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Holocene floodplain soils along the Río Mamoré, northern Bolivia, and their implications for understanding inundation and depositional patterns in seasonal wetland settings

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Abstract

The Llanos de Moxos (LM) in the Bolivian Amazon basin host one of the largest seasonally inundated savannah landscapes on Earth. Very little is known of this area with regard to sedimentary dynamics, soil formation, or their relationship to longer-term climatic and hydrological variability in this setting. Here we present a detailed study of three floodplain depositional sequences building up the inundation savannah along the Mamoré River in the central LM. Pleistocene sands and silts (unit 1) are overlain by thick overbank deposits of mostly Holocene age (unit 2) on both sides of the Mamoré River, and underline the importance of extensive flooding processes for the late Quaternary sedimentary and geomorphic evolution of the LM. The fine-grained distal overbank sediments show signs of strong modification by hydromorphic processes and reflect spatial and temporal variations in flood inundation patterns. In addition, widespread dark gray to blackish soil and paleosol horizons are intercalated with the overbank sediments, but lack evidence for significant weathering. These horizons are likely the result of cumulative incorporation of organic material under conditions of particularly low sedimentation rates. Therefore, their formation

should be linked to processes such as channel migration, longer-term meanderbelt evolution, avulsive drainage reorganization, changes in sediment supply or climatically induced variations in flooding type, frequency or magnitude. The common occurrence of these floodplain soils in the early to mid-Holocene may thus reflect the combined effects of hydrological, geomorphic and sedimentary changes during a drier mid-Holocene.

Keywords: seasonal wetlands, savannah, Amazon, hydromorphic soils, floodplain, paleosol

1. Introduction

The lowlands of the Amazon basin host some of the largest rivers on Earth, combine up to ~15% of the world's freshwater runoff (Gaillardet et al., 1997) and play a major role for our understanding of weathering, erosion, and sediment transport, storage and recycling over geological timescales (Berner et al., 1983; Moreira-Turcq et al., 2004; Bouchez et al., 2012). Therefore, significant efforts have been dedicated to the geochemical and mineralogical characterization of runoff and sediment load throughout the Amazon basin (Gibbs, 1967; Franzinelli and Potter, 1983; Roche and Jauregui, 1988; Konhauser et al., 1994; Elbaz-Poulichet et al., 1999; Guyot et al., 1999, 2007; Vital and Stattegger, 2000; Bouchez et al., 2012). Particularly in the Andean foreland, floodplains and their backswamps represent important traps for much of the sediment produced in the headwaters of the Andes and/or the Brazilian shield (e.g., do Nascimento et al., 2015). Despite the large potential that riverine wetlands and floodplains have for assessment of longer-term sediment production, weathering and transport rates in the Amazon basin, few studies have utilized these extensive sedimentary archives (Martinelli et al., 1993; Aalto et al., 2003; Dosseto et al., 2006; Wittmann et al., 2009, 2011; Melack and Hess, 2011; Moquet et al., 2011) and most have only addressed relatively recent timescales.

Along the headwaters of the Río Madeira and its tributaries Beni and Mamoré Rivers in northern Bolivia, much of the floodplain coincides with the Llanos de Moxos (LM) - one of the largest seasonally inundated savannah landscapes on Earth (Hamilton et al., 2002). In this extremely flat region, more than ~80,000 km² can be subject to up to four months of inundation in extreme years (Hamilton et al., 2004) (Fig. 1). In contrast to their counterparts in coal-forming perennial wetlands, however, very little is known with regard to the strongly

seasonal character of soil formation and sedimentary dynamics in inundation savannah settings or their relationship to longer-term variability of environmental and hydrological conditions (Wright et al., 2000; Greb et al., 2006). This is despite the major role these particular tropical and subtropical environments play for the sustainable use of hydrological resources on a global scale (Dugan, 1992), and limits their use as modern analogues for the interpretation of buried soils (hereafter termed paleosols) in alluvial sediments in the rock record (e.g., Platt and Keller, 1992; Kraus, 1997; Aslan and Autin, 1998; Wright et al., 2000; Varela et al., 2012; Beilinson and Raigemborn, 2013). Alluvial surface soils in the LM show abundant pedological evidence for dominantly hydromorphic conditions (Boixadera et al., 2003). Recent studies on the type and chronologies of the paleochannel systems in the LM inundation savannah have documented large-scale changes in inundation patterns and/or frequencies, likely as the result of the interplay between environmental and tectonic drivers (Plotzki et al., 2011; Lombardo, 2014; Plotzki et al., 2015). In this context, periods of reduced flooding intensities in the southeastern LM seem to be expressed in the development of regional-scale paleosols, which underlie mid- to late Holocene fluvial and lacustrine sediments (Lombardo et al., 2012), or are exposed in cut-banks along the outer meanderbelts of most major rivers in the LM (Iriondo, 1999). These paleosols thus seem to represent major marker horizons in floodplain stratigraphy, but so far no detailed data are available to clarify their significance with regard to their paleoenvironmental, hydrological, geomorphic or even tectonic controls.

In order to characterize the pedogenic and sedimentary processes underlying the formation of these floodplain soil/paleosol-sediment-sequences, we here present a detailed study of three floodplain depositional sequences building up the inundation savannah along the Río Mamoré in the central LM. Our combined pedogenic, sedimentary, mineralogical,

geochemical, and chronological dataset therefore provides an opportunity towards exploring inundation patterns, sedimentary dynamics, and tectonic evolution in the southern Amazon basin over the late Quaternary, and more specifically the Holocene. In addition it contributes new data towards an improved understanding of sedimentary and pedogenic processes operating in terrestrial seasonal wetland environments in general.

2. Regional setting

The study area is situated in the Llanos de Moxos in northeastern Bolivia (Fig. 1), which are confined by the crystalline Brazilian Shield to the east, and by the Andes in the southwest. Elevation above sea level is around 150 m (Hanagarth and Sarmiento, 1990). The LM are drained by the Mamoré River toward the Madeira River in the north, and are very flat with gradients of less than 10 cm km^{-1} (Dumont and Fournier, 1994). The Mamoré River originates at the footslopes of the Eastern Cordillera. As a white-water river it carries a significant load of sediments ($64 \times 10^6 \text{ t a}^{-1}$ at the confluence with the Beni River; Guyot et al., 1999). Mean annual discharge of the Mamoré River at Trinidad (Fig. 1) is $2850 \text{ m}^3 \text{ s}^{-1}$. Inner-annual variability is high with discharge ranging from less than $500 \text{ m}^3 \text{ s}^{-1}$ in the dry season to $7600 \text{ m}^3 \text{ s}^{-1}$ in the rainy season (Ronchail et al., 2005; HYBAM-Observatory / ORE Hybam Observatoire de Recherche en Environment, 2014). In contrast to rivers with Andean headwaters, clear-water rivers originate in the plains or in the Brazilian Shield, and carry only minor amounts of sediments (Sioli, 1984; Bourrel et al., 2009).

Climate in the LM is controlled by the South American Summer Monsoon, leading to heavy convective rainfall in austral summer and markedly dry conditions in winter (Zhou and Lau, 1998; Garreaud et al., 2009). Peak discharge during austral summer causes inundations for

several months per year (Bourrel et al., 2009). Inundations cover an area between 30,000 and 78,000 km², occasionally reaching 150,000 km² (Hanagarth and Sarmiento, 1990; Hanagarth, 1993; Hamilton et al., 2004). Two main mechanisms of flooding are distinguished as the result of spatial variations in precipitation patterns. With dominant rainfall over the Andes, white-water river flood peaks tend to be rapid, and overbank flow has the potential to transport sediment far into the floodplain (allogenic flooding); in contrast, dominant precipitation in the lowlands leads to waterlogging of soils and the slow dispersal of flood water from clear-water rivers into the floodplain (autogenic flooding), restricting sediment transport into the floodplain (Fig. 2a, b), and reducing the potential of geomorphic modification (e.g., crevassing) along the white-water rivers (Aalto et al., 2003; Bourrel et al., 2009). Tree vegetation in the LM savannah landscape is generally limited to slightly elevated areas less affected by the seasonal inundations such as forest islands and gallery forest along fluvial levees (Beck, 1983; Hanagarth, 1993).

3. Methods

3.1 Field work

Along the Mamoré River, the gallery forest is locally interrupted where meanders have cut into the adjacent grasslands of the inundation savannah at the outer margins of the Mamoré River meanderbelt. Here, low water levels in the dry season (June - October) expose floodplain soil/paleosol-sediment-sequences in the outer cutbanks (Fig. 2c, e). We studied two of these exposures ~15 km west of Trinidad on the western river bank (site Trinidad-1/M-2a, depth 7 m; Fig. 2a, c) and near San Pedro ~50 km downstream of Trinidad on the

eastern bank (site M-4, depth 10 m; Fig. 2a, e). In addition, a complementary sediment core of 8 m length and 4 cm diameter was extracted in the floodplain ~3.5 km southwest of M-2a with a motor-driven percussion corer (M-2b, Fig. 2a, d). Sedimentary characteristics and soil properties such as colour, grain size, redoximorphic and pedogenic features were described for the two profiles and the sediment core according to FAO (2006). Bulk samples for laboratory analysis were taken from all major layers and horizons in M-2a, b and M-4.

3.2 Laboratory analyses

Bulk sediment samples were stored in plastic bags, air-dried, and separated in two aliquots. From one aliquot sesquioxides were removed from the samples by adding 0.3 M sodium citrate-solution, 1 M sodium carbonate-solution and sodium dithionite, and organic matter was removed by oxidation with 30% H_2O_2 . A solution of sodium hexametaphosphat with sodium carbonate was added as dispersant agent. Grain size measurement was conducted with the Malvern Mastersizer 2000 (Hydro S) with a stirring speed of 2000 rpm and a measurement time of 50 s. Fraunhofer theory was used for calculating the particle size distribution results. Results were recalculated to “pipette clay content” with the equation by (Konert and Vandenberghe, 1997) in order to compensate for the underestimation of clay content by the laser particle measurement with respect to the pipette method (i.e., the international standard method) (Di Stefano et al., 2010). For the other aliquot, soil colour was determined on moist samples using the Munsell soil colour chart. The air-dried samples were milled (vibration disk mill). Total organic carbon and total nitrogen content were measured with a VaRío Macro Elemental Analyzer. Also, organic loss on ignition (LOI) was measured following combustion at 450°C. For determination of elemental compositions pellets were

pressed with a mixture of 4 g sample material and 0.9 g wax (Licowax C Micropowder PM), and measured by x-ray fluorescence spectroscopy (Phillips 2400 XRF). Finally, clay mineralogy was analysed by x-ray diffractometry for selected samples (XRD, Phillips PW 1830). The diffractograms of Mg^{2+} saturated specimens were obtained for normal conditions, after solvation with ethylene glycol and after heating to 550°C. Then, amounts of clay minerals were measured semi-quantitatively by XRD peak height (Biscaye, 1965; Brindley and Brown, 1980). In addition, selected diffractograms from profile M-2a and M-4 were visually checked for minor peaks indicating the presence of iron oxides such as hematite, goethite, and lepidocrocite.

3.3 Geochronology

A total of 15 samples was taken to establish a chronological framework for the studied floodplain soil/paleosol-sediment-sequences. AMS ^{14}C analysis was conducted at the Poznan Radiocarbon Laboratory. For four samples, charcoal or plant remains were dated. For the remaining samples, humins and humic acids from the bulk organic sediment were dated separately in order to assess possible contamination of the samples with older (Räsänen et al., 1991) or younger carbon as the humic acid fraction provides crucial information about the degree of contamination and thus age reliability (Pessenda et al., 2001; Grootes et al., 2004). Radiocarbon ages were calibrated using Calib 7.1 online (Stuiver and Reimer, 1993) in combination with the Southern Hemisphere calibration data set (Hogg et al., 2013) and are indicated as cal ka BP.

Four additional samples were taken for Optically Stimulated Luminescence (OSL) by forcing opaque steel tubes into the sandy sediments. Material from the outer parts of the tubes was

used for determination of dose relevant elements, whereas equivalent dose (D_e) was measured on material extracted from the inner part of the tubes. Carbonates and organic matter in the samples were dissolved using HCl and H_2O_2 , respectively. Only one sample contained sufficient coarse, sand-sized material, and fine grain measurement was conducted for all samples. For this purpose, the 4–11 μm fraction was enriched by settling following Stokes' law and subsequently etched with H_2SiF_6 , followed by HCl treatment. For additional coarse grain measurement (sand fraction, see Table 3), quartz grains from sample AP-1 were isolated using heavy liquid separation and subsequently etched with 40% HF for 60 min, followed by HCl treatment. The single-aliquot regeneration (SAR) protocol of (Murray and Wintle, 2000) was applied for Equivalent Dose (D_e) determination, using a Risø TL/OSL DA-20 reader equipped with blue diodes and Hoya U-340 detection filters. The purity of samples (i.e., the presence of feldspars) of 2 mm aliquots (ca. 100-150 grains) was checked by exposure to IR diodes. Preheating at 230°C for 10 s was used prior to all OSL measurement, for which we found good reproducibility in dose recovery tests and negligible thermal transfer (Supplementary Data Table 1). Palaeodose calculation is based on ca. 5 D_e measurements for the fine grain fraction and ca. 50 individual D_e measurements for the coarse grain fraction. Due to the small number of grains on the aliquots we expect that the OSL signal is dominated by emissions from very few or even a single grain. If incompletely bleached grains are present, the Minimum Age Model (MAM) of Galbraith et al. (1999) should be used, as has previously been applied to sediments in the LM (Plotzki et al., 2013, 2015). However, application of the MAM is restricted to coarse grain samples. As a consequence, we used the MAM assuming an overdispersion value of 0.20. Where only fine grain samples could be prepared, the Central Age Model (CAM) was applied. Therefore, results based on fine grain samples should be regarded as maximum ages. The concentration of dose rate relevant elements (K, Th, U) was determined using high-resolution low-level

gamma spectrometry. No indication for radioactive disequilibrium in the U decay chain was observed (Preusser and Degering, 2007; Zander et al., 2007). Dose rates and OSL ages were calculated using ADELE software (Kulig, 2005). Cosmic dose rate was calculated using present day depth and correcting for geographical position (latitude, altitude) following the approaches in Adamiec and Aitken (1998). Present day water content and the field hydrological context was used to estimate average water content during sediment burial. All OSL ages are indicated in ka.

4. Results

4.1 Sedimentary and pedogenic characteristics

All studied soil/paleosol-sediment-sequences share some general sedimentary and pedogenic characteristics, but also differ in a number of aspects. Firstly, all sequences are subdivided into a coarser basal part mostly consisting of loam and sandy loam (unit 1) which is overlain by ~4-5 m of very fine-grained clay deposits (unit 2; Fig. 3, Table 1). With ~55-85%, clay is the dominant grain size fraction in the fine-grained deposits of unit 2, but silt content is highly variable (~15-45%) and even minor amounts (up to ~5%) of fine sand are present (Fig. 4). Sediments do not show signs of stratification and appear mostly massive, but exhibit blocky angular structure in the upper 2 (M-2a) to 3 m (M-4) (Table 1). Due to the high clay content ped surfaces often appear bright, but no clay cutans or slickensides were observed macroscopically.

Throughout the profiles, matrix colours are very variable and range from gray and brown to yellowish and orange (Table 1). Particularly in unit 1, matrix colour is dominantly gray

(brownish to light gray). In the upper portions of unit 2, the profiles are characterized by horizons of generally darker brownish or yellowish colours (e.g., 10YR 5/3, 10YR 4/2 or 2.5Y 4/2), which vary in thickness between 20 and 150 cm at M2-a and b, respectively. They are overlain by thin veneers of lighter sediment at M-2a and b, and show desiccation cracks at M-4 (Fig. 3). Horizons of markedly darker brown, yellow, gray or black colour (e.g., 5YR 3/1 - 4/2, 2.5Y 5/2, 7.5YR 4/1) and variable thicknesses of 40-80 cm are present in the lower part of unit 2 (Fig. 3, Table 1). All of these horizons have clear upper boundaries and pronounced but more gradual lower boundaries. They were described as surface (Ah) and buried (Ahb) soil horizons based on the contrast with the over- and underlying sediments of lighter colour (C horizons). The presence of mm-scale gypsum crystals roughly coincides with these dark horizons in M-2a and M-4 (Fig. 3). They occur in ~200 cm, and ~100 and 270-400 cm depth, respectively, but were not observed in the M-2b core.

Large parts of all profiles show intense mottling as typical for hydromorphic overprinting and are thus referred to as Cg (unit 2) and II-Cg (unit 1) horizons. In the fine-grained deposits of unit 2, mottles are commonly up to 1 cm in diameter and vary between reddish brown and yellowish brown colours. In the more sandy loams of unit 1 at the base of the profiles, mottling is expressed in up to 10 cm masses of brighter browns and orange. While in M-2a and M-4 reddish brown dominates the colours in the upper half of the profiles, M-2b shows a reverse pattern with more yellowish mottles in the upper 200 cm. Black and mm-scale iron and manganese concretions were also observed below 260 cm at M-2a, and between 110 and 300 cm at M-4, respectively.

4.2 Clay mineralogy

Based on the data from selected samples at M-2a and M-4, clay mineralogy in both soil/paleosol-sediment-sequences is dominated by illite (~30-60%) and kaolinite (~30-55%) which together account for 80-95% of the clay (Fig. 5, Supplementary Data Table 3). Minor amounts of smectite (up to 20%) and illite/smectite mixed layer clays (up to 10%) occur in the profiles. In contrast, only negligible amounts of chlorite (~0-1%) are present in M-2a and M-4, whereas modern fluvial sediment consistently contains up to 19% of chlorite (Supplementary Data Fig. 1). In general, however, the clay mineralogical composition corresponds well with clay mineralogy measured in bedload and suspended sediment along the Mamoré River (Guyot et al., 1993, 2007). Even though highest smectite contents were measured in the Ahb2 at M-4, no clear relationship seems to exist between the paleosol and soil horizons (Ah/Ahb) and clay mineralogy across all profiles. Finally, the presence of goethite and lepidocrocite was documented by visual interpretation of diffractograms to a depth of ~305 cm at M-2a and ~260 cm at M-4 (Supplementary Data Fig. 1). With one exception, no iron oxides were identified in the diffractograms below these depths.

4.3 Geochemistry

The down-profile variation of textural and selected geochemical parameters in M-2b and M-4 further characterises the soil/paleosol-sediment-sequences at our study sites (Fig. 4, Supplementary Data Table 2). Grain size can be used to distinguish the coarser deposits of unit 1 at the base of the sequences from the finer clay deposits of unit 2 above (Fig. 4). The fine-grained deposits of unit 2 can also be clearly distinguished from the coarser unit 1 based

on their significantly higher carbon (Fig. 6a) and nitrogen content, pointing to two contrasting different depositional environments for these two units. There is no clear relationship in grain size between the Ah and Ahb, and under- and overlying C/Cg horizons. In particular the clay/silt ratio as a measure of in-situ weathering (Wambeke, 1962) is highly variable and does not correspond to soil and paleosol horizons (Fig. 4), thus suggesting that textural variability is directly related to depositional processes. In contrast, Ah and Ahb horizons are characterized by elevated C/N values of up to ~4.5 (Fig. 4) and overall higher LOI values of ~10-25% (Fig. 6b), confirming the overall minerogenic character of these organic-rich horizons in contrast to the less organic sediments of the C horizons. Highest values of carbon and nitrogen are generally found in the Ah horizons of the upper profiles, whereas values of <1.25% (C) and <0.15% (N) are characteristic of the buried fAh horizons, pointing to the ongoing degradation of organic matter in the former (Fig. 6c).

The distinction of paleosol/soil horizons (Ah, Ahb) versus sediment (C and Cg horizons) is not well reflected in the total element data (Supplementary Data Table 1). Even though the weathering indices show slightly elevated values in some organic paleosol/soil horizons (Figs. 4, 6d), this relationship is not systematic and high values are also characteristic in some C horizons. In both profiles, however, commonly used weathering indices such as the Chemical Index of Alteration (CIA) (Kronberg and Nesbitt, 1981) or the Rb/Sr ratio have significantly higher values than modern fluvial sediment samples from the Mamoré River (Fig. 6d). Lower CIA values in the basal loams and sandy loams of unit 1 point to reduced weathering intensities, and possibly indicate a relationship with grain size. Particularly at M-4, the Cg horizon (~100-300 cm) exhibits overall low values and high variability which is not reflected in grain size alone. In combination, this seems to imply a limited degree of in-situ weathering unrelated to the formation of Ah and/or Ahb horizons, suggesting that chemical

variations in the profiles are reflective of (i) processes affecting the whole depth of the deposit (i.e., hydromorphy), and/or (ii) variations in chemical composition of parent material (i.e., sediment provenance).

Fe₂O₃ values range between ~4 and 12% in both profiles but show some marked difference in down-profile trends (Fig. 4). M-4 is characterised by a gradual down-profile decrease in Fe₂O₃, whereas values in M-2b are more variable and show a pronounced peak in Fe₂O₃ in ~180 cm depth in the C horizon (Fig. 4). In both profiles, these trends are also resembled by the P₂O₅ content, which is characterized by overall values of ~0.1% but exhibits a marked peak coinciding with the Fe₂O₃ peak in M-2b. In contrast to Fe₂O₃ and phosphorus, manganese (MnO) and sulphur (S) values are generally low throughout the profile at M-2b, but exhibit marked peaks coinciding with the C horizon between ~100 and 300 cm depth at M-4. In addition, minor peaks also characterize the uppermost parts of the Ah horizon at M-2b and the fAh2 horizon at M-4 (Fig. 4). In summary, neither Fe₂O₃, MnO nor S exhibit obvious relationships with the Ah or Ahb horizons (Fig. 4). Instead, variation in the hydromorphically mobile elements such as Fe, Mn and S (Brümmer, 1974) is best explained when compared against the colour of mottling the profiles (Fig. 6e, Table 1). In this context, yellowish brownish colours have lowest overall values in MnO and Fe₂O₃, as expected for dominantly reducing conditions. In contrast, highest measured values of Fe₂O₃, MnO and S correspond to samples in which black Mn concretions were observed in the field. Interestingly, high S values also coincide with the observation of gypsum crystals in the field. Therefore, hydromorphic and redox conditions as expressed in the intense mottling seems to exert an important control on the total element composition in the profiles.

Both Ti and Zr are considered immobile (e.g., Allen and Hajek, 1989), and thus contain valuable information with regard to sediment provenance. In our profiles, however, Zr content is strongly related to grain size (e.g., fine sand, $r^2=0.80$, Supplementary Data Table 1). In contrast, TiO_2 content shows no relation to grain size ($r^2=-0.13$) but exhibits both strong down-profile variations and differences between both profiles (Fig. 4). At M-4, overall TiO_2 content is low and does not show variability within the fine-grained unit 2, whereas higher values characterize the coarse loams of unit 1. At M-2b overall TiO_2 content is higher with a slight increase towards the top of the profile. In the southeastern LM, the K/Mg ratio has recently been suggested to reflect variations in provenance with lower values indicative of increased contributions from Andean source areas (Plotzki et al., 2015). Similar to TiO_2 content, this ratio does not show systematic differences between C/Cg and Ah/Ahb horizons (Fig. 4). Instead, the down-profile variations in TiO_2 and K/Mg exhibit some similarities between both profiles (Fig. 4) with highest values towards the top. Whereas no significant difference in TiO_2 and K/Mg can be observed between units 1 and 2 at M-2b, the finer-grained sediments of unit 2 at M-4 show much lower and/or variable values in comparison to unit 1. Particularly in combination with TiO_2 content, the K/Mg ratio allows the distinction between the more recently deposited sediment within the uppermost 50-100 cm (i.e., Ah horizons) from the underlying deposits (Fig. 6f). The high TiO_2 content and K/Mg ratios in Ah horizons compare well with samples from modern fluvial sediment in the Mamoré River (Fig. 6f).

4.4 Chronology and deposition rates

A total of 13 samples was measured for ^{14}C dating, and the results span a period between ~2.6 and 13.7 cal ka BP. Two samples on charcoal and plant remains yielded modern ages

and are not further considered for the chronological framework of our profiles. For nine samples, paired measurements of humins and humic acids were conducted, and show good correspondence for eight of them (<25% age difference, generally between 1-11%). Only one sample (M-2b at 180 cm depth) resulted in a major age difference of ~5 ka between both fractions and is consequently not further considered. The general accuracy of the remaining radiocarbon ages is corroborated by three fine grain OSL ages from the fine-grained upper part of the profiles, which yielded ages of 2.94 ± 0.37 , 5.14 ± 0.37 and 15.8 ± 1.16 ka, and are in good stratigraphic correspondence with the radiocarbon ages (Tables 2, 3). A paired fine and coarse grain OSL age is available for an additional sample from the coarse-grained sediments of unit 1 in the basal part of M-4. Here, the fine grain age of 37.4 ± 2.8 ka is more than twice that of the coarse grain age (17.8 ± 2.5 ka). Given the difference in depositional environments and dated material between this sample and the samples in the overlying clays, this may well be related to the post-depositional vertical mobility of fine grains within the coarser, sandy sediment matrix, and/or the presence of partial bleaching as a common problem with dating fine-grained sediments from a fluvial context (Wallinga, 2002; Hu et al., 2010). In combination with the stratigraphic correspondence with the overlying ^{14}C and OSL ages, this suggests age on coarse grains as more reliable.

Based on these considerations, our results imply mostly Holocene ages for the clays in the upper parts of the profile, and late Pleistocene ages (~11.8-18 ka) for the coarser deposits in the basal parts. At M-2a the prominent organic paleosol horizon (Ahb2) has likely formed between ~9.2 and 6.3 cal ka BP. At M-4 a similar paleosol horizon (Ahb2) is somewhat older and dates to the time between 12.9 until 7.3 cal ka BP, whereas a younger organic horizon (Ahb1) is late Holocene in age (~3.6-2.6 cal ka BP).

Additionally, sediment deposition rates were calculated from the coarse grained MAM OSL age, as well as radiocarbon ages from humins and charcoal (Table 2). The resulting rates are very variable and range from $0.12 \pm 0.01 \text{ mm a}^{-1}$ in the Ahb2 at to $2.36 \pm 0.89 \text{ mm a}^{-1}$ within the coarse grained basal sediments at M-2b. Interestingly, the rates show a very good correspondence with the type of soil/paleosol horizon (Fig. 7). While rates calculated for Ahb horizons and the combined Ah-C horizons in the upper parts of the profile are well below 0.4 mm a^{-1} , all remaining rates are above this value. Even though only four additional rates are available, the results seem to show comparatively low but slightly higher values ($0.4 \pm 0.20 \text{ mm a}^{-1}$ and $0.52 \pm 0.01 \text{ mm a}^{-1}$) for sediments (C/Cg horizons) in the fine-grained clays of unit 2. In contrast, ages within the coarse basal deposits of unit 1 at M-2b yield much higher rates (Fig. 7), underlining the difference in depositional processes between the two units.

5. Discussion

All three profiles share three essential characteristics with regard to their sedimentary and pedological characteristics, which form the base for the interpretation of depositional environments and post-depositional modification in the Mamoré River floodplain: they exhibit coarser sediments at the base, contain dark soil and paleosol horizons, and are strongly affected by redoximorphic features.

5.1 Late Quaternary variations in depositional environments

Firstly, the profiles are divided into two distinct sedimentary/lithological units. Unit 1 at the base of the profiles consists of loams, sandy loams and sands, and unit 2 predominantly comprises 4-5 m of clays and loam. A number of processes are capable of transporting

sediments in floodplain settings but grain size generally decreases as a function of distance away from their riverine source (Guccione, 1993; Asselman and Middelkoop, 1995; Cazanacli and Smith, 1998). The coarse sands and loams of unit 1 are therefore likely related to deposition in proximal floodplain settings such as a levee or channel. In fact, clay content of ~15-25 % in unit 1 corresponds well with data from samples associated with crevasse splays in a levee setting along the Mamoré and Beni Rivers as reported by Aalto et al. (2003). In contrast, the extremely fine-grained sediments of unit 2 are interpreted to reflect deposition in a distal floodplain setting (i.e., “flood basin”, “backswamp”) (Aslan and Autin, 1998; Kraus and Aslan, 1999) by decanting of suspended sediments originating from the main white-water rivers such as the Mamoré River (Aalto et al., 2003; Gautier et al., 2007). However, varying amounts of silt and fine sand indicate variations related to the distance to the channel as nearest sediment source, or to the type or energy of the depositional process itself. In particular fine sand layers within unit 2 may be interpreted as the result of more dynamic processes such as crevasse splays penetrating into the more distal floodplain.

The transition from coarser-grained levee or channel deposits of unit 1 to the clayey deposits corresponding to overbank deposition in a distal floodplain (unit 2), could be interpreted as the result of lateral migration and avulsive shifts in the Mamoré River meanderbelt (Aslan and Autin, 1998). However, the similarity of the sedimentary sequences on both sides of the river (M-4 vs M-2a and b), and the complete lack of levee and/or channel facies within unit 2, rather point to a more regional shift in fluvial processes from a bedload and coarse-load dominated Pleistocene system towards more suspended load dominated systems during the Holocene (Fig. 2a, b). Significant climate-driven changes in vegetation patterns were reported from the southern Amazon basin at the Pleistocene-Holocene transition (Burbridge et al., 2004; Mayle et al., 2004; Whitney et al., 2011), and may have also affected the fluvial

regimes in the Peruvian and Bolivian lowlands by changing sediment yields and runoff in the headwaters (May et al., 2008b; Rigsby et al., 2009; Latrubesse et al., 2012). In fact, paleochannel dimensions demonstrate that large meandering systems drained the southeastern LM in lateglacial times, also showing that the overall drainage network has undergone major spatial and hydrological transformations since the late Pleistocene (Plotzki et al., 2015). The exact timing of change along the Mamoré River is hard to constrain based on the few studied sites, but has probably occurred sometime between ~11 and 18 cal ka BP. Clearly, an improved chronological framework and additional sedimentary sequences in the LM are required to test the scenario of a major change in fluvial regime towards the end of the Pleistocene.

5.2 Pedogenic processes in the distal floodplain

All profiles and both stratigraphic units are strongly affected by mottling which in turn expresses the post-depositional modification by hydromorphic processes as the result of varying redox conditions and microbial processes (Brümmer, 1974; Berthelin, 1988; Van Breemen and Buurman, 1998). Mottles vary from yellowish and brownish, to reddish brown and even blackish, indicating the varying presence of goethite and lepidocrocite as confirmed by XRD data, or ferrihydrite in the profile (Schwertmann and Taylor, 1989; Scheinost and Schwertmann, 1999). In particular in combination with grayish matrix colours, the orange mottle colours as indicative for lepidocrocite and ferrihydrite in the coarse-grained unit 1 at the base of all profiles reflect frequently waterlogged conditions related to a higher average seasonal groundwater table in proximity to the channel (Aslan and Autin, 1998) and/or favoured by the porosity of the sediment. The strong influence of repeated, seasonal reduction and oxidation is also reflected by elements preferentially affected by

hydromorphism such as iron, manganese or sulphur (Brümmer, 1974), which vary with colour and type of mottling, thus showing that hydromorphic processes have a marked influence on the overall chemistry in the profiles. More specifically, mottles are smaller and are mostly reddish brown in the overlying unit 2 pointing to the widespread presence of goethite and poorly crystalline ferrihydrite in the upper parts of profiles M-2a and M-4. Below ~300 m no crystalline iron oxides seem to be present, generally confirming the strong influence of waterlogging related to the groundwater table in the lower part of the profiles. Interestingly, in profile M-2b only a few mottles of yellowish brown colour as indicative of goethite (Scheinost and Schwertmann, 1999) were observed in the upper ~3 m. In comparison with profiles M-2a and M-4 at the river banks, this may suggest stagnic conditions and processes of pseudo-gleying (Wright et al., 2000), thus pointing to generally lower groundwater tables but frequently stagnating surface water in the more distal setting of M-2b. Thereby, mottling colours may reflect a difference in overall hydromorphic conditions affecting the profile, suggesting the presence of a seasonal wetland catena with soil formation strongly influenced by the distance to the Mamoré River (Fig. 8a), and controlled by a combination of hydrologic setting (e.g., flooding regime, drainage conditions, groundwater table), sedimentary characteristics, and subtle variations in topographic relief (Aslan and Autin, 1998; Wright et al., 2000).

The thicker Ah horizon and the absence of gypsum crystals in the closed core of profile M-2b in contrast to the exposures along the river banks (i.e., M-2a, M-4) may provide a further hint to the differing hydromorphic setting between the profiles. The formation of secondary gypsum crystals from ascending evaporation fronts near the capillary fringe is common in floodplain soils (Aslan and Autin, 1998), but the coincidence of high Mn, Fe_2O_3 , and S content with Mn concretions and gypsum crystals may also imply a relationship between

gypsum and redox conditions. In this context, preferential precipitation of gypsum may be facilitated by oxidation of ferrous sulfide which could have formed previously under more reducing conditions within the more organic Ah and Ahb horizons (Aslan and Autin, 1998).

These dark (gray to blackish) horizons of variable thickness formed in unit 2 were described as soil and paleosol horizons in the field based on their colour properties. Although dark matrix colour can result from various soil properties (i.e., iron sulfide, manganese oxide, humus, microcharcoal), these horizons can be clearly distinguished from the surrounding sediment (i.e., C/Cg horizons) only based on their characteristically high organic content (e.g., LOI, C and N). Thus, their organic content provides the most likely explanation with regard to their gray to blackish colour (Shields et al., 1968; Calhoun et al., 2001), which can persist even if the organic matter is oxidized and mineralized (Holliday, 2004) and the resulting organic carbon values are comparatively low (Aslan and Autin, 1998). A relatively low degree of humification in these horizons is expressed by high C/N ratios, and is common under warm and particularly wet conditions (Post et al., 1985). Values generally agree with other published results from the LM (Hanagarth, 1993). Particularly for Ahb horizons, high C/N ratios and relatively low nitrogen content may also reflect nitrogen loss by denitrification of organic matter decomposition under waterlogged, anaerobic conditions (Groffman and Tiedje, 1988; Pinay et al., 2000).

Apart from their organic content, however, these horizons do not show evidence for particularly strong in-situ weathering. While soil profiles in well-drained sandy to silty abandoned levee deposits in the southeastern LM (Fig. 8b) are characterised by the presence of clay cutans and well-developed Bt horizons (Lombardo et al., 2015), in our profiles there is no indication for clay translocation such as cutans. This could be the result of Pleistocene

soil ages in the southeastern LM in comparison with the relatively young, Holocene ages in our profiles. More likely, however, this reflects the significantly finer texture of the distal floodplain deposits, which effectively delays clay illuviation in the profile, but which may also be the combined effect of poor drainage conditions and continued, slow sedimentation in contrast to the elevated topography on the levees in the southeastern LM (Aslan and Autin, 1998; Lombardo et al., 2015). Despite some desiccation cracks developed in the upper part of the profile at M-4, there is no field evidence for slickensides reflecting vertisol formation and self-mulching, due to shrinking and swelling in a seasonally inundated floodplain environment. While this may be interpreted as the result of insufficient exposure and drainage (Wright et al., 2000), it could also be the result of sand and silt content within the clay deposits (Aslan and Autin, 1998), and/or the relatively low percentage of expandable clay minerals (e.g., smectites) in our profiles compared with more distal locations in the LM where clay related mulching is common (Boixadera et al., 2003). Only negligible amounts of chlorite were detected in our profiles, although chlorite is clearly present in clays extracted from suspended sediment and bedload in the modern Mamoré River (Guyot et al., 1999). This is in agreement with Boixadera et al. (2003), and may indicate the complete decomposition of pedogenically unstable chlorite (Bain, 1977; Allen and Hajek, 1989; Velde, 2012), or its transformation into illite/smectite mixed layer clays (Righi et al., 1993). In seasonally reducing environments such as the LM, illite/smectite mixed-layer clays also form from illites and even smectites (Brinkman et al., 1973; Righi et al., 1993).

In comparison with fluvial sediments from the Mamoré River, our results indicate some degree of in-situ weathering (i.e., transformation of clays) in the sediments of all profiles. However, weathering indices and the clay mineralogy in the Ah and Ahb horizons do not significantly differ from the surrounding sediment (C/Cg horizons), implying that in-situ

weathering processes do not explain the formation of the Ah/Ahb horizons. Instead, some degree of in-situ weathering as reflected in elemental data may be related to variations in parent material such as changes in provenance or the varying contribution of recycled and pre-weathered floodplain sediments farther upstream (Dunne et al., 1998), or to redox processes as reflected in the strongly developed redoximorphic features in our profiles (Aslan and Autin, 1998).

The sediment deposition rates as calculated from our radiocarbon and OSL based chronology clearly seem to corroborate this scenario, and are lowest in the Ah/Ahb horizons. The sedimentation rates associated with the Ah/Ahb horizons are slightly lower than rates in C horizons, and an order of magnitude lower than rates reported from ^{210}Pb dating of flood sediment along the Mamoré River in close proximity to the channel (Aalto et al., 2003). Whilst this difference could be interpreted as an artefact of different methods and timescales of analysis, in general there seems to be a good agreement between the various methods for determining floodplain sedimentation rate (Hobo et al., 2010). Even though our deposition rates need to be considered with caution as a result of the remaining uncertainties in relation to our limited chronological dataset, they compare well with the range of most reported rates in distal floodplain environments (Kraus and Bown, 1993; Walling and He, 1998; Makaske et al., 2002; Alexandrovskiy et al., 2004) whereas published rates in proximal floodplain settings correspond to the reported overall higher rates in proximal floodplain and levee environments (Mertes, 1994; Walling and He, 1998; Moreira-Turcq et al., 2004; Hobo et al., 2010). This confirms that sediment supply and sedimentation rate decrease exponentially away from the channel (Guccione, 1993; Asselman and Middelkoop, 1995; Cazanacli and Smith, 1998; Törnqvist and Bridge, 2002), representing a first order control on the formation of Ah and Ahb horizons within the distal floodplain sediments of unit 2. Therefore, the Ah

and Ahb horizons could to be interpreted as cumulic soil horizons which would have formed under very low but continued deposition of overbank sediment. In this context, the formation of cumulic soil horizons does not reflect geomorphic stability, and is not necessarily linked to climate or regional-scale environmental conditions (Alexandrovskiy et al., 2004), but needs to be interpreted as a threshold dominated response to sedimentation rates below $\sim 0.3 \text{ mm a}^{-1}$. As more time is available between sedimentation events, soil maturity is expected to increase with lower sedimentation rates (Bown and Kraus, 1987; Alexandrovskiy et al., 2004). In our profiles, however, Ah/Ahb horizons only show minimal weathering intensities as typical for Entisols and Inceptisols, very likely reflecting the dominance of hydromorphic processes which largely inhibits other pedogenic processes (Kraus and Aslan, 1993; Aslan and Autin, 1998).

In combination, our data suggest that the soil/paleosol-sediment-sequences along the Mamoré River are subject to two interacting variables in the distal floodplain environment (Fig. 8a): (i) redox features, hydromorphic conditions and overall weathering in the profiles appear to be controlled by a the hydrologic setting (e.g., flooding duration, groundwater depth) and properties (e.g., porosity, grain size), possibly forming seasonal wetland catenas similar to other extensive wetland settings in the world (Aslan and Autin, 1998; Wright et al., 2000); in addition, (ii) the cumulic but weakly weathered organic soil horizons (Ah/Ahb) seem to be mainly coupled to a threshold in sedimentation rates along the Mamoré River, with increasing thickness away from river channel due to decreasing sedimentation rates. Therefore, they do not likely provide information on soil maturity, showing that time is not the dominant soil forming factor in these seasonal wetland settings (Bown and Kraus, 1987; Aslan and Autin, 1998).

5.3 Controls on the formation of Holocene floodplain paleosols

A range of processes potentially contribute to the variability in hydrological setting and sedimentation rate in floodplains, and can be related to changes in geomorphic and/or climatic boundary conditions (Fig. 9).

First, within the meanderbelt, dynamic lateral migration of the channel (Fig. 9) is characteristic for meandering systems such as the Beni and the Mamoré Rivers in the LM. Thus, the distance between the channel and any given point in the floodplain is changing over time in response to lateral migration, and could explain variations in sedimentation rates and hydrologic setting in the floodplain. Lateral migration rates of individual meander bends have been estimated for the Beni River to average 30 m a^{-1} with maxima of up to 140 m a^{-1} (Gautier et al., 2007), and are similarly dynamic along the Mamoré River (Charriere et al., 2004; Constantine et al., 2014). These rapid channel shifts, however, are mostly constricted to the meanderbelt width, and in addition occur on decadal timescales in contrast to the long-term changes in sedimentation rate as expressed by the few Ah/Ahb horizons in our profiles. Aside from lateral channel migration as related to the evolution of individual meander bends, meanderbelt systems can also be subject to longer-term shifts in the mean average position of the channel, e.g., through slow tectonic movements, eventually leading to meanderbelt asymmetry (Leeder and Alexander, 1987). A tectonically induced major reorganization in flooding dynamics and inundation patterns during the Holocene has recently been suggested for the LM (Lombardo, 2014) and should be reflected in a stratigraphic asymmetry between both sides of the river. Whilst meanderbelt asymmetry characterizes the Mamoré River in the northern LM, the channel position does not exhibit any obvious asymmetry for a long reach of several hundred km in the vicinity our studied profiles. In addition, periods of lower

sedimentation rates as inferred from the Ahb horizons occur on both sides of the meanderbelt and - based on our chronology - show some temporal overlap. Therefore, tectonics and resulting meanderbelt asymmetry alone are unlikely to explain the variation in sedimentation rates at our profiles and the related formation of soil horizons in the distal floodplain.

Second, changes in climate and environmental conditions may potentially trigger a range of changes within the fluvial system. (i) Environmental changes could influence the production and/or erosion rates of sediment in the Andean catchments of the Mamoré River, causing periods of reduced sediment supply along the lowland reach of the river (Fig. 9). Changes in sediment supply should affect the floodplain synchronously regardless of the hydrological setting, but this is not reflected by our chronological data. Along the Andean piedmont marked changes in sedimentary dynamics have been deduced from fluvial and alluvial sediments during the early to mid-Holocene (May et al., 2008a), and possibly coincide with environmental changes throughout the Central Andes characterized by decreasing lake levels (Abbott et al., 1997), increased fire frequencies (Mourguiart and Ledru, 2003; Urrego et al., 2009) and minimal extents of Andean ice caps (Buffen et al., 2009). Varying erosion rates leading to varying proportions of sediments from the upper catchment, however, are not reflected in the provenance and overall geochemical data of our profiles, but could be further examined through the application of cosmogenic nuclides (Wittmann et al., 2011). (ii) Longer-term precipitation and inundation patterns may vary in response to climatic changes e.g., ENSO (Aalto et al., 2003; Ronchail et al., 2005). During periods of increased precipitation in the lowlands vs. the upper catchments (e.g., during positive El Niño phases), flooding along clear-water rivers may inundate much of the floodplain, thus reducing the potential for the dispersal of sediment-laden white-water flood waters far into the floodplains. In this scenario, the low sedimentation rates in the distal floodplain would correspond to

increasing El Niño conditions. (iii) In addition to the spatial characteristics of flooding, climatic changes can also cause significant changes in magnitude and frequency of discharge, ultimately affecting overbank sedimentation rates in the floodplain. The few available data for the Mamoré River suggest significantly reduced discharge in the middle Holocene (Plotzki et al., 2013) which would correspond to the idea of generally decreased sediment transport capacities and deposition along the proximal Andean piedmont and megafans in the upper Río Grande catchments (May et al., 2008b). Longer periods of reduced flooding intensities should increase the chance for geomorphic stability with negligible sediment supply and noticeable weathering in relation with these horizons, which was not detected in our data. Also, in the scenario of climatically induced change in sedimentation rates the formation of Ah/Ahb horizons should be roughly synchronous throughout the LM. Even though a temporal overlap in the mid-Holocene is reflected in the radiocarbon dates from the Ahb horizons at M-4 and M2a, further stratigraphic and chronological data are required to test this scenario.

Third, changes in discharge and sediment supply over time can also be related to re-organizations in the drainage network as caused by channel avulsions and capture (e.g., Aslan and Autin, 1999), and may thus provide a mechanism to explain variations in regional scale changes in inundation pattern and deposition rates (Fig. 9). Today the Río Grande contributes the majority of sediment to the Mamoré River (Guyot et al., 1994; Constantine et al., 2014) and actively constructed a distributary channel network ~60 km southeast of the M-4 profile between ~7-4 ka (Lombardo et al., 2012). Interestingly, this coincides with comparatively high sedimentation rates and the deposition of more muddy layers at M-4 between ~7-4 cal ka BP. A difference in sediment provenance between M-4 and M-2b on the western margin of the Mamoré River may also be reflected in the Ti and K/Mg data (Plotzki et al., 2015). The

abandonment of the distributary system to the southeast, and the connection of the Río Grande with the Mamoré River system occurred sometime in the mid- to late Holocene (Lombardo et al., 2012; Plotzki et al., 2013). This may therefore explain the increase in K/Mg towards the top of both profiles, and could suggest that low sedimentation rates before ~6-7 ka as reflected by the Ahb horizon of roughly uniform thickness on both sides of the Mamoré River (Fig. 8a) may provide evidence for previously decreased hydrological and geomorphic dynamics along the reach of the current Mamoré River. Further data on the spatial and temporal evolution of the drainage network in the extensive Andean foreland basin upstream of our profiles will allow a more complete reconstruction of these large-scale geomorphic changes, and help to clarify their effects on sedimentation rates and soil formation in the LM.

6. Conclusions

The combined pedological, geochemical, mineralogical and chronological data from the three investigated profiles suggest a combination of geomorphic and climatic controls on their formation. The presence of thick overbank deposits of mostly Holocene age (unit 2) on both sides of the Mamoré River underlines the importance of extensive flooding processes for the late Quaternary sedimentary and geomorphic evolution of the LM. These fine-grained distal overbank deposits are strongly modified by well-developed redoximorphic features such as mottles and nodules and suggest a strong control of hydrological conditions related to temporal and spatial inundation pattern on weathering and post-depositional modification of the sediments. Spatial variations in redoximorphic features between our profiles also imply the presence of seasonal wetland catenas, possibly reflecting varying groundwater levels and flooding durations as a function of proximity to the Mamoré River channel. Widespread dark gray to blackish paleosols are intercalated with the overbank sediments. Despite their distinct

appearance they lack evidence for significant weathering and are likely the result of cumulative incorporation of organic material under conditions of particularly low sedimentation rates as typical in cumulic alluvial soils. Therefore, their paleoenvironmental and –climatic significance is not straightforward, and their formation may be linked to interactions between a variety of geomorphic processes within the dynamically evolving framework of the LM fluvial system. The Holocene floodplain paleosols along the Mamoré River may thus reflect the combined effect of hydrological, geomorphic and sedimentary changes related to overall drier mid-Holocene climatic conditions, but further geochemical analysis and dating efforts are needed to elucidate the regional validity of paleosol chronology, as well as the timing and impact of upstream re-organizations in the fluvial network (Lombardo, 2014; Plotzki et al., 2015). For this purpose, improved chronologies are required to enable a more accurate picture of varying sedimentation rates. In combination, more detailed investigations of the organic matter preserved in the widespread paleosol horizons offer prospects of studying climate related late Quaternary vegetation changes in the distal floodplain, and may even provide a first step required for the assessment of spatial and temporal variations in long-term carbon storage capacity in the inundation savannah of the southern Amazon basin. Our results imply that sedimentary and pedogenic aspects of seasonal flooding processes need to be considered with any future attempt of unravelling the longer-term Cenozoic stratigraphic record of foreland basin fills in southern Amazonia (Wright and Marriott, 1993) and interpreting distal floodplain deposits and paleosols in the geological record worldwide (Platt and Keller, 1992; Kraus and Aslan, 1993; Kraus, 1997; Wright, 1999; Wright et al., 2000).

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Figures

Fig. 1. Geographical setting of the Llanos de Moxos. Insets show location within South America and Bolivia, respectively. Blue shading shows the inundation area in the Llanos de Moxos based on Brakenridge et al. (2014).

Fig. 2. Geographic setting and field impressions from the studied floodplain paleosol-sediment sequences along the Mamoré River cut-banks. a) Natural colour MODIS Aqua satellite image (17-02-2014) showing the dispersal of sediment plumes from the major white-water rivers into the inundated floodplain southwest of Trinidad (1 - Mamoré River, 2 - Securé River, 3 and 4 - Isiboró and Tijamuchí Rivers). b) True colour (band combination 1-4-3) Landsat 8 image (19-02-2014) illustrating the setting of profiles M-2a and M-2b with regard to the overbank sediment plume (brown colour) extending ~4 km from the channel (blue dashed line) into the inundated floodplain (black). c) Paleosol-sediment-sequences at M-2a (Trinidad-1); note the presence of a gallery forest on top of the river bank. d) Grassland vegetation in the inundation savannah at coring site M-2b approximately 3.5 km southwest of M-2a. e) Bank exposure at M-4 (San Carlos). Arrows mark the position of paleosol horizons. f) Red and ochre matrix of II-Cg horizon (> 4 m depth) with light brown and grey mottles and Fe-Mn nodules section M-2b. g) Grey-coloured silty and slightly sandy loam with yellowish red to strong brown mottles in Cg horizon (> 4 m depth) in M-4. h) Dark grey Ahb1 horizon (0.7 to 1.1 m depth) with reddish mottles in section M-4.

Fig. 3. Stratigraphic, sedimentary and field pedological illustration of the floodplain paleosol-sediment sequences at San Carlos (M-4) and Trinidad (M-2a and b) including the results from radiocarbon and OSL dating.

Fig. 4. Down-profile variations of selected textural and geochemical parameters (see Fig. 3 for legend); white circles mark locations of sample points.

Fig. 5. Summary of clay mineralogical composition of selected samples at M-2a and M-4 as compared to Mamoré River parent sediment reported in Guyot et al. (2007) and Guyot (1993).

Fig. 6. Relationships between selected textural and geochemical variables. White dots mark sample depth in the sedimentary core M-2b and profile M-4.

Fig. 7. Deposition rates in the Mamoré River floodplain as calculated from selected radiocarbon and OSL ages (this study) and ^{210}Pb chronology (Aalto et al., 2003).

Fig. 8. Stratigraphic and soil geomorphic model in the central Llanos de Moxos. a) Conceptual illustration of the location of the studied profiles within a wetland catena along the Mamoré River. b) Contrasting soil geomorphic relationship in the southeastern Llanos de Moxos where the presence of Pleistocene and Holocene levees creates topographic relief, and the hydrological setting may have been artificially altered (Lombardo et al., 2015).

Fig. 9. Conceptual diagram illustrating potential controls on sedimentation rates and the formation of floodplain soils along the Mamoré River in the LM, Bolivia (1.1 dynamic lateral migration of the channel; 1.2 longer-term shifts in the meanderbelt; 2.1 changes in sediment production and/or erosion rates in the Andean catchments; 2.2 spatial variations of inundation patterns in response to longer-term precipitation changes; 2.3 climatically induced changes in flooding magnitude and frequency; 3 drainage network re-organization caused by channel avulsions and capture).

Supplementary Data Fig. 1. X-Ray diffractograms from profile M-2a and M-4 with arrows indicating the identification of goethite and lepidocrocite in the samples (x, y and z indicate measurements on air-dried, ethylene glycol saturated, and heated (550°C) samples, respectively).

Tables:

Table 1: Stratigraphy, soil horizon and morphology of sites M-2a, b and M-4.

Depth (cm)	Horizon	Upper boundary ^a	Texture ^b	Colour matrix	Colour mottles	Structure ^c	Other features
M-2a (Trinidad-1)							
0-20	(C)	-	MS	light gray (10YR 8/2)	-	SG	-
20-40	Ah(b)	c	HC	dull yellowish brown (10YR 5/3)	-	AB	-
40-180	Cg	g	HC	brownish gray - grayish brown and dull reddish brown (10YR 4/1, 5YR 4/2 - 5/3)	reddish brown (2.5YR 4/8)	AB	-
180-260	Ahb	c	HC	brownish black - grayish brown (5YR 3/1 - 4/2)	reddish brown (2.5YR 4/8)	MA	gypsum crystals
> 260	IICg	g	HC SiC	brownish gray - grayish brown, and (5YR 5/1 - 6/2)	reddish brown and dull yellow orange (2.5YR 4/8, 10YR 6/3 - 6/4)	MA	black Fe-Mn concretions, large mottles
M-2b (Laguna Seca)							
0-20	(C)-Ah	-	HC	yellowish gray - grayish yellow (2.5Y 6/1 - 6/2)	-	(MA)	-
20-150	Ah	g	HC SiC	dark grayish yellow - grayish yellow (2.5Y 4/2 - 6/2)	-	(AB)	-
150-200	Cg	g	HC	grayish yellow - dull yellow (2.5Y 6/2 - 6/4)	yellowish brown (10YR 5/6)	(MA)	-
200-260	Ahb1	c	HC	dark greyish yellow (2.5Y 5/2)	-	(MA)	-
260-270	C	g	HC	dull brown (7.5YR 6/3)	-	(MA)	-
270-295	Ahb2	c	HC	dark grayish yellow and grayish brown - light brownish gray (2.5Y 5/2, 7.5YR 6/2 - 7/1)	reddish brown (2.5YR 4/8)	(MA)	-
295-450	Cg	g	HC	dull brown - dull orange (7.5YR 6/3 - 7/3)	reddish brown (2.5YR 4/8)	(MA)	-
450-500	IICg	g	CL	grayish yellow - light gray, and dull orange (2.5Y 7/2 - 8/1, 7.5YR 6/4)	dull yellow orange (10YR 6/3 - 6/4)	(MA)	large mottles
>500	IICg	g	L SL	orange (7.5YR 7/6)	bright yellowish brown (10YR 6/6)	(MA)	large mottles
M-4 (San Pedro)							
0-40	Ah	-	HC	grayish yellow brown (10YR 4/2)	-	AB	-
40-70	Cg	g	HC	dull yellowish brown (10YR 4/3)	reddish brown (2.5YR 4/8)	AB	-
70-110	Ahb1	c	HC	brownish gray (7.5 YR 4/1)	reddish brown (2.5YR 4/8)	AB	gypsum crystals
110-300	Cg	g	HC SiC	dull yellowish brown (10 YR 5/3)	reddish brown - bright reddish brown (2.5YR 4/8 and 5YR 5/8)	AB	black Fe-Mn concretions
300-400	Ahb2	c	HC	black - brownish gray (7.5 YR 2.5 -4/1)	reddish brown (5YR 4/6)	MA	gypsum crystals
> 400	Cg - IICg	c	HC SiL (clay-rich) SL (clay-rich)	brownish gray (10YR 6/1)	reddish brown - bright brown (5YR 4/6 and 7.5 YR 5/8)	MA	

^a **Boundary:** g = gradual, c = clear.
^b **Texture:** HC = heavy clay, C = clay, SiC = silty clay, L = loam, CL = clay loam, SiL = silt loam, SL = sandy loam, Si = silt, MS = medium sand, LS = loamy sand.
^c **Structure:** AB = angular blocky, MA = massive, SG = single grain (according to FAO, 2006).

Table 2: Results from radiocarbon dating (x denotes samples used for calculating sedimentation rates)

Site	Latitude	Longitude	Field Code	Lab Code	Material	Dated fraction	Depth (cm)	¹⁴ C (a) BP / pMC	Cal a BP			Cal a BP**	Rates
M-2a	-14.839	65.062	1_11	Poz-34217	Soil / Organic Sediment	Humins (RES)	200	5560±40	6282	-	6394	6338±56	x
M-2a	-14.839	65.062	1_11	Poz-39531	Soil / Organic Sediment	Humic Acids (NaOH)	200	5600±40	6303	-	6396	6350±47	
M-2a	-14.839	65.062	1_9	Poz-34216	Soil / Organic Sediment	Humins (RES)	237	8290±50	9125	-	9393	9259±134	x
M-2a	-14.839	65.062	1_9	Poz-39532	Soil / Organic Sediment	Humic Acids (NaOH)	237	7440±50	8175	-	8311	8243±68	
M-2b	-14.862	65.076	B1 II 80-85	Poz-16786	Soil / Organic Sediment	Humic Acids (NaOH)	180	7960±50	8631	-	8968	8800±169	
M-2b	-14.862	65.076	B1 II 80-85	Poz-16787	Soil / Organic Sediment	Humins (RES)	180	11970±60	13605	-	13823	13714±109	
M-2b	-14.862	65.076	B1-III-37	Poz-13524	Charcoal	Charcoal	237	108.63±0.46*	modern			-	
M-2b	-14.862	65.076	B1 IV 45-50	Poz-16788	Soil / Organic Sediment	Humic Acids (NaOH)	345	9280±60	10296	-	10497	10397±101	
M-2b	-14.862	65.076	B1 IV 45-50	Poz-16789	Soil / Organic Sediment	Humins (RES)	345	9220±50	10249	-	10397	10323±74	x
M-2b	-14.862	65.076	B1 V 50-55	Poz-16790	Plant Remains	TOC	450	106.77±0.33*	modern			-	
M-2b	-14.862	65.076	B1-VII-98	Poz-13525	Charcoal	Charcoal	698	10180±310	11258	-	12381	11820±562	x
M-2b	-14.862	65.076	B1-VIII-60	Poz-13526	Charcoal	Charcoal	760	10440±60	12097	-	12413	12255±158	x
M-4	-14.348	64.964	169_A	Poz-39577	Soil / Organic Sediment	Humins (RES)	75	2510±30	2466	-	2701	2584±118	x
M-4	-14.348	64.964	169_A	Poz-39612	Soil / Organic Sediment	Humic Acids (NaOH)	75	2945±35	2963	-	3137	3050±87	
M-4	-14.348	64.964	169_B	Poz-39578	Soil / Organic Sediment	Humins (RES)	110	3435±35	3585	-	3690	3638±53	x
M-4	-14.348	64.964	169_B	Poz-39613	Soil / Organic Sediment	Humic Acids (NaOH)	110	4140±50	4523	-	4806	4665±142	
M-4	-14.348	64.964	169_C	Poz-39579	Soil / Organic Sediment	Humins (RES)	305	6470±40	7321	-	7417	7369±48	x
M-4	-14.348	64.964	169_C	Poz-39614	Soil / Organic Sediment	Humic Acids (NaOH)	305	7200±60	7879	-	8028	7954±75	
M-4	-14.348	64.964	169_D	Poz-39580	Soil / Organic Sediment	Humins (RES)	320	7670±40	8382	-	8452	8417±35	x
M-4	-14.348	64.964	169_D	Poz-39616	Soil / Organic Sediment	Humic Acids (NaOH)	320	7160±50	7872	-	7998	7935±63	
M-4	-14.348	64.964	169_E	Poz-39581	Soil / Organic Sediment	Humins (RES)	400	11130±60	12872	-	13045	12959±87	x
M-4	-14.348	64.964	169_E	Poz-39617	Soil / Organic Sediment	Humic Acids (NaOH)	400	10060±60	11345	-	11691	11518±174	

* pMC

** Mean

Table 3: Results from OSL dating.

Site	Sample	Depth (cm)	Grain size (µm)	n	od	K (%)	Th (ppm)	U (ppm)	Mois. (%)	W max. (%)	W (%)	D (Gy ka ⁻¹)	D _E CAM (Gy)	D _E MAM (Gy)	Age CAM (ka)	Age MAM (ka)
M-4	AP5f	60	4-11	5	-	2.71 ± 0.04	19.62 ± 0.82	3.73 ± 0.10	10	34	22 ± 10	4.78 ± 0.60	14.05 ± 0.18	-	2.94 ± 0.37	-
M-4	AP6f	300	4-11	5	-	2.85 ± 0.04	18.53 ± 0.89	3.59 ± 0.05	16	21	18 ± 3	4.90 ± 0.35	25.21 ± 0.37	-	5.14 ± 0.37	-
M-4	AP7f	410	4-11	5	-	1.99 ± 0.03	17.29 ± 0.48	3.31 ± 0.05	16	16	16 ± 2	4.09 ± 0.29	64.67 ± 1.07	-	15.8 ± 1.16	-
M-4	AP1f	590	4-11	6	-	1.26 ± 0.02	11.30 ± 0.53	2.51 ± 0.03	10	12	11 ± 2	2.89 ± 0.20	107.96 ± 2.82	-	37.4 ± 2.8	-
M-4	AP1c	590	100-150	50	0.27	1.26 ± 0.02	11.30 ± 0.53	2.51 ± 0.03	10	12	11 ± 2	2.49 ± 0.11	53.45 ± 2.13	44.23 ± 5.88	21.5 ± 1.3	17.8 ± 2.5

Supplementary Data Tables:

Table 1: Results from dose recovery and thermal transfer test on coarse grain sample AP1c

Profile	Sample	n	Given dose (Gy)	Recovered dose (Gy)	RSD (%)	Recovery ratio	Thermal transfer (Gy)	Thermal transfer (%)
M-4	AP1c	4	49.15	53.57 ± 5.63	10.5	1.09	-0.41 ± 0.50	-0.8

Table 2: Results from grain size analysis, C and N analysis, and XRF analysis for profiles M-2b and M-4.

Sample Nr	Depth cm	f. Sand %	Silt %	Clay %	N %	C %	C/N	LOI %	SiO2 %	TiO2 %	Al2O3 %	Fe2O3 %	MnO %	MgO %	CaO %	K2O %	Na2O %	P2O5 %	S %	Cl %	Ni %	Cr %	V %	Co %	Cu %	Zn %	Pb %	Ba %	Rb %	Sr %	Y %	Nb %	Zr %	Ga %	Mo %	Cs %	As %	Ce %	Nd %	Hf %	CIA
M-2b-1	2.5	0.40	19.72	79.88	0.28	2.45	8.66	15.00	59.140	0.571	25.170	4.140	0.013	1.370	0.306	3.780	0.351	0.163	0.083	0.016	0.002	0.010	0.025	0.004	0.005	0.013	0.010	0.101	0.028	0.016	0.005	0.002	0.023	0.003	0.000	0.013	0.000	0.024	0.009	0.000	85.01
M-2b-2	12.5	1.47	27.79	70.75	0.17	0.98	5.98	16.83	58.680	0.673	23.500	7.800	0.015	1.220	0.329	3.310	0.309	0.114	0.054	0.005	0.003	0.011	0.023	0.007	0.004	0.013	0.007	0.097	0.023	0.015	0.005	0.003	0.030	0.003	0.003	0.009	0.000	0.024	0.010	0.000	85.62
M-2b-3	17.5	1.86	29.05	69.09	0.16	1.02	6.39	14.45	59.910	0.682	23.970	6.050	0.016	1.250	0.354	3.390	0.343	0.107	0.051	0.002	0.004	0.011	0.022	0.006	0.003	0.013	0.009	0.099	0.024	0.015	0.005	0.003	0.032	0.003	0.000	0.013	0.000	0.014	0.010	0.003	85.43
M-2b-4	22.5	2.63	25.90	71.47	0.16	1.07	6.86	18.07	59.110	0.620	24.640	6.190	0.021	1.260	0.432	3.280	0.297	0.108	0.049	0.008	0.003	0.009	0.025	0.007	0.006	0.013	0.008	0.102	0.023	0.015	0.005	0.002	0.030	0.003	0.003	0.013	0.000	0.018	0.007	0.000	86.01
M-2b-5	27.5	1.15	20.36	78.49	0.20	1.45	7.28	27.53	57.390	0.569	26.000	6.170	0.033	1.320	0.531	3.280	0.253	0.103	0.050	0.004	0.005	0.010	0.026	0.007	0.007	0.014	0.009	0.103	0.025	0.015	0.005	0.002	0.025	0.004	0.000	0.009	0.000	0.016	0.010	0.002	86.48
M-2b-6	52.5	0.10	35.57	64.34	0.26	1.67	6.43	13.42	61.450	0.575	23.930	4.520	0.012	1.150	0.554	3.050	0.280	0.083	0.033	0.005	0.005	0.011	0.024	0.005	0.006	0.011	0.007	0.112	0.023	0.013	0.005	0.002	0.023	0.003	0.003	0.015	0.000	0.025	0.011	0.000	86.04
M-2b-7	72.5	0.47	43.40	56.14	0.17	1.06	6.20	11.26	63.760	0.582	22.860	4.070	0.008	1.140	0.482	3.100	0.369	0.080	0.023	0.006	0.002	0.023	0.022	0.004	0.005	0.010	0.005	0.113	0.021	0.012	0.004	0.002	0.026	0.003	0.000	0.013	0.000	0.017	0.006	0.002	85.26
M-2b-8	92.5	0.17	41.13	58.70	0.16	0.91	5.59	13.72	62.800	0.579	22.770	5.190	0.009	1.170	0.501	3.210	0.345	0.085	0.016	0.004	0.002	0.010	0.022	0.004	0.003	0.011	0.005	0.100	0.022	0.013	0.004	0.002	0.024	0.003	0.000	0.013	0.000	0.018	0.010	0.000	84.88
M-2b-9	127.5	0.39	23.61	76.00	0.12	0.66	5.71	11.80	61.260	0.515	22.600	6.920	0.016	1.290	0.616	2.770	0.253	0.165	0.010	0.006	0.003	0.009	0.020	0.006	0.004	0.013	0.007	0.094	0.022	0.014	0.004	0.002	0.023	0.003	0.000	0.008	0.000	0.014	0.002	0.000	86.13
M-2b-10	142.5	0.75	18.36	80.89	0.13	0.70	5.21	17.89	57.910	0.514	23.100	9.230	0.019	1.290	0.621	3.080	0.260	0.236	0.010	0.006	0.003	0.010	0.022	0.008	0.005	0.016	0.008	0.101	0.024	0.015	0.005	0.002	0.020	0.003	0.000	0.009	0.002	0.017	0.010	0.000	85.36
M-2b-11	152.5	2.12	17.13	80.75	0.10	0.46	4.44	15.71	56.720	0.523	24.210	9.030	0.019	1.360	0.563	3.370	0.285	0.273	0.011	0.005	0.003	0.011	0.020	0.008	0.002	0.015	0.007	0.110	0.024	0.015	0.004	0.002	0.019	0.004	0.000	0.011	0.000	0.024	0.006	0.000	85.16
M-2b-12	167.5	4.60	26.33	69.07	0.11	0.42	3.85	9.37	54.120	0.555	23.240	12.260	0.016	1.370	0.467	3.420	0.324	0.477	0.012	0.004	0.004	0.010	0.022	0.012	0.003	0.017	0.007	0.105	0.023	0.014	0.004	0.002	0.018	0.003	0.002	0.008	0.002	0.016	0.011	0.000	84.66
M-2b-13	182.5	4.22	29.69	66.09	0.11	0.43	3.88	9.32	54.040	0.542	23.030	12.810	0.014	1.350	0.466	3.260	0.322	0.506	0.014	0.006	0.000	0.011	0.022	0.006	0.000	0.018	0.006	0.099	0.022	0.013	0.004	0.002	0.019	0.003	0.002	0.012	0.000	0.020	0.009	0.000	85.05
M-2b-14	242.5	0.39	16.44	83.18	0.13	1.00	7.68	14.30	60.320	0.491	23.100	7.330	0.012	1.260	0.705	2.570	0.209	0.114	0.012	0.007	0.003	0.010	0.023	0.007	0.008	0.018	0.007	0.101	0.024	0.015	0.005	0.002	0.019	0.003	0.000	0.010	0.002	0.021	0.009	0.000	86.89
M-2b-15	267.5	2.21	15.45	82.34	0.12	0.50	4.19	8.66	58.940	0.503	24.870	6.510	0.013	1.420	0.677	2.860	0.203	0.088	0.009	0.000	0.002	0.009	0.024	0.006	0.005	0.015	0.007	0.097	0.023	0.016	0.004	0.002	0.021	0.004	0.000	0.007	0.003	0.017	0.005	0.000	86.93
M-2b-16	272.5	1.28	17.37	81.35	0.13	0.70	5.55	17.71	60.120	0.535	24.360	5.790	0.015	1.350	0.714	2.910	0.194	0.080	0.013	0.004	0.002	0.011	0.023	0.006	0.007	0.015	0.009	0.116	0.026	0.017	0.005	0.002	0.024	0.004	0.000	0.010	0.000	0.024	0.015	0.000	86.45
M-2b-17	277.5	0.68	15.20	84.12	0.14	0.83	5.88	17.66	59.130	0.507	24.410	6.930	0.014	1.340	0.697	2.780	0.202	0.086	0.014	0.009	0.002	0.013	0.023	0.006	0.006	0.014	0.008	0.111	0.024	0.016	0.005	0.003	0.022	0.003	0.002	0.016	0.000	0.022	0.009	0.000	86.90
M-2b-18	282.5	0.88	15.17	83.96	0.12	0.68	5.63	14.55	59.340	0.512	24.890	6.060	0.013	1.390	0.713	2.820	0.193	0.080	0.012	0.005	0.003	0.014	0.024	0.006	0.006	0.015	0.008	0.109	0.024	0.017	0.005	0.002	0.021	0.003	0.002	0.013	0.000	0.014	0.013	0.000	86.98
M-2b-19	287.5	1.03	15.54	83.43	0.16	0.64	4.00	14.79	58.560	0.528	24.310	7.550	0.012	1.360	0.711	2.900	0.208	0.095	0.013	0.007	0.003	0.017	0.024	0.006	0.006	0.017	0.009	0.109	0.025	0.017	0.005	0.002	0.023	0.004	0.003	0.009	0.000	0.014	0.014	0.003	86.42
M-2b-20	292.5	2.60	15.78	81.62	0.12	0.66	5.41	13.29	59.150	0.522	24.940	6.450	0.015	1.370	0.668	2.900	0.176	0.072	0.014	0.011	0.003	0.010	0.022	0.006	0.005	0.014	0.008	0.108	0.024	0.016	0.005	0.002	0.023	0.004	0.001	0.014	0.000	0.021	0.009	0.002	86.95
M-2b-21	297.5	3.97	16.90	79.13	0.12	0.38	3.26	9.20	58.220	0.507	24.910	7.680	0.012	1.430	0.588	2.960	0.197	0.077	0.014	0.003	0.002	0.009	0.020	0.006	0.003	0.013	0.006	0.091	0.023	0.014	0.003	0.002	0.025	0.004	0.000	0.008	0.000	0.014	0.009	0.003	86.93
M-2b-22	352.5	3.28	18.67	78.06	0.11	0.48	4.35	7.74	58.540	0.506	24.950	7.280	0.013	1.4																											

Table 3: Clay Mineralogy of selected samples from M-4 and M-2a.

Name	Depth (cm)	Illite (%)	Kaolinite (%)	Smectite (%)	Ill./Smec. (%)	Chlorite (%)	Horizon
M4-2	20	62	28	-	10	0	Ah
M4-4	60	41	53	-	6	0	Ah
M4-6	100	48	46	-	6	0	C
M4-8	140	53	41	-	5	0	C
M4-10	180	53	41	-	6	0	C
M4-12	220	49	43	-	7	0	C
M4-14	260	46	47	-	7	0	C
M4-16	300	36	48	16	-	0	fAh
M4-18	340	33	49	18	-	0	fAh
M4-20	380	32	49	20	-	0	fAh
M4-23	575	38	52	10	-	0	C
M4-24	625	40	47	13	-	0	C
M2-2	42	53	42	4	-	1	Ah
M2-3	62	35	62	1	-	1	C
M2-5	150	52	43	4	-	1	C
M2-6	203	38	54	8	-	1	fAh
M2-7	238	32	56	6	-	1	fAh
M2-8	312	32	58	9	-	1	C
M2-9	387	29	57	13	-	1	C
M2-10	428	37	57	5	-	1	C
M2-11	467	49	44	6	-	1	C
M2-12	507	53	41	5	-	1	C