsolidated counterparts in the instance of the youngest alluvial deposits formed in association with the units island (Fig. 6) and lower alluvial plain (Fig. 7).

5.1. Facies association A (active channel)

This association consists of either single or amalgamated, fining-upward, or less commonly, blocky (e.g., profile RBN5 in Fig. 7 and RBN42 in Fig. 8) successions bounded by sharp, concave-up basal surfaces with erosional reliefs of 2 to 4 m. These surfaces are mantled by lags of quartz, iron-cemented mudstone and sandstone pebbles, which are also dispersed throughout the association. Association A contains the coarsest-grained lithofacies of the study area, including massive (facies Sm) or cross-bedded, fine- to coarse-grained sandstone (facies Sc) and, secondarily, massive (facies Sm) or intraformational (facies Gmi) conglomerates (Table 2). Cross sets of stratified-sandstones are of small- to medium-size (i.e., usually 0.1 to 0.3 m thick, only exceptionally reaching up to 1 m thick). Either tabular or trough cross sets are present. Noteworthy is that these strata are often organized into fining-upward successions and they consistently display unidirectional foresets with avalanche angles and rare internal low-angle (i.e., <15°) dipping reactivation surfaces. Plant remains (mainly branches and leaves) are locally dispersed in this association, particularly in the sandy facies. The maximum thickness of facies association A is 5 m.

Fining- and thinning-upward successions bounded at the bases by erosional surfaces with concave-up geometry were characteristics used to relate facies association A to confined flows within channels. Such features are in agreement with the sedimentary record of many other channels described elsewhere (e.g., McLaurin and Steel, 2007).

Lag deposits overlying basal surfaces, added to fining- and thinning-upward successions, are consistent with sedimentation during waning flows, as typical of channels. The cross-stratified sandstones attest channel fills formed by small- to medium-scale, 2D- or 3D-bedforms that migrated downstream at channel bottoms.

5.2. Facies association B (abandoned channel and floodplain basin)

This association is interbedded or, most often, grades downward into the association A, completing its fining-upward nature. Association B also sharply overlies association A or is interbedded with association C, being composed of the finest lithofacies of the study area. These lithofacies include, in order of decreasing importance, laminated mudstone (facies MI), massive pelite (facies Mm), heterolithic deposits (facies H) and locally peat (facies P). Lower in the sections, these deposits are dark gray-colored, but upward they usually grade into light gray, pale to whitish yellow colors, when there is also a transition from facies Mm and H to facies Mm. The latter is dominantly reddish to yellowish and indurated, and contains iron concretions and root marks. Peat deposits (facies P) occur lower in the sections together with facies Lm and form laminae or layers up to 0.4 m in thickness. This facies consists either of concentrations of pure plant debris or a mixture of plant debris with a muddy matrix. Log and leave fragments are also locally dispersed throughout facies Lm and, to a lesser extent, facies H and Mm. In the latter, the plant fragments may be highly degraded. Not rarely, association B can be as thick as 6 to ~8 m (e.g., profiles RBN44 and RBN8 in Fig. 7).

Facies association B is related to abandoned channel and floodplain environments due to its genetic association with active channel deposits. In particular, the upward gradation from channel deposits and the dominantly muddy lithologies attest that this association was formed by mud settling during low energy flows within abandoned channels. The occurrence of association B external to channel deposits records deposition of mud from suspension in flat-lying floodplains. These areas are suitable to subaerial exposure and soil development, processes suggested in facies association B by the indurated horizons with root marks and iron concretions (facies Mm). Abandoned channels and floodplain mudstones may contain stagnant areas with anoxic conditions, characteristic that favours the preservation of plant remains and also peat formation, as recorded in facies P.

5.3. Facies association C (crevasse splay/levee)

At the outcrop scale, this association, one of the most frequent in the study area, is typically characterized by discontinuous, lenticular beds that pinch out into association B or, less often, are cut down by association A. Facies association C is distinguished from the other deposits by several coarsening-upward packages usually <0.5 m thick that grade upward from association B, and which have upper sharp contacts marked by rootlets and undistinguishable invertebrate trace fossils. In many instances, the thickness of the individual cycles increases upward (e.g., profiles RBN17 and RBN82 in Fig. 6; RBN84A in Fig. 7), although the opposite pattern, i.e., cycles that are thinner-upward, is also present (e.g., profile RBN13 in Fig. 7). These cycles consist of moderately to well sorted, very fine- to fine-grained sandstones, with local medium-grained sandstones. These deposits are mostly massive (facies Sm) or locally cross-stratified (facies Sc; see profile RBN13 in Fig. 7). Peat layers <10 cm thick and with highly fragmented plant debris are locally present (e.g., profile RBN 29 in Fig. 7). Plant remains may also occur dispersed within this association.

The sandy nature with grain sizes finer than in association A, the lenticular nature, and the thickening-upward successions are main features that allowed the attribution of facies association C to crevasse spays. Many crevasse splay deposits of similar characteristics have been recognized in the sedimentary record (e.g., Makaske, 2001;
The fact that these sandy lenses grade upward from facies association B conforms to this proposed interpretation. Such superposition of facies association records increased flow stage, with the consequent episodic progradation of sand from channels into adjacent floodplains, forming suspension lobes.

5.4. Facies association D (point bar)

This association is subordinate in the study area, being recorded only locally in profiles RBN24 and RBN60 (Fig. 9). It forms packages up to ~1 m thick that either occur isolated within association B or are interbedded with association A. These deposits extend laterally only a few tens of meters along the outcrops. They consist of moderately to well sorted, fine- to medium-grained sandstone internally with large scale, low-angle (i.e., <12°) dipping, inclined beds (facies Si). The inclined sandstone beds form lateral packages up to 0.2 m thick, which are separated by reactivation surfaces marked by laminae or a few cm-thick layers of mudstone. As recorded in the other facies associations, these deposits may also contain dispersed plant debris.

Facies association D was attributed to point bar deposits mostly based on the large-scale, low angle-dipping inclined beds, related to lateral deposition in inner meander loops during channel migration (e.g., Nichols, 1999). Reactivation surfaces within these deposits indicate alternation between episodes of sand deposition and non-deposition and or erosion of previously deposited sediments. Such characteristics are compatible with changes in flow energy due to seasonal fluctuations (Bridge and Demicco, 2008; Miall, 2014).

6. Facies content and chronology of geomorphological units

As described in the previous section, the studied deposits, including the ones representative of the sedimentary substratum over which the Branco River valley became established, are related to environments typical of fluvial systems. The distribution of these deposits within the individual geomorphological units is described in the present section, together with their chronological context based on 23 OSL ages (Table 3) and 22 radiocarbon ages (Table 4). These data provide the basis to reconstruct the fluvial dynamics before the establishment of the modern landscape, as well as discuss its main controlling factor.

6.1. Alluvial island

Alluvial islands in the study area contain facies associations A to C. These deposits are typically friable in the islands, lacking any evidence of iron cementation or concretions. Dominant deposits are related to abandoned channels/floodplains (association B) and crevasse splays (association C). These deposits may compose the entire section, as in profiles RBN17 and RBN82 (Fig. 6). Active channel deposits (association A) are subordinate, occurring only as a thin package in profile RBN21 (Fig. 6). These deposits locally occur in proportion comparable to the other deposits, as in profile RBN7 (Fig. 6), where they form interbeddings. In addition, the active channel deposits dominate locally, as in profile RBN10 (Fig. 6). Six 14C ages were obtained from the muddy facies of deposits representative of alluvial islands, which vary from 2779 to 2929 to 1352–1446 cal yrs BP, with the other ages being only a few hundred years old (Table 4 and Fig. 6).

6.2. Alluvial plain

The lower alluvial plain consists mainly of alternations between facies associations A and B, with the former being volumetrically much more expressive than in the alluvial islands. In fact, alluvial plains is recorded by the amalgamation of association A (see profiles RBN42 and
Intraformational massive conglomerate (Gmi): Conglomerate consisting exclusively of mudstone pebbles within a matrix of moderately sorted, fine- to medium-grained sandstone. Massive bedding. Local reworking of mud layers during episodes of high energy flows, without time to generate stratification.

Low angle inclined sandstone (Si): Moderately to well sorted, fine- to medium-grained sandstone internally with large-scale (i.e., metric), low angle-dipping bedding. Deposition by lateral accretion forming bars.

Cross-bedded sandstone (Sc): Medium- to coarse-grained sandstone with planar or trough cross-stratification, in general, of small- to medium-scale (sets thickness = 10 to 20 cm). Deposition due to migration of small- to medium-scale, either 2D or 3D mega-ripples under unidirectional, low flow regime.

Massive sandstone (Sm): Massive sandstone varying from very fine- to coarse-grained, generally moderately to well sorted. Locally highly bioturbated, with undistinguishable trace fossils. Sediment mixing due to weathering and/or pedogenesis/bioturbation.

Heterolithic deposits (H): Mudstone layers interbedded with very fine- to medium-grained sandstone forming lenticular, streaky and, less commonly, wavy beddings. Episodic sedimentation with alternation between slow mud deposition from suspension under low flow energy and frequent sand input during relatively higher flow energy.

Massive siltstone (Mm): Massive mudstone/siltstone. Locally endured, mottled and wavy beddings.

Laminated mudstone (Me): Laminated mudstone with parallel horizontal lamination.

Peat (P): Mudstone or sandstone with high amount of organic plant remains, locally highly fragmented as “coffee ground”. Slow horizontal deposition of mud from suspensions under low flow energy. Accumulation of plant remains under anoxic conditions.

<table>
<thead>
<tr>
<th>Lithofacies (code)</th>
<th>Description</th>
<th>Sedimentary process</th>
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<tbody>
<tr>
<td>Massive conglomerate (Gm)</td>
<td>Massive conglomerate of pebbles supported by poorly sorted, fine to coarse-grained sandy matrix. Pebbles chiefly of quartz and subordinately ferruginous concretions.</td>
<td>High energy flows transporting sediments as bedloads. The lack of structure may be due to weathering and/or rapid sedimentation, with no time for sediment organization.</td>
</tr>
<tr>
<td>Intraformational massive conglomerate (Gmi)</td>
<td>Conglomerate consisting exclusively of mudstone pebbles within a matrix of moderately sorted, fine- to medium-grained sandstone. Massive bedding.</td>
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<tr>
<td>Massive sandstone (Sm)</td>
<td>Massive sandstone varying from very fine- to coarse-grained, generally moderately to well sorted. Locally highly bioturbated, with undistinguishable trace fossils.</td>
<td>Sediment mixing due to weathering and/or pedogenesis/bioturbation.</td>
</tr>
<tr>
<td>Heterolithic deposits (H)</td>
<td>Mudstone layers interbedded with very fine- to medium-grained sandstone forming lenticular, streaky and, less commonly, wavy beddings.</td>
<td>Episodic sedimentation with alternation between slow mud deposition from suspension under low flow energy and frequent sand input during relatively higher flow energy.</td>
</tr>
<tr>
<td>Massive siltstone (Mm)</td>
<td>Massive mudstone/siltstone. Locally endured, mottled and wavy beddings.</td>
<td>Slow deposition of mudstone and siltstone from suspension under very low flow energy with subsequent subaerial exposure and pedogenesis.</td>
</tr>
<tr>
<td>Laminated mudstone (Me)</td>
<td>Laminated mudstone with parallel horizontal lamination.</td>
<td>Slow horizontal deposition of mud from suspensions under low flow energy.</td>
</tr>
<tr>
<td>Peat (P)</td>
<td>Mudstone or sandstone with high amount of organic plant remains, locally highly fragmented as “coffee ground”.</td>
<td>Accumulation of plant remains under anoxic conditions.</td>
</tr>
</tbody>
</table>

In some instances, association B is interbedded only with association C, forming deposits up to 12 m thick, as in profile RBN29 (Fig. 7). Deposits derived from this unit recorded seven ¹⁴C ages and two OSL ages. The radiocarbon ages are as young as 111.0 pMC (percent modern carbon), i.e., a few decades, at the top of the profiles, being within the same range as the ages obtained for the alluvial islands. However, older radiocarbon ages between 4224 and 4410 and 4956–5075 cal yrs BP and the OSL ages of 3557 (±322) and 4100 (±331) yrs were also recorded. Interestingly, these older ages are not randomly distributed, but they increase systematically downward in the lithostratigraphic profile. Such age distribution led to discard contamination by older organic matter, thus the alluvial plain deposits were most likely formed since the mid-Holocene. Profiles from the upper alluvial plain may be entirely composed of abandoned channel/floodplain deposits, as recorded in profiles RBN4 and RBN6, as well as most of RBN5, RBN8, RBN13 and RBN35 (Fig. 7). The four latter profiles display sandy intervals at the bases, which are related to active channel (RBN5 and RBN35) and crevasse splay deposits (RBN8 and RBN13). In addition, this unit is also represented by profiles with equal proportions of active channel and abandoned channel/floodplain strata (RBN16) or it may be dominated by crevasse splay deposits (RBN84). Radiocarbon ages from the upper alluvial plain unit range from 6825 to 6981 to 7577–7678 cal yrs BP. OSL ages varied from 10,426 (±598) to 18,746 (±1583) yrs. As in the lower alluvial plain, these ages increase systematically downward in the lithostratigraphic profiles. It follows that the formation of the upper alluvial plain deposits initiated in the Late Pleistocene and continued to form until the early to mid-Holocene.

Undifferentiated alluvial plains of the Branco River tributaries were locally recorded at three locations, represented by profiles RBN 26, RBN35, RBN42, RBN61 and RBN67 (Fig. 8). These contain only facies associations representative of active channels (association A) that either grade upward into abandoned channel/floodplain deposits (association B) (RBN26, RBN35 and RBN67) or are entirely represented by active channel deposits (RBN61). Radiocarbon ages are as young as 1302–1396 cal yrs BP, with OSL ages being as young as 817 (±68) yrs. However, one radiocarbon age obtained in a bank of the Água Boa do Uninivi River, close to its confluence to the Branco River, recorded 5268–5323 cal yrs BP. In addition, a much older OSL age, i.e., 42,850 (±3277) yrs, was recorded at the base of the profile RBN35, upstream in this same tributary.

6.3. Sedimentary basement and inactive megafan surface

The sedimentary basement consists chiefly of facies associations related to active channel and abandoned channel/floodplain environments. Active channel deposits are generally coarse-grained and grade upward into the latter, forming fining-upward successions. However, these associations are frequently in sharp contact with each other (e.g., profiles RBN 100, RBN101, and RBN103 in Fig. 9), and eventually contains only active channel deposits (e.g., RBN107). The uppermost portions of most profiles are dominated by abandoned channel/floodplain deposits. As opposed to the previous units, crevasse splay deposits were present only in one profile (RBN23) of the Içá Formation. Point bar deposits, not documented in any other units, were recorded in profiles RBN24 and RBN60. An important characteristic of the sedimentary basement is the discontinuity surface in its top, marked by a paleosol up to 5 m thick. This paleosol is characterized by root marks, representing an indurated interval with in situ concretions of iron oxides/hydroxides. OSL dating of five samples derived from these deposits reported ages distinctively much older than the ones recorded in any other depositional units, with values ranging from 102,054 ± 9054 to 245,766 ± 23,816 yrs. These samples had equivalent doses near the reliable maximum doses (2Dq) possible to be estimated for quartz grains using blue stimulation OSL signal.

Only the southernmost edge of the Caracaraí megafan deposits were accessed in this study, which corresponds to two localities upstream in the Água Boa do Uninivi River. Both profiles recorded loose and well-sorted sands at their tops. In profile RBN55, these deposits unconformably overlie the facies associations related to active channels and abandoned channels/floodplains. The OSL age of 22,241 ± 1659 yrs and 2928 ± 232 yrs were obtained from deposits below and above the unconformity, respectively (Fig. 9). In the profile RBN57, there are only moderately to well-sorted sandstones, with the lower succession displaying a fining-upward trend related to active channel deposits. Two OSL ages from this locality recorded 6262 ± 410 and 4676 ± 371 yrs.

Table 2

<table>
<thead>
<tr>
<th>Lithofacies (code)</th>
<th>Description</th>
<th>Sedimentary process</th>
</tr>
</thead>
<tbody>
<tr>
<td>Massive conglomerate (Gm)</td>
<td>Massive conglomerate of pebbles supported by poorly sorted, fine to coarse-grained sandy matrix. Pebbles chiefly of quartz and subordinately ferruginous concretions.</td>
<td>High energy flows transporting sediments as bedloads. The lack of structure may be due to weathering and/or rapid sedimentation, with no time for sediment organization.</td>
</tr>
<tr>
<td>Intraformational massive conglomerate (Gmi)</td>
<td>Conglomerate consisting exclusively of mudstone pebbles within a matrix of moderately sorted, fine- to medium-grained sandstone. Massive bedding.</td>
<td>Local reworking of mud layers during episodes of high energy flows, without time to generate stratification.</td>
</tr>
<tr>
<td>Low angle inclined sandstone (Si)</td>
<td>Moderately to well sorted, fine- to medium-grained sandstone internally with large-scale (i.e., metric), low angle-dipping bedding.</td>
<td>Deposition by lateral accretion forming bars.</td>
</tr>
<tr>
<td>Cross-bedded sandstone (Sc)</td>
<td>Medium- to coarse-grained sandstone with planar or trough cross-stratification, in general, of small- to medium-scale (sets thickness = 10 to 20 cm).</td>
<td>Deposition due to migration of small- to medium-scale, either 2D or 3D mega-ripples under unidirectional, low flow regime.</td>
</tr>
<tr>
<td>Massive sandstone (Sm)</td>
<td>Massive sandstone varying from very fine- to coarse-grained, generally moderately to well sorted. Locally highly bioturbated, with undistinguishable trace fossils.</td>
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</tr>
<tr>
<td>Massive siltstone (Mm)</td>
<td>Massive mudstone/siltstone. Locally endured, mottled and with incipient reddish or yellowish iron concretions, as well as root marks.</td>
<td>Slow deposition of mudstone and siltstone from suspension under very low flow energy with subsequent subaerial exposure and pedogenesis.</td>
</tr>
<tr>
<td>Laminated mudstone (Me)</td>
<td>Laminated mudstone with parallel horizontal lamination.</td>
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<td>Peat (P)</td>
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<td>Accumulation of plant remains under anoxic conditions.</td>
</tr>
</tbody>
</table>
7. Evolution of the depositional system

The analysis combining sedimentological, geomorphological and chronological data indicates that in the Mid-Late Pleistocene, i.e., long before the establishment of the Branco River, the study area was dominated by fluvial deposits of the Içá Formation. Although this unit contains facies associations similar to the ones representative of the evolution of the Branco River, it appears to record a fluvial system with more energetic flows, as suggested by the prevalence of coarse-grained (i.e., sandy and conglomeratic) channel deposits. Previous publications suggested that coarse-grained deposits of the Içá Formation from other areas of the Amazonian lowlands represent braided river systems (e.g., Maia et al., 1977). However, the association of coarse-grained channel deposits with point bars, abandoned channels and crevasse splay strata sustains meandering channels. Many meandering channels have been reconstructed from the sedimentary record based on similar deposits (e.g., Miall, 1996; Willis and Tang, 2010; Labrecque et al., 2011). Deposits with faciological characteristics comparable to the Içá Formation from central Amazonia were also attributed to meandering fluvial systems (Nogueira et al., 2013; Rossetti et al., 2015).

The fact that the sedimentary basement related to the Içá Formation is topped by a discontinuity surface with a paleosol with in situ concretions of iron oxides/hydroxides is compatible with a temporal hiatus with non-deposition and/or erosion between these fluvial deposits. The characteristics of this ferruginous paleosol are similar to those from other lateritic paleosols recorded throughout Amazonia, which have been related to a climate with alternating and well-defined wet and dry seasons (e.g., Rossetti, 2004, 2006). Thus, the discontinuity surface at the top of the sedimentary basement records a significant time gap from ~100,000 to ~42,000 yrs, formed under a climate with alternating wet and dry seasons. Only after this climatic period is that the accumulation of alluvial deposits was initiated along the Branco River valley.

Following the establishment of the unconformity at the top of the sedimentary basement, the northern Amazonia experienced a drastic change in drainage style, with the onset of distributary networks that favoured the development of several megafans (Rossetti et al., 2012a, 2014b). These are recorded by triangular-shaped landforms, represented in the study area by the Viruá and Caracará megafan deposits (Fig. 4). The fact that all megafans of northern Amazonia have axes pointing to Precambrian basement rocks of the Guiana Shield and are distributed radially toward the center of the Pantanal Setentrional Basin is consistent with a common origin, previously related to the tectonic evolution of this basin (Rossetti et al., 2014b). The overall northeast/southeast orientation of the Viruá and Caracará megafans and of their main feeder channels (Fig. 4) indicates sediments sourced chiefly from northeast to southwest. Although the feeder paleochannel of the Caracará megafan is sharply interrupted by the Branco River alluvial valley, the morphological analysis suggests its northeastward continuity. This channel would have supplied the Caracará megafan with sediments sourced from northeastward, i.e., from the Guiana Shield. Although the distributary channels in a megafan typically change direction, the overall orientation of both the feeder paleochannel of the
Caracaráí megafan and the megafan suggests a depositional system with a roughly northeast- to southwest-trending distributary drainage before the establishment of the Branco River. The river onset would have interrupted the sediment supply into the Caracaráí megafan and eventually deactivated the megafan as a significant depositional site. When the Branco River cut down into the sedimentary basement to produce its currently extensive southward-flowing valley is an issue that can be discussed with basis on the chronology of the Caracaráí megafan. The ages available for the inactive megafan surface are only from terminal deposits, which may include sediment reworked from small local drainages, thus being unrelated to the distributary network responsible for the megafan development. Nonetheless, a plausible hypothesis is that the Branco River was formed from the expansion of pre-existing, smaller-scale drainage basins.

The age obtained for the undifferentiated unit along of the Água Boa do Univini River suggests sediment deposition as old as 42,850 ± 3277 yrs. Understanding the origin of these sedimentary deposits is difficult, because this age is much older than the ages of the alluvial deposits in the Branco River valley. This age is also much younger than the ages of the sedimentary basement recorded in the study area and of correlatable deposits previously documented in other Amazonian areas (e.g., Rossetti et al., 2015). Our hypothesis is that at 42,850 ± 3277 yrs the studied segment of the river did not undergo active deposition. Instead, a proto-drainage network in the lower Branco River was disconnected from upstream tributaries of this river, such as the Uraricoera, Tacutu and Mucajai Rivers. These rivers might have deviated their courses southward to connect with less extensive drainage basins, forming the Branco River basin with its long southward-flowing valley.

The present data indicate that, after downcutting, there was a first phase of sediment aggradation within the Branco River valley from at least 18.7 ky to 7592–7754 cal yrs BP, represented by the upper alluvial plain unit. Deposits within this unit indicate that the Branco River was probably a mixed-load anastomosed river during this time-frame, as suggested by sandy and muddy facies related to abandoned channels and floodplains, as well as crevasse splays interbedded with active channel sandstones, as also observed elsewhere (e.g., Miall, 2014).

Sediment aggradation was proceeded by a renewed phase of valley downcutting, which left behind the upper alluvial unit in a position distant from the river course. During this time, a significant volume of these deposits was eroded, particularly in the central and lower parts of the study area, where the upper alluvial plain is either absent or discontinuous (Fig. 3). There was no substantial change in the river dynamics, with the river remained as a mixed-load system. This is evidenced by the mixture of muddy and sandy lithologies formed within active channel, abandoned channel/floodplain and crevasse splay environments. The ages obtained for these deposits attest river downcutting sometime before 4956–5075 cal yrs BP, when sediment aggradation initiated in the new space created along the river valley. The fact that deposits as young as 111.0 pMC are recorded along the profiles of the lower alluvial plain indicates that this area is still a site of active sedimentation. The muddy and sandy deposits with such young age suggest no change in sediment load over time (e.g., Sander et al., 2014, 2015). Indeed, the change from a clear-water river in the dry period into a white-water river in the wet period attests that this river is still seasonally prone to sand and mud deposition. However, the river morphology changed from anastomosing (type 3 anabranching river according to Nanson and Knighton, 1996) to sinuous with long rectilinear segments. This is currently an anabranching sand-bed river, with several local avulsions and lacking crevasse splays, but such environments are well represented in the sedimentary record of the upper and lower
Fig. 7. Lithostratigraphic profiles and chronology of the lower (a) and upper (b) alluvial plains. The legend is the same as in Fig. 6. Profile locations are in Fig. 3.
alluvial plains, as well as alluvial islands. Considering this fact, and also the age similar to the lower alluvial plain deposits, we can conclude that some alluvial islands correspond to fragments detached from the lower alluvial plain during river diversion, and some are emerging from the stabilization of mid channel sandy bars. The proposed change in fluvial style perhaps occurred only very recently, i.e., over the last centuries, as suggested by the sedimentary record of the upper and lower alluvial plains and alluvial islands, which contain very recent crevasse splay and floodplain deposits.

8. Control on fluvial evolution over time

The two main factors with higher impact on the development of drainage basins are climate and tectonics (e.g., Latrubesse and Rancy, 1998; Souza Filho and Stevaux, 2004; Bertani et al., 2014). Which of these factors had the greatest effect on the paleoenvironmental changes of the Branco River basin is not a straightforward question. Therefore, both hypotheses must be addressed with caution.

Numerous publications have recorded variations from dry to wet and cold to warm climates during the Late Pleistocene-Holocene in several Amazonian areas (e.g., Van der Hammen and Hooghiemstra, 2000; Mayle and Power, 2008; Govin et al., 2014). A compilation of several Amazonian areas (e.g., Van der Hammen and Hooghiemstra, 2000; Mayle and Power, 2008; Govin et al., 2014). Which of these factors had the greatest effect on the paleoenvironmental changes of the Branco River basin is not a straightforward question. Therefore, both hypotheses must be addressed with caution.

Noteworthy is that the continued long-term cooling that succeeded the short period of deposition of the undifferentiated unit recorded no further sedimentation. Instead, while the climate became dominantly drier during the Last Glacial, the Branco River valley forced its way through the landscape. This fluvial valley developed during a time unsuitable to increased stream power resultant from climatically-regulated erosion. Such incompatibility with the climatic signal reinforces the proposed interpretation that the studied segment of this river was shaped by river capture due to a tectonic cause.

The water table during the LGM is expected to have been at the lowest stage. According to the glacio-eustatic sea-level fluctuation model of terrace formation, this corresponds to a time with increased erosion and downcutting. Instead, the Branco River valley initiated its incision during the Last Glacial, the Branco River valley forced its way through the landscape. This fluvial valley developed during a time unsuitable to increased stream power resultant from climatically-regulated erosion. Such incompatibility with the climatic signal reinforces the proposed interpretation that the studied segment of this river was shaped by river capture due to a tectonic cause.

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The sedimentation of the undifferentiated alluvial plain along the Água Boa do Univivi River occurred during the early stages of the Last Glacial episode, when the climate was wet but more seasonal than the present (Fig. 10). This pattern is incompatible with the glacio-eustatic sea-level fluctuation model of terrace formation (e.g., Maddy, 1997) applied to explain the terrace formation along the Amazonian lowland rivers (Itron and Kalliola, 2010). This is because the model considers incision and terrace downcutting during glacial lower sea-level conditions and aggradation during warmer interglacial higher sea-level with inundation and burial of former terraces. Deposition on the undifferentiated alluvial plain could be explained from the climatically-modulated fluvial terrace development. In this model, alluviation is typically associated with the long-term peniglacial cooling, while erosion and terrace downcutting are related to higher stream power and higher sediment supply during the Last Glacial to early Holocene transition (Bridgland, 2000; Gerlach et al., 1997; Starkel, 2003; Bridgland and Westaway, 2008).

Noteworthy is that the continued long-term cooling that succeeded the short period of deposition of the undifferentiated unit recorded no further sedimentation. Instead, while the climate became dominantly drier during the Last Glacial, the Branco River valley forced its way through the landscape. This fluvial valley developed during a time unsuitable to increased stream power resultant from climatically-regulated erosion. Such incompatibility with the climatic signal reinforces the proposed interpretation that the studied segment of this river was shaped by river capture due to a tectonic cause.

The water table during the LGM is expected to have been at the lowest stage. According to the glacio-eustatic sea-level fluctuation model of terrace formation, this corresponds to a time with increased erosion and downcutting. Instead, the Branco River valley initiated its incision during the LGM through an aggradational phase that produced part of the sedimentary record represented by the upper alluvial plain. In addition, it is noteworthy that deposition of this unit continued during the following glacial/interglacial transition, when the climate might have graded from cold and dry to dominantly wet (cf. Cordeiro et al., 2011; D’Ápolito et al., 2013; see B and C in Fig. 10). Deposition prolonged through the subsequent phase of dry and/or seasonal climate that lasted until the late-mid Holocene. This deserves a note because all these climatic changes did not affect the terrace evolution, as indicated by the continuous sediment aggradation and formation of the upper alluvial plain. In particular, climate-regulated terraces would typically experience downcutting during the wet conditions of the Tardiglacial due to higher
stream power imposed by increased water and sediment inflows, not recorded in the study area. Thus, the history of terrace erosion and deposition along the Branco River is not fully explained neither applying the glacio-eustatic base level fluctuation model, nor the purely climatic model. An allochtonous factor related to tectonic processes seems more likely and also more consistent with the unpaired nature of the upper

Fig. 9. Lithostratigraphic profiles and chronology from sedimentary basement (a) and inactive megafan surface (b) deposits.
alluvial plain. In general, terraces on opposite sides of a river having material that resists equally to erosion tend to present similar heights. In the instance of the Branco River, although the sedimentary basement from both margins of the Branco River is similar, this river downcut unevenly forming unpaired terraces, more suggestive of a tectonic control. Many other terraces with similar characteristics have been related to tectonic influence (e.g. Ouchi, 1985; Bridge and Demicco, 2008).

An increasing number of publications have shown oversimplified aspects of the climatically-related model of fluvial terrace evolution and the importance of tectonic uplift and subsidence to produce complex successions of fluvial terraces worldwide (e.g., Lewin et al., 1995; Van Balen et al., 2000; Starkel, 2003; Gibbard and Lewin, 2009; Wegmann and Pazzaglia, 2009; Wang et al., 2010; Pazzaglia, 2013). Rossetti et al. (2014a) have also concluded that changes in tectonic subsidence rates played an important role on terrace evolution along the Madeira River, a southern Amazonian tributary. The proposed influence of tectonic events on the Amazonian terraces is compatible with the neotectonic history proposed in numerous publications, as reviewed by Rossetti (2014). Therefore, although a tectonic investigation for the studied area is beyond the scope of the present work, the tectonic influence on the Branco River evolution can be defended based on the following three indirect evidences. First, the Branco River valley cuts through an extensive area of the eastern Pantanal Setentrional. This constitutes the largest wetland of the Amazonian lowland, with numerous inactive megafans formed during the Late Pleistocene to Holocene (Zani and Rossetti, 2012; Rossetti et al., 2012a, 2014b; Cremon et al., 2014). As highlighted in these previous publications, the megafans of the Pantanal Setentrional are distributed radially toward the central part of a widespread tectonic depression separated from Precambrian rocks of the Guiana Shield by numerous NW- and NE-trending faults (Rossetti et al., 2014b; Rossetti et al., 2016).

Second, the Branco River is anomalously long and essentially rectilinear. The river encountered its way from lowland areas of the Tacutu Basin into the Pantanal Setentrional wetland, cutting through a long segment of basement rocks of the Guiana Shield 90 to 150 m higher than the mean elevation of these basins (60 m; Fig. 1). Because the Pantanal Setentrional basin corresponds to an area currently under tectonic subsidence (e.g., Rossetti et al., 2014b; Rossetti et al., 2016), the most likely is that this change in river course took place due to tectonics.

Similar drainage anomaly has been commonly associated with tectonic causes, as recorded in the Rhine (Preusser, 2008) and Yangtze (Burbank and Anderson, 2012) drainage basins.

Third, the truncation of the Caracaraí megafan by younger Late Pleistocene-Holocene fluvial deposits constitutes a major paleographic change in the region. The modification from a distributary into a tributary drainage network serves further explanation. Such change in drainage pattern in the Viruá megafan, located in the eastern margin of the Branco River, was related to the combination of monsoonal climate with mild tectonic activity, the latter being responsible for the creation of a shallow depression where these megafan deposits were accommodated (Rossetti et al., 2012a). These authors also stated that the abandonment of the Viruá megafan was due to the deviation of its feeder channel by fault capture. Further studies are needed in order to decipher if the Caracaraí megafan had a similar origin. However, its close proximity to the Viruá megafan, the similar shape and orientation, as well as the sediment derived from same source rocks to the northeast (Fig. 4), suggest a comparable genesis. In addition, the abandonment of the Caracaraí megafan seems to have been triggered by the establishment of the Branco River, as evidenced by the fact that this river crosses the feeder paleochannel of the Caracaraí megafan. If this interpretation is correct, the abandonment of this megafan was most likely caused by the interruption of sediment influx sourced from the Guiana Shield due to drainage capture by the Branco River (Fig. 11b).

Despite the perceptible tectonic influence in the evolution of the Branco River, the most likely is that this factor interacted with climate, particularly since the mid-Holocene (Fig. 11c). Slightly before 5 ky, there was a short period of downcutting in the upper alluvial plain under a dry, or at least, seasonal climate (Latrubesse and Franzinelli, 1998, 2005; D’Ápolito et al., 2013; see A and B in Fig. 10). Prolonged droughts due to the intensification of dry seasons might have promoted terrace downcutting, because under such conditions erosion is due to the lowering of the water table. A new phase of sediment aggradation resulted in the accumulation of the lower alluvial plain deposits. The sedimentation of this unit took place under the increased mid-Holocene humidity phase, recorded in northern Amazonia (Sifeddine et al., 2001; Barbossa et al., 2004; Rossetti et al., 2005; Toledo et al., 2010; D’Ápolito et al., 2013) and also worldwide, and when the global climate was nearly 40% wetter than the Younger Dryas (Maslin and
The morphological characterization based on remote sensing and sedimentological analysis with radiocarbon and OSL chronology led to the conclusion that the sedimentary fill of the Branco River in northern Amazonia is only 18.7 ky old. This inference is supported by the fact that they truncate a Late Pleistocene megafan preserved to the west of the river. This residual landform was formed by sediments sourced from highland areas of the Guiana Shield to the northeast, and a long segment of the Branco River had not been established while the megafan was under development. A plausible hypothesis is that the Branco River was formed by the connection of pre-existing smaller-scale drainage basins. River capture due to tectonic reactivation may have played an important role in diverting a northeastern river system with discharge into the Caribbean Sea, allowing its connection with smaller-scale drainages to form the extended southward flowing Branco River with discharge into the Negro-Amazon basins.

9. Conclusion

The morphological characterization based on remote sensing and sedimentological analysis with radiocarbon and OSL chronology led to the conclusion that the sedimentary fill of the Branco River in northern Amazonia is only 18.7 ky old. This fluvial valley became established after a long time gap of non-deposition and/or erosion that followed the per-flooded even on higher terraces, depositing pelitic sediments on the upper alluvial plain. Therefore, the good correspondence between sediment aggradation in the lower alluvial plain and the wet stage since the mid-Holocene implies that the climate might have played an important role in diverting a northeastern river system with discharge into the Caribbean Sea, allowing its connection with smaller-scale drainages to form the extended southward flowing Branco River with discharge into the Negro-Amazon basins.

**Acknowledgements**

This work was founded by the projects #2010/09484—2 and #13/50475-5, granted by the Research Funding Institute of the State of São Paulo-FAPESP. The Brazilian National Council for Scientific and Technological Development-CNPq and Brazilian Federal Agency for Support ad Evaluation of Graduate Education-CAPES provided the research grant 472131-2009-5 and the doctoral grant BEX 0487/15-5, respectively. The Brazilian Geological Survey—CPRM provided transportation during Burns, 2000). Wet conditions might have contributed to increase the water discharge in the Branco River. This river would have reached bankfull levels and flooded even on higher terraces, depositing pelitic sediments on the upper alluvial plain. Therefore, the good correspondence between sediment aggradation in the lower alluvial plain and the wet stage since the mid-Holocene implies that the climate might have been a key element influencing the latest evolution of the Branco River. This inference is supported by the fact that many modern alluvial plains along Amazonian rivers developed also from the mid-Holocene (latrubesse and Kalicki, 2002; Rossetti et al., 2005; Rozo et al., 2012; Guimarães et al., 2012).

Table 4

<table>
<thead>
<tr>
<th>Sample</th>
<th>Measured age (yrs BP)</th>
<th>$^{13}$C/$^{12}$C</th>
<th>Conventional age ($^{14}$C yrs BP or pMC*)</th>
<th>Calibrated age (2σ in yrs BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>RBN04.1</td>
<td>6840 ± 40</td>
<td>−28.7‰</td>
<td>6780 ± 40</td>
<td>7577–7678</td>
</tr>
<tr>
<td>RBN07.2</td>
<td>590 ± 30</td>
<td>−27.0‰</td>
<td>560 ± 30</td>
<td>588–642</td>
</tr>
<tr>
<td>RBN08.1</td>
<td>6010 ± 30</td>
<td>−22.8‰</td>
<td>6050 ± 30</td>
<td>6825–6981</td>
</tr>
<tr>
<td>RBN10.1</td>
<td>2860 ± 30</td>
<td>−30.8‰</td>
<td>2790 ± 30</td>
<td>2779–2929</td>
</tr>
<tr>
<td>RBN12.1</td>
<td>109.8 ± 0.3*</td>
<td>−31.1‰</td>
<td>111.2 ± 0.3*</td>
<td>86–119</td>
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<tr>
<td>RBN16.3</td>
<td>9170 ± 30</td>
<td>−19.5‰</td>
<td>9260 ± 30</td>
<td>10,366–10,524</td>
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Fig. 10. Chronological framework derived from the present work and their comparison with other paleoenvironmental data from northern Amazonia for the last 70 ky. A - Latrubesse and Franzinelli (1998, 2005); B - D’Apollito et al. (2013); C - Cordeiro et al. (2011); and A.P. - Alluvial Plain.
fieldwork campaigns. Antonio Lisboa and Beatriz Lisboa from the ICMBio—Chico Mendes Institute for Biodiversity Conservation helped with the logistics during part of these campaigns. The authors are also grateful for the comments, suggestions and corrections made by Dr. Takashi Oguchi and two anonymous reviewers, which helped to improve this work.

Map. KML file containing the Google map of the most important areas described in this article.

Appendix A. Supplementary data

Supplementary data associated with this article can be found in the online version doi: http://dx.doi.org/10.1016/j.geomorph.2016.07.030. These data include the Google maps of the most important areas described in this article.

References


