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Lacustrine records of Holocene flood pulse dynamics in the Upper Paraguay River watershed (Pantanal wetlands, Brazil)

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ABSTRACT

The Pantanal is the world's largest tropical wetland and a biodiversity hotspot, yet its response to Quaternary environmental change is unclear. To address this problem, sediment cores from shallow lakes connected to the Upper Paraguay River (PR) were analyzed and radiocarbon dated to track changes in sedimentary environments. Stratal relations, detrital particle size, multiple biogeochemical indicators, and sponge spicules suggest fluctuating lake-level lowstand conditions between ~11,000 and 5300 cal yr BP, punctuated by sporadic and in some cases erosive flood flows. A hiatus has been recorded from ~5300 to 2600 cal yr BP, spurred by confinement of the PR within its channel during an episode of profound regional drought. Sustained PR flooding caused a transgression after ~2600 cal yr BP, with lake-level highstand conditions appearing during the Little Ice Age. Holocene PR flood pulse dynamics are best explained by variability in effective precipitation, likely driven by insolation and tropical sea-surface temperature gradients. Our results provide novel support for hypotheses on: (1) stratigraphic discontinuity of floodplain sedimentary archives; (2) late Holocene methane flux from Southern Hemisphere wetlands; and (3) pre-colonial indigenous ceramics traditions in western Brazil.

Introduction

Wetlands have long received recognition for their role in global biogeochemical cycles, their value as a waterfowl habitat and as a freshwater resource for developing societies (Davis, 1994; Wuebbles and Hayhoe, 2002; Zedler and Kercher, 2005). Over the past century, most wetland conservation issues have focused on threats whose impacts are felt largely by local stakeholders (e.g., fragmentation, deforestation, invasive species, and pollution). In contrast, climate change presents a threat to wetland services on a much broader scale. Tropical wetlands, for example, are known to be sensitive to perturbations in air temperature or in the hydrologic cycle (Chauhan and Gopal, 2001). Based on IPCC AR4 projections, these ecotones are at substantial risk for terrestrialization, which may in turn have significant implications for carbon cycling and the Earth's energy balance (Bates et al., 2008). Modeling studies argue for increased CH_4 flux to the atmosphere from tropical wetlands under warming scenarios, potentially inducing a positive feedback on global climate (Shindell et al., 2004). Such warming also impacts aquatic thermal regimes that could result in widespread loss of biodiversity through anoxic events and habitat degradation (Hamilton, 2010). Other studies stress the potential for oxidation of carbon formerly sequestered in flooded soils (Mitsch et al., 2009). However, accurate model predictions of the impacts of global change on terrestrial ecosystems are a challenge due to the complex interactions among hydrology, vegetation, nutrients, soils, and fungi at many scales (e.g., Heimann and Reichstein, 2008). Unfortunately, few in-depth paleo-records are available to assess the response of unaltered tropical wetlands to well-known climatic perturbations of the Quaternary and thus validate

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model simulations. Our central purpose with this study is to expand the development of such records in the Pantanal, an enigmatic wetland wilderness mostly located in western Brazil (Fig. 1).

The Pantanal is the world's largest savanna floodplain-type wetland and a biodiversity hotspot (Junk et al., 2006; Lopes et al., 2007). The Pantanal owes its existence to seasonal flooding of the Upper Paraguay River (PR), which inundates the region with a slow-moving flood pulse produced by austral summer rainfall (Junk et al., 1989). The arrival of flood waters from the PR and its tributaries is critical for: (1) primary productivity and nutrient cycling on floodplains (Hamilton et al., 1997; de Oliveira and Calheiros, 2000); (2) the life cycles of birds and fish, which are important to the regional economy (PRODEAGRO, 1997; Súarez et al., 2004; Alho, 2008); (3) the community structure of flora (Nunes da Cunha and Junk, 2001); and (4) human activities including ranching, agriculture, and riverine navigation (Hamilton, 1999; da Silva and Girard, 2004). The PR is unregulated by dams within the Pantanal and hydrographs show that the flood pulse is sensitive to drought (da Silva and Girard, 2004). As a result, deposystems directly influenced by PR flooding provide an attractive opportunity to assess the impact of environmental change on the river and indeed, the greater Pantanal wetland itself.

Aspects of the Quaternary history of western Brazil have been studied using a variety of approaches, including: (1) floodplain geomorphology (Tricart, 1982; Assine and Soares, 2004); (2) alluvial and loess stratigraphy (Stevaux, 2000); (3) carbonate analyses (Boggiani and Coimbra, 1995; Bertaux et al., 2002); (4) palynology (Ledru, 1993; Whitney et al., 2011); (5) sponge paleoecology (Parolin et al., 2007); and (6) soil and sediment geochemistry (Victoria et al., 1995; de Oliveira Bezerra and Mozeto, 2008). In his regional synthesis of the Upper Parana River watershed, Stevaux (2000) interpreted two distinct intervals of aridity (~40,000–8000 and ~3500–1500 cal yr BP) separated by a humid phase from ~8000 to 3500 cal yr BP. Accumulations of organic carbon in several lakes of the Pantanal at ~6500 cal yr BP have been interpreted as further evidence of a middle Holocene climatic optimum (de Oliveira Bezerra and Mozeto, 2008). New data from Whitney et al. (2011) support Holocene wetting, but they also concluded that drought affected the Pantanal from ~10,000 to 3000 cal yr BP, similar to other sites in tropical South America apparently responding to variability in insolation and Atlantic sea surface temperature (SST; Seltzer et al., 2000; Sifeddine et al., 2001; Abbott et al., 2003). The lack of coherence among the available records demonstrates that more research is needed to clarify the Pantanal's environmental history.

Here, we report the results of sediment core analyses from two shallow lakes in the central Pantanal (Fig. 1). The water levels in both of these basins are dominantly controlled by the PR flood pulse (McGlue et al., 2011). As a consequence, sedimentological, geochemical, and biological lake level indicators preserved in cores provide novel insights on Holocene flood pulse dynamics and enable comparisons with the growing paleo-database for the Pantanal wetlands.

Setting

The Pantanal (Portuguese for "swamp") extends from ~lat 16° to 21°S in a low-altitude basin that occupies portions of Brazil, Bolivia, and Paraguay (Fig. 2A). The basin is elliptically shaped and seismically active, and its formation has been linked to Andean tectonism (Horton and DeCelles, 1997; Ussami et al., 1999). The PR flows from north to south near the western margin of the basin (Fig. 2A). Vegetation in the Pantanal is a mixture of *cerrado* (tropical savanna), Amazon-derived semi-deciduous forest, Chaco-derived seasonal dry

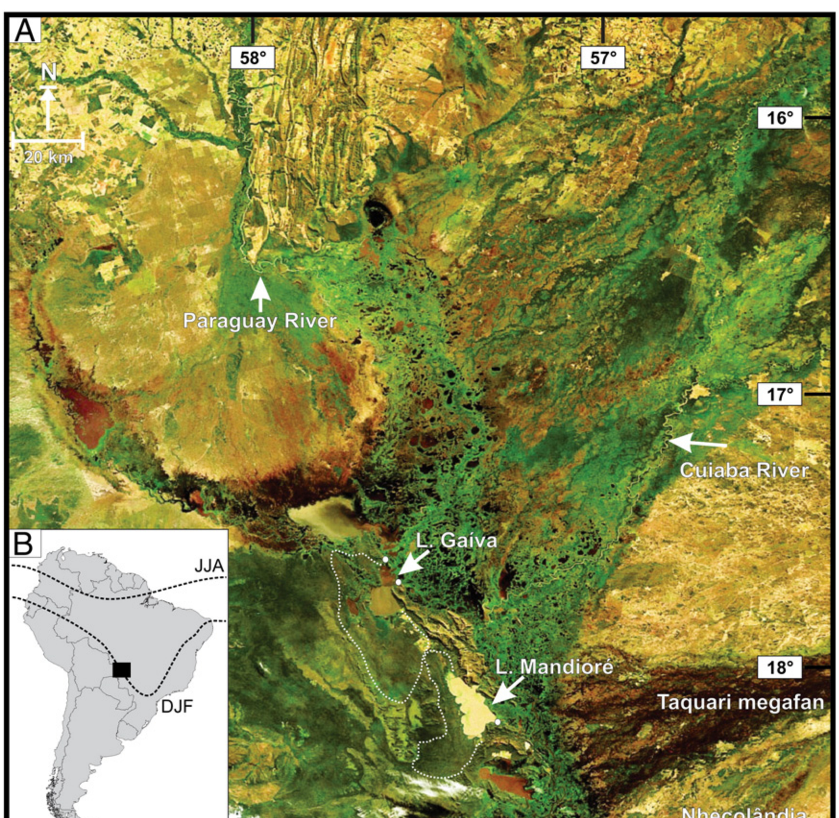

Figure 1. (A) Satellite image of the central Pantanal, Brazil and Bolivia (publicly available from CHELYS SRL, http://www.eosnap.com/tag/pantanal). Vegetation appears in green and water appears in shades of gray and blue. Lagoas Gaíva and Mandioré (western watershed limits are marked by dashed white lines) are situated on the western margin of the Paraguay River (PR), adjacent to the Serra do Amolar. The PR flood pulse travels from north to south, entering the lakes at locations marked by white dots. (B) Inset map of South America, with the Pantanal marked by a black box. Migration pathway of the Intertropical Convergence Zone is marked by stippled lines. Summer monsoon rainfall initiates the PR flood pulse, which controls limnogeological processes in the study lakes. DJF = December, January, and February. JJA = June, July, and August.

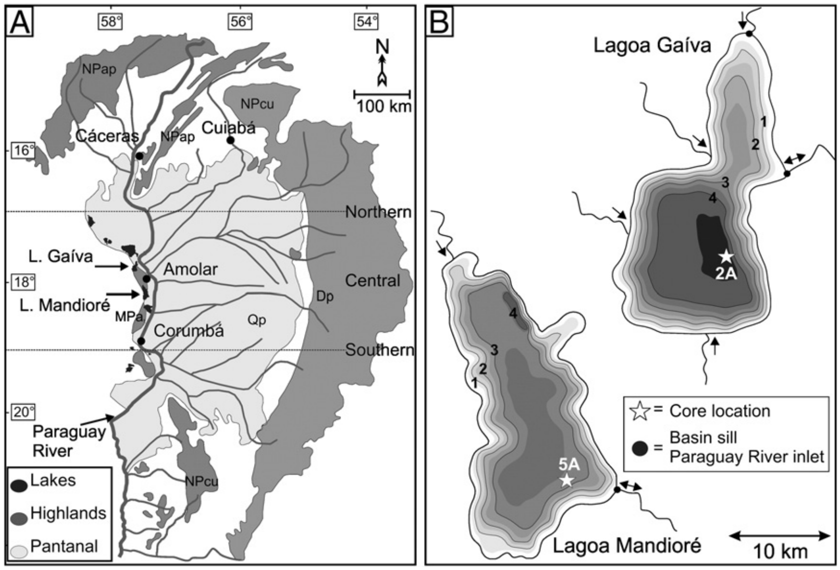

Figure 2. (A) Overview map of the Pantanal wetlands showing the locations of Lagoas Gaíva and Mandioré with respect to the cities of Corumbá and Cáceres (adapted from Assine and Silva, 2009). Dp = Devonian Paraná Group (sandstones, shales). MPa = Mesoproterozoic Amolar Group (metamorphosed sandstones, conglomerates). NPap = Neoproterozoic Alto Paraguai Group (mixed sedimentary units). NPcu = Neoproterozoic Cuiabá Group (metamorphosed sedimentary units). Qp = Quaternary Pantanal Formation (alluvium). (B) Locations of sediment cores used in this study. Bathymetric contour maps (in meters) are from McGlue et al. (2011).

forest, and aquatic plants (Prance and Schaller, 1982). The spatial distribution of plant communities is controlled by patterns of flooding, topography, and soil type (Pinder and Rosso, 1998). Several large rivers enter the Pantanal laterally and form megafans north and east of the PR (Assine and Silva, 2009). Lagoa Gaíva (LG) and Lagoa Mandioré (LM) sit adjacent to a Proterozoic mountain belt along the western margin of the PR, straddling the border of Brazil and Bolivia (CPRM, 2006). Both lakes are shallow (~4–6 m w.d.), dilute, well-oxygenated (probably the result of polymixis) and seasonally connected to the PR by channels (Fig. 2B; McGlue et al., 2011).

Climate in the Pantanal is strongly influenced by the South American summer monsoon (SASM; Zhou and Lau, 1998). Moisture derived from the Amazon basin is carried to the region by the South American low-level jet in the austral summer (Garreaud et al., 2009). The amount of summer precipitation in the Pantanal follows a strong N–S gradient, with areas in the north receiving ≥1500 mm. At the study sites (central Pantanal), >1000 mm of rain falls in a single prolonged wet season from late October to early April. The mean annual temperature is ~25°C and evaporation exceeds precipitation during most of the year (Por, 1995). Northwesterly winds characterize the austral summer, whereas east-northeasterly winds are prominent in the winter. Retention of the PR flood pulse in the northern Pantanal (through floodplain storage processes; Hamilton, 1999) delays the onset of full inundation in the central and southern Pantanal by several months. Lake levels track the arrival of PR flood waters and are less sensitive to direct precipitation early in the austral summer (McGlue et al., 2011).

Methods and materials

A short sediment core (core 2A, length 132 cm) was retrieved from LG using a modified square rod piston corer (Fig. 2B). Care was taken not to over-penetrate the lake floor and radioisotopes confirm that the modern sediment–water interface was collected (McGlue et al., 2011). An aluminum barrel was hammered into the southern end of LM in order to collect a 116-cm-long sediment core (core 5A). Although the primary focus of this study is LG core 2A, limited datasets from LM core 5A augment our interpretations of lake levels, especially for the late Holocene.

Following collection, the cores were shipped to the University of Minnesota-Limnological Research Center (UMN-LRC) where they were split, photographed, and described (Schnurrenberger et al., 2003). Ten sediment sub-samples were sieved to ~63 μm to isolate fine organic matter for radiocarbon dating. These sediments were subjected to an acid–base-acid sequential pre-treatment, followed by conversion to graphite at the University of Arizona (UA) Accelerator Mass Spectrometry facility or Beta Analytical Inc. Conventional ^{14}C dates were converted to calendar ages using CALIB 6.0 and the INTCAL09 curve (Table 1; Reimer et al., 2009). Radiocarbon data were used to: (1) identify hiatuses, (2) calculate sedimentation rates where deposition was relatively continuous, and (3) establish chronological context for paleoenvironmental interpretations. The lakes are hydrologically open for a portion of each year and the watershed geology is dominated by low-grade metamorphic rocks. Therefore, we assumed that the carbon inventory is in equilibrium with the atmosphere and any ^{14}C reservoir effects are minimal. Groundwater influence on the basins has not been quantified but is believed not to undermine age determinations.

Detrital particle size was measured on sub-samples (at 2-cm intervals) to assess depositional processes and energy using a Malvern Mastersizer ® with a Hydro 2000 aqueous dispersion unit. Samples were stirred at 3500 rpm and sonicated both during and prior to measurement to guarantee dispersion. Organic matter (OM), biogenic silica (BiSi) and any traces of carbonate were removed prior to analysis. Each sample was measured four times and results are reported as a mean value coarse (≥63 μm) to fine (<63 μm) ratio. Elemental and stable isotopic analyses of OM were conducted at UA to provide insights on biomass production, preservation, and provenance. Total organic carbon (TOC), total nitrogen (presented as atomic C:N) and $\delta^{13}C_{OM}$ were measured on a continuous-flow gas-ratio mass spectrometer (Finnigan Delta PlusXL ®) coupled to a Costech® elemental analyzer. Prior to combustion in the elemental analyzer, all samples were rinsed with 1 M HCl for 1 h. Standardization is based on acetanilide for elemental concentration and NBS-22 and USGS-24 for $\delta^{13}C_{OM}$. Precision was better than ± 0.09 for $\delta^{13}C_{OM}$ based on repeated internal standards. Biogenic silica determinations were completed at UMN-LRC using a time series wet chemical extraction technique (DeMaster, 1979). These data provide a measure of the diatom, sponge, and phytolith remains in sediments; reported values have a precision of ≤1.0%.

Table 1
Radiocarbon data for LG and LM. OM = sedimentary organic matter. Ages in parentheses were excluded from the age model.

Lab number	Sample name	Core depth (cm)	Material	$\delta^{13}C$ (‰)	Age (^{14}C yr BP)	Error	Median calibrated age (cal yr BP)	2-σ range (cal yr BP)
AA89103	LG09-2A-12	12	OM	−27.8	(555)	(56)	(558)	(518–601)
AA89098	LG09-2A-28	28	OM	−26.2	880	44	807	699–915
Beta-290022	LG09-2A-34	34	OM	−24.5	1510	40	1415	1313–1517
Beta-290210	LG09-2A-41	41	OM	−24.2	1830	30	1781	1698–1864
AA89099	LG09-2A-53	53	OM	−23.7	4596	56	5260	5051–5468
AA89100	LG09-2A-80	80	OM	−22.0	5422	57	6158	6005–6311
AA89101	LG09-2A-114	114	OM	−22.1	6282	61	7208	7005–7411
AA95519	LG09-2A-123	123	OM	−22.7	8012	93	8863	8600–9126
AA89110	LG09-2A-129	129	OM	−24.1	9708	54	11,018	10,798–11,238
AA89102	LG09-2A-131	131	OM	−25.4	(5458)	(94)	(6215)	(5996–6434)
AA87373	LM08-5A-110	110	Root	−27.0	4244	42	4749	4627–4871

Sponge spicules were analyzed at the Laboratório de Estudos Paleoambientais da Fecilcam (LEPAFE) following the procedure outlined in Volkmer-Ribeiro and Turcq (1996). Fifteen horizons (3–5 samples/horizon) from approximately equal intervals in the core were studied to identify megascleres, microscleres, and gemmoscleres. These spicules were used to characterize species assemblages and discriminate between lotic and lentic paleoenvironments (e.g., Parolin et al., 2008). Additionally, macroscopic charcoal, fish remains, and mineral grains with oxidation coatings were counted from residues of 125-μm screen-washed samples at a 2-cm sampling interval. Wet sediment samples were disaggregated in deionized water using repeated freeze–thaw cycles. Wet weights were measured for a separate aliquot from each sample, which was oven-dried and reweighed to determine water content and to calculate original dry weights for sieved samples. Residues were counted using a stereomicroscope and values are reported as abundance per gram of dry sediment.

The trace element composition of LG core 2A was determined at the University of Minnesota Duluth using an ITRAX® X-ray fluorescence core scanner operated at 1-cm resolution with 60 s scan times. This work focuses on Rb/K, an index of chemical weathering of watershed rocks, as K is preferentially removed in the hydrolysis of feldspar and mica. Because chemical weathering is a slow process, the paleolimnological significance of Rb/K over short time scales more likely relates to weathered material delivered by runoff to the basin.

Results

Geochronology

Eight of the ten radiocarbon measurements were used to determine the chronology of deposition at LG (Table 1). The shallowest sample (depth 12 cm) returned an older ^{14}C age than predicted based on the decay of excess ^{210}Pb (McGlue et al., 2011), possibly due to refractory carbon contamination. Additionally, the deepest sample (depth 131 cm) yielded a spurious age, due to low initial carbon. Both of these data points were excluded from further consideration. The age model that emerges is non-linear over the past ~11,000 cal yr BP (Fig. 3). From ~11,000 to ~6200 cal yr BP, our datasets suggest that crevasse flows entered LG from the north, punctuating the stratal record through erosion of the basin floor. In contrast, we interpret that there was a drought-related hiatus from ~5300 to 2600 cal yr BP, when the rate of sediment accumulation was negligible. Our age model suggests that sedimentation has been relatively continuous since ~2600 cal yr BP. From LM core 5A, a radiocarbon measurement made on a plant root found below a major unconformity surface yielded an age of ~4700 cal yr BP (Fig. 4A).

Stratigraphy and multi-indicator paleolimnology

LG core 2A was divided into five units based on macroscopic bedding features and microscopic components observed on smear slides.

Unit V (~11,000–8900 cal yr BP)

Unit V is comprised of massive gray–green silty clay that transitions into crudely bedded tan clayey sand near the upper contact (Fig. 5). These sediments are low in TOC, whereas $\delta^{13}C_{OM}$ is heavy and C/N reaches maximum values (>25; Fig. 5). Detrital coarse:fine and oxidized grains increase towards the top of the unit, suggesting an upward increase in environmental energy (Figs. 5 and 6). Sponge spicules are mainly from *Trochospongilla paulula* and *Oncosclera navicella*; these species favor lentic and lotic habitats, respectively (Fig. 7; Volkmer-Ribeiro, 1999; Volkmer-Ribeiro and Pauls, 2000). We interpret these indicators to reflect a lowstand lake that was influenced by PR floods. BiSi and TOC suggest that aquatic productivity was relatively low, which is supported by minor abundances of algae observed on smear slides. Values of $\delta^{13}C_{OM}$ and C/N indicate mixed organic parent material, likely including a component of terrestrial C_4 grasses (mean $\delta^{13}C$ of modern C_4 grass in the Pantanal is −12.4‰; Victoria et al., 1995) or macrophytes rich in ^{13}C.

Radiocarbon data bracket the timing of the transition between Units V and IV to ~8900–7200 cal yr BP. An irregular contact exhibiting high coarse:fine and oxidized grains separates these units (Figs. 5 and 6). Today, similarly coarse sediments are largely confined to the diffuse northern end of the lake, near the basin sill. We interpret these sediments as the distal expression of crevasse splay activity, most likely produced by relatively weak PR floods that occurred during foreshortened summer rains. Scour of underlying LG sediments from these flows punctuates the stratal record and helps to explain the non-linear age relations in the core during the early Holocene (Fig. 3).

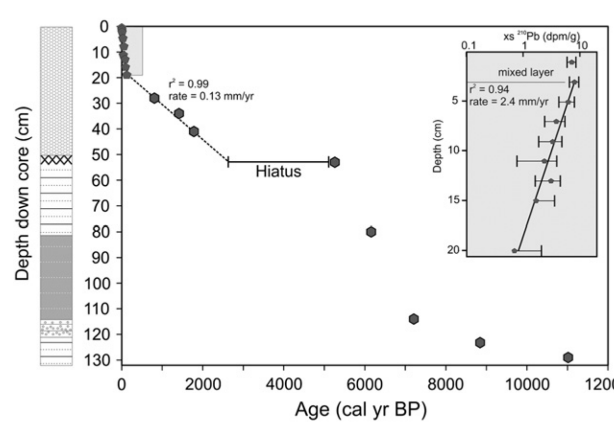

Figure 3. Radiocarbon age–depth model for Lagoa Gaíva core 2A. The gray inset box shows the ^{210}Pb age model for the upper ~20 cm of the core (McGlue et al., 2011). The stratal record is punctuated from ~11,000 to 6200 cal yr BP, most likely because of erosive flows that entered the basin associated with sporadic Paraguay River flooding. A hiatus from ~5300 to 2600 cal yr BP is marked by a major reduction in sedimentation rate. See text for details.

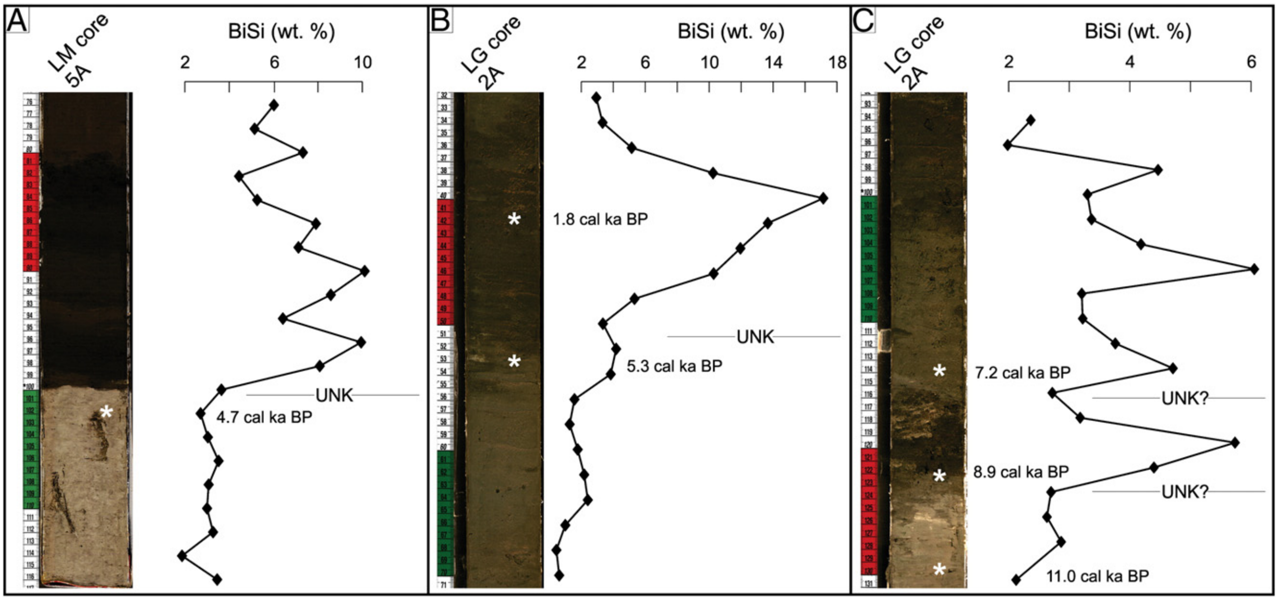

Figure 4. Core photo and biogenic silica data from: (A) Lagoa Mandioré (LM), with an unequivocal unconformity surface that was produced by the complete desiccation of the southern end of the lake; (B) Lagoa Gaíva (LG), with a subtle hiatus marking the same episode of drought from ~5300 to 2600 cal yr BP. This subtle hiatus marks an important shift from siliciclastic to biogenic ooze-dominated sediments. Panel (C) shows likely erosional unconformities (UNK) in LG created by crevasse flows in the early Holocene. Stars mark the locations of radiocarbon dates.

Unit IV (commencing after ~8900 cal yr BP)

Unit IV is a massive, upward fining package of tan clayey sand and brown–green silty–sandy clay with variable amounts of OM (Fig. 5). Detrital coarse:fine and oxidized grain counts are high near the base of the unit, then decline toward the upper contact (Figs. 5 and 6). Rb/K follows an opposite trend, decreasing slightly above the contact then increasing toward the middle of the unit (Fig. 6). C/N shows an overall declining trend in Unit IV, but values generally remain above 14. In contrast, values of TOC and $\delta^{13}C_{OM}$ markedly increase, reaching maximums of ~5.8 wt.% and −20.9‰, respectively. Abundant *Corvoheteromeyenia australis* spicules indicate a eutrophic wetland or lentic environment, possibly influenced by weak flooding (Fig. 7; Volkmer-Ribeiro, 1999). We interpret that LG was at lowstand and the PR was confined to its channel for much of Unit IV time. Carbon isotope and C/N values indicate a mixed OM provenance, likely influenced by oxidation in a shallow lake environment. The accumulation of charcoal fragments suggests that fires burned on the drought-conditioned landscape and were subsequently flushed into the basin (Fig. 6). Wildfires were probably most prevalent at the Unit V/IV transition and early in Unit IV time, when Rb/K data indicate reduced precipitation.

Unit III (ending at ~5300 cal yr BP)

Unit III is comprised of silty–sandy clay with variable OM content and massive gray-green silty clay (Fig. 5). Just above the Unit III/IV contact, detrital coarse:fine and oxidized grains are relatively high, whereas Rb/K sharply declines (Fig. 6). Detrital coarse:fine, TOC, and $\delta^{13}C_{OM}$ generally decrease toward the top of Unit III (Fig. 5). BiSi remains relatively static at <4.0 wt.%. High amplitude variability in C/N is conspicuous, with values ranging from 18 to 11 (Fig. 5). C/N numbers in this range are typically associated with mixed aquatic and terrestrial parent material (e.g., Meyers and Teranes, 2001). Lentic and wetland sponge spicules (dominantly *C. australis*) are abundant near the base of the unit, whereas lotic spicules (e.g., *O. navicella*) and fish remains begin to appear consistently near the middle of the unit (Figs. 6 and 7). We interpret Unit III as deposited in a

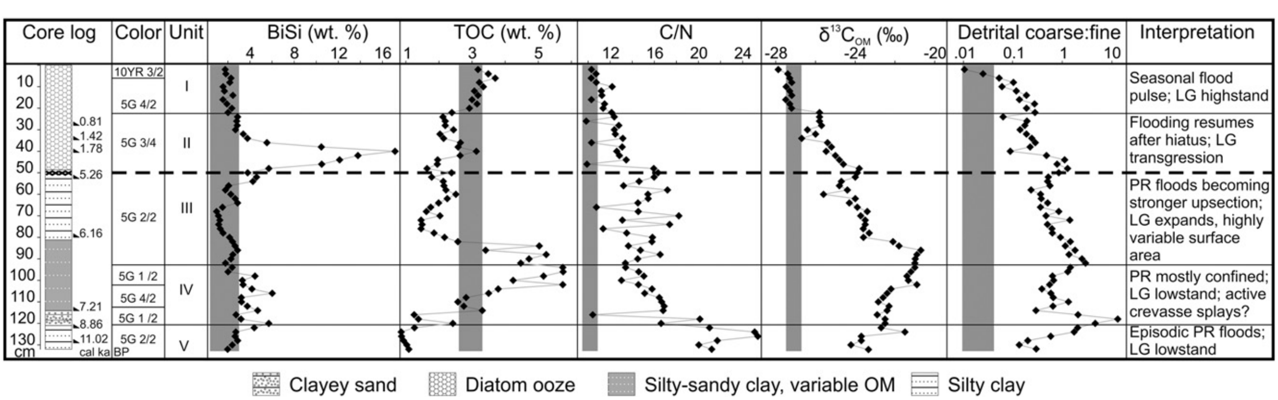

Figure 5. Lithostratigraphy, biogeochemistry and detrital coarse:fine datasets for Lagoa Gaíva (LG) core 2A. The dark vertical bars represent the range of values measured on modern lake-bottom sediments within 2–3 km of the core site (n = 6 nearest neighbors; data from McGlue et al., 2011). Note that a coarse:fine of 1 denotes equal amounts of detrital sand and silt + clay. Strong Paraguay River (PR) flooding appears after ~2600 cal yr BP, causing a transgression in LG. Flooding was weak and sporadic during the early to mid Holocene. A dashed line marks the location of a major drought-related hiatus. BiSi, biogenic silica; C/N, carbon to nitrogen ratio; OM, organic matter; TOC, total organic carbon.

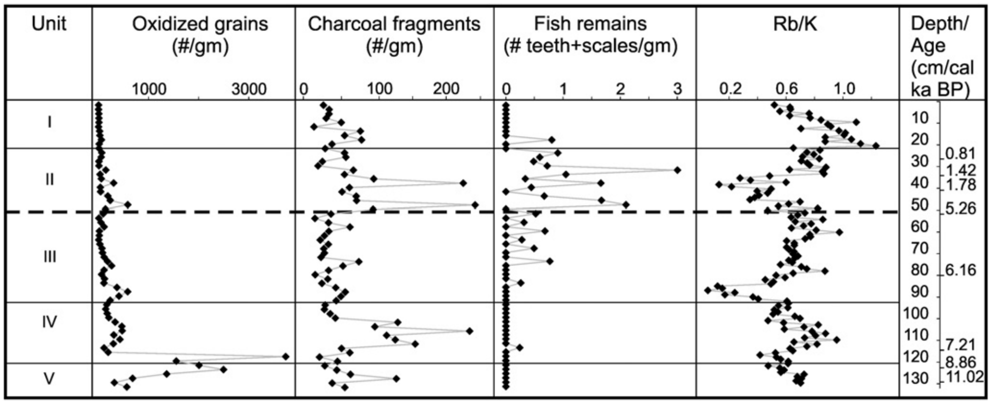

Figure 6. Screen (125 μm) washed residues and Rb/K chemostratigraphy for Lagoa Gaíva core 2A. Note 1-cm sampling interval for Rb/K data. Sands with oxidation patinas and charcoal fragments likely get flushed into the basin following intervals of drought. Fish fossils begin to appear consistently after 6200 cal yr BP and peak with transgression after ~2600 cal yr BP. High (low) Rb/K values indicate an increase (decrease) in the transport of weathered material to the lake, providing a relative index of precipitation. Three pronounced troughs mark the punctuated Holocene, whereas a peak in Unit I may reflect higher rainfall during the Little Ice Age.

fluctuating environment, with water levels initially low, followed by lake expansion under the influence of PR floods of increasing strength by ~6200 cal yr BP.

Holocene hiatus (~5300–2600 cal yr BP)

Evidence of a Holocene hiatus between Units III and II is subtle and marked by an indistinct stratigraphic contact with a change to a darker hue of green with scattered red-brown flecks (Fig. 4B). Particle sizes at the hiatus are relatively large (Fig. 5), whereas sponges are dominantly from lentic species (*Metania spinata* and *Trochospongilla variabilis*; Fig. 7). By contrast, this episode is dramatically expressed in southern LM, with a sharp contact separating highly oxidized yellow clay with plant roots from overlying BiSi-rich brown silty clay (Fig. 4A). This hiatus likely occurred due to isolation and evaporation of LG and LM as the PR was confined within its channel during an extended drought. Extrapolation of the linear sedimentation rate (0.13 mm/yr) that characterizes Unit II to its basal contact indicates that the late Holocene hiatus spanned ~2800 cal yr BP.

Unit II (~2600–440 cal yr BP)

Unit II consists of dark green diatom ooze (Fig. 5). Just above the hiatus, charcoal fragments and oxidized grains increase, and Rb/K declines (Fig. 6). C/N and $\delta^{13}C_{OM}$ follow trends toward lower values upward toward the top of Unit II. There is an important shift to silt-dominated detritus (decreasing coarse:fine) at ~1800 cal yr BP, suggesting a transgression that curtailed lowstand depositional processes (Fig. 5). Spicules from *O. navicella*, *T. variabilis*, *Corvospongilla seckti*, and *Radiospongilla amazonensis* indicate a major environmental transition and the initiation of PR flooding into a formerly turbid, eutrophic, lowstand lake (Fig. 7). Flooding is also marked by peaks in BiSi content and fish remains, as planktonic diatoms (typically *Aulacoseira* sp.) and nekton flourished in a deeper, more productive lake (Figs. 5 and 6). Anoxic events and fish kills are observed in the PR following extended drought today (the *dequada* phenomenon; Hamilton et al., 1997). Following this analog, fish may have sought refuge in LG following late Holocene aridity, leading to an accumulation of scales + teeth during Unit II time.

Unit I (~440 cal yr BP to present)

Unit I is comprised of massive, dark green, clayey ooze (Fig. 5). Detrital coarse:fine decreases to ≤0.1 due to an increase in clays, whereas C/N declines and $\delta^{13}C_{OM}$ becomes depleted, reaching values consistent with a lacustrine algal signal (Fig. 5). TOC increases to ~3.0–3.5 wt.% and BiSi remains around 2.0%, similar to the modern lake (McGlue et al., 2011). Rb/K increases above the Unit II/I contact, suggesting increased precipitation (Fig. 6). Total sponge spicules decline and the assemblage becomes dominated by *M. spinata* with

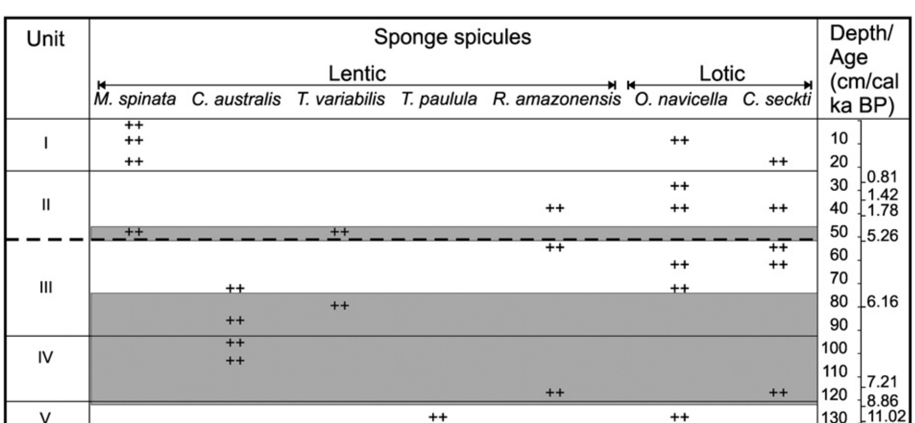

Figure 7. Sponge spicule data for Lagoa Gaíva (LG) core 2A. Note that sponge spicules present in recent LG sediments were documented in McGlue et al. (2011). Lentic sponge remains typically mark intervals interpreted to be most affected by drought (shaded in gray), implying confinement of the PR within its channel and at least partial isolation of LG on the floodplain. Lotic species appear consistently after ~2600 cal yr BP, likely due to the influence of strong PR flooding. ++ denotes presence in sample. See text for details. Sponge data have been archived at LEPAFE (code L.86, 87, 88, 89, 90.C.08).

minor *O. navicella* and *C. seckti* (Fig. 7). Taken together, these indicators signal lake-level highstand and a strong seasonal flood pulse by ~440 cal yr BP.

Discussion

Depositional continuity in floodplain lake strata

Floodplain lake sediments record the response of a fluvial landscape to climatic forcing. It is important to note, however, that the majority of facies and geochronological datasets from overbank and floodplain environments indicate discontinuous deposition (Mertes, 1994; Willis and Behrensmeyer, 1994; Aalto et al., 2003; Lewin et al., 2005). Detailed studies of paleochannel-lake stratigraphy generally reach similar conclusions, as the infilling of abandoned channels with sediment is intimately tied to episodic processes, including chute erosion, lateral channel migration, and fall-out of suspended load (e.g., Holbrook, 2010). Although shoreline morphology shows that the study lakes are not oxbows, the level of the PR nonetheless influences basin-filling processes due to its control on the interaction between lake-surface elevation and basin sill (e.g., Bohacs et al., 2003).

The depositional history captured in LM core 5A is clearly discontinuous and provides compelling evidence of complete desiccation of the southern end of the lake at ~4700 cal yr BP (Fig. 4A). Similarly striking stratigraphic evidence of desiccation in LG is absent, hinting at the possibility of a continuous record, as suggested by Whitney et al. (2011). However, careful examination of physical stratal surfaces coupled with a high density of radiocarbon dates confirms that the record is unequivocally punctuated prior to ~2600 cal yr BP (Fig. 4B,C). Sediments in core 2A are pervasively massive, which suggests a zone of active bioturbation and supports the interpretation of very shallow lake levels. This is expected, in as much as the volume of LG (~0.31 km^3) is probably incapable of buffering major hydroclimatic perturbations, leading to rapid lake-level fluctuations. Clearly, LG and LM are responding to the same episode of low effective precipitation (P:E) and PR discharge in the late Holocene, as the latter is situated <100 km downstream of the former. We conclude that differences in basin shape and coring location control how this episode is manifested in stratal archives recovered in cores. This is consistent with our modern process study, which showed that shallow maximum depths, fetch orientations, and shoreline topography affect the continuity of sediment accumulation in both lakes, with the stratal record in LM more acutely punctuated than LG (McGlue et al., 2011).

In LG core 2A, the direction of detrital coarse:fine follows a consistent pattern (higher) with respect to hiatuses in the stratigraphic record. At least three such examples are evident, attesting to the punctuated nature of the record prior to 2600 cal yr BP (Fig. 4B, C). Although the magnitude of the response varies in time, oxidized grains (higher) and Rb/K (lower) also track these events. Particle-size data are especially valuable in the study of floodplain lakes, as the impact of erosive flows is not clearly reflected in more traditional biogeochemical indicators. Moreover, depositional continuity must be measured at the scale of bedding (typically 10s of cm or less for overbank environments; Bridge, 1984) in floodplain lakes for a robust chronological assessment. Linear extrapolation of age data across bedding planes is fraught with complication in shallow lowland lake cores, a point likewise emphasized by Ledru et al. (1998a). Clearly, radiocarbon dating at high density is necessary to reduce time-averaging of multicentennial to millennial events in these types of archives.

Paleoenvironments in the Pantanal

Early to mid-Holocene

Stevaux (2000) suggested a wet phase persisted in the Parana River watershed from ~8000 to 3500 cal yr BP. A number of other studies argue in favor of early Holocene humidification culminating in a climatic optimum around 6500 cal yr BP (Assine and Soares, 2004; de Oliveira Bezerra and Mozeto, 2008). The most recent study is a detailed pollen diagram from LG, which argues for a major climatic transition at 12,900–12,200 cal yr BP based on a shift from floodplain forest to savanna grassland vegetation (Whitney et al., 2011). These authors posit that temperature increased in the terminal Pleistocene and that the Pantanal experienced moderate drought from ~10,000 to 3000 cal yr BP.

Our data from LG supports the early to mid-Holocene trend argued for by Whitney et al. (2011). We interpret the significant differences among the units of LG core 2A to reflect a step-wise response of the lake basin to variability in PR flooding. From 11,000 to 6200 cal yr BP, detrital sands are common in LG strata, and the ability of a river to move sand is a function of the amount of water available and gravitational potential energy (e.g., Heins and Kairo, 2007). Because the longitudinal profile of the PR from Cáceres (where it enters the Pantanal) to the study area near Amolar shows only minor topographic variability (<150 m over ~300 km), stream power is most likely controlled by discharge. Given the compelling evidence for terminal Pleistocene aridity across western Brazil, increased PR discharge in the early Holocene appears to be a necessary condition for coarse siliciclastic deposition in LG. However, our paleolimnological reconstruction indicates that early Holocene PR flooding was highly dissimilar to the flood pulse that defines the Pantanal wetlands today. We suggest that the transition to greater P:E, coupled with soil stabilization through savanna grassland development, probably caused incision and confinement of the PR and low water levels in floodplain lakes as a boundary condition for the early Holocene, with PR discharge significantly lower than present due to low insolation and a weak SASM (Berger, 1992). Complex paleochannel patterns and incision upstream of LG appear to reflect these environmental conditions, but only relative geochronology is available for these features (Assine and Silva, 2009). We conclude that infrequent or weak floods may have driven crevasse development across PR levees, sending unconfined flows into LG. The pattern of coarse:fine variation, which shows upward coarsening overlain by finer particles at the boundaries of Units V–IV and IV–III, is consistent with progradation of thin, sandy lobes into a lowstand lake (e.g., Fielding, 1984; Bridge, 1984; Aslan and Blum, 1999).

Another clue about water levels in the early Holocene comes from the geochemistry of OM (Supplementary Fig. 1). Elevated C/N and enriched $\delta^{13}C_{OM}$ values in Units V–III most likely reflect mixed-source OM impacted by oxidation and reworking in a lowstand lake setting. A number of other factors could be responsible for enriched $\delta^{13}C_{OM}$ values, including high rates of algal productivity, reservoir effects, or low concentrations of aqueous CO_2 (Talbot and Johannessen, 1992). However, insights from lithofacies and C/N are most consistent with a fluctuating lowstand environment in the early Holocene, making alternative interpretations of $\delta^{13}C_{OM}$ less viable (Fig. 5). Although caution is warranted in the interpretation of bulk $\delta^{13}C_{OM}$, the ~4.0–8.0‰ increase relative to modern values provides confidence for a lowstand interpretation. Commonly used biogeochemical indicators of productivity and biomass, BiSi and TOC, are of limited individual value in constraining lake level during this time. However, elevated early to mid-Holocene carbon burial characterizes a number of lake systems in the Pantanal and has previously been interpreted as a signal of warm, wet climate (e.g., de Oliveira Bezerra and Mozeto, 2008). In LG, sediments with high TOC wt.% at the top of Unit IV and basal Unit III are also characterized by high coarse:fine and lentic sponge spicules, which suggests a lowstand lake impacted by weak flood flows, not a highstand lake influenced by a true flood pulse (Figs. 5 and 7). However, it is important to note that limited knowledge of the age and origins of the Pantanal's large lakes represents a shortcoming for all sediment core-based paleolimnological research. If LG occupies a former tributary of the PR, then lake formation by aggradation of the trunk channel is a potential alternative interpretation of the early Holocene stratigraphy, with Unit IV representing an initial lacustrine phase and Unit III an expansion of the lake with stronger flooding. Based on multiple indicators, we favor the lowstand interpretation and conclude that expanded dry-season length likely influenced a pre-existing lake system at this time (an inference supported by Pleistocene paleoecological data; Whitney et al., 2011), but future research involving sub-bottom imaging should help resolve this question. This interpretation is consistent with pollen studies from Brazilian savanna lakes, which argue for dry conditions at the Pleistocene–Holocene transition (Salgado-Labouriau, 1997). Climate data and models suggest that during low insolation phases, precipitation in the South American continental interior should be antiphased with northeastern Brazil, and well-dated speleothem isotope records detailed in Cruz et al. (2009) provide compelling evidence for this relationship in the early Holocene.

The first strong signal of increased P:E is at the conclusion of elevated carbon burial in Unit III, at the transition to green silty clay with elevated Rb/K at ~6200 cal yr BP (Figs. 5 and 6). The decline in coarse:fine and negative ~3.5–5.0‰ shift in $\delta^{13}C_{OM}$ indicate suspension fall-out and mixed algal–terrestrial OM deposition in a variable lake setting until ~5300 cal yr BP. Lotic sponges and the first appearance of fish fossils in this interval further demonstrate connectivity with the PR (Figs. 6 and 7). In the southern Pantanal, riverine tufas indicative of strong baseflow and a shift to higher P:E have been dated to ~6500 cal yr BP (Sallun Filho et al., 2009). Although relatively low at the start of the Holocene, insolation at 15°S gradually began to increase after ~7500 cal yr BP (Berger, 1992), which likely caused the SASM to strengthen. Regional keystone datasets support higher precipitation in step with insolation at this time, including speleothem $\delta^{18}O$ from the Botuverá cave in southern Brazil and pollen records from central Brazil (Ledru et al., 1998b; Cruz et al., 2005).

Mid- to late Holocene

Marchant and Hooghiemstra (2004) documented major changes in hydrology and vegetation at ~4000 cal yr BP at many sites in tropical South America that suggest a shift toward greater precipitation or shorter dry seasons. The database compiled by Stevaux (2000) indicates an interval of widespread aridity from 3500 to 1500 cal yr BP for the Upper Parana watershed and other lowland areas in central South America. Our data from LG and LM, plus other records from the Pantanal wetlands are somewhat different (Table 2). Evidence of high moisture levels and regular PR flooding first appear in the LG record in Unit II after ~2600 cal yr BP. This interpretation is broadly consistent with the vegetation reconstruction described in Whitney et al. (2011) and with a number of other key tropical sites around South America where aridity persisted beyond ~4000 cal yr BP (e.g., Thompson et al., 1995). Carbonate deposits in the southern Pantanal point toward aridity around 4100 cal yr BP, and stalagmite growth bands show more frequent instances of drought between ~3800 and 2500 cal yr BP (Bertaux et al., 2002; Assine and Soares, 2004). Insolation at lat 15°S was relatively high (Berger, 1992) and SSTs in the tropical Pacific appear to have played a stronger role in the modulation of continental P:E and thus transgression at LG after ~3000 cal yr BP (Koutavas et al., 2006).

Unit I is an intriguing component of the LG stratigraphy, as its inception (~440 cal yr BP) falls during the Little Ice Age, a period marked by a southerly Intertropical Convergence Zone position and higher rainfall in several regions of tropical South America from ~AD 1250 to 1810 (Haug et al., 2001; Polissar et al., 2006; Reuter et al., 2009; Bird et al., 2011). A positive excursion of Rb/K near the base of Unit I and decreasing coarse:fine moving toward the top of the unit are consistent with higher precipitation and suspension fall-out processes in a highstand lake, respectively. Higher lake levels are likewise implied by the abrupt shift to $\delta^{13}C_{OM}$ values sustained below −27.0‰ and the progressive decline in C/N at the start of Unit I.

Implications

The shorelines of large lakes like LG and LM were the sites of permanent settlements for pre-colonial indigenous groups, who took advantage of lake-margin topography to cope with PR flooding (Peixoto, 2009). Archeological remains excavated from mounds just south of the study area point to the inception of an important ceramics tradition amongst indigenous people at ~2600 cal yr BP (Felicissimo et al., 2010). Microanalysis of potsherds has shown that fluvial sponge skeletons were selected by artisans as an additive to clays, likely in an effort to increase the strength of pottery fired at relatively low temperatures (Felicissimo et al., 2010). This is highly feasible based on

Table 2
Summary of available late Quaternary paleo-datasets for the Brazilian Pantanal. Note that January insolation at lat 15°S is relatively low (≤445 W m^{-2}) from ~10,000 to 7000 cal yr BP (Berger, 1992). Most datasets suggest Pleistocene aridity in the Pantanal and greater environmental variability in the Holocene. Our lacustrine records show that the PR flood pulse did not approach modern conditions until the late Holocene. See text for details.

Region	Site	Age range (method)	Paleoenvironments	Reference
Northern Pantanal	Rio Paraguay fan	Pleistocene-present (relative)	Paleo-channels, incision patterns linked to dry terminal Pleistocene and wet, variable Holocene	Assine and Silva (2009)
Central Pantanal	L. Gaíva	~45,000 cal yr BP to present (^{14}C on terrestrial, bulk OM)	Pollen, diatoms suggest transition to savanna vegetation and wet climate at ~12,900 cal yr BP; moderate drought ~10,000–3000 cal yr BP	Whitney et al. (2011)
	L. Negra and L. Castelo	~18,600 cal yr BP to present (^{14}C on bulk OM)	TOC in lake sediments linked to a wet Holocene, with a “climatic optimum” at ~6500 cal yr BP	de Oliveira Bezerra and Mozeto (2008)
	L. Gaíva and L. Mandioré	~11,000 cal yr BP to present (^{14}C on <63 μm OM)	Multiple lake sediment and biological indicators suggest sporadic, weak PR flooding in early-mid Holocene; drought 5300–2600 cal yr BP; stronger flood pulse after ~2600 cal yr BP	This study
	Rio Taquari	8650 ^{14}C yr BP (charcoal)	Floodplain incision, sand sheet deposition linked to wet early Holocene following dry late Pleistocene	Soares et al. (2003)
Southern Pantanal	Miranda and Aquidauana	3820 ^{14}C yr BP (shell $CaCO_3$)	Mollusk shells embedded in fluvial strata linked to a dry climate around 4000 cal yr BP	Assine and Soares (2004)
	Rio Nabileque	Pleistocene–present (relative)	Abandoned meander belt, incision patterns linked to Holocene climatic variability	Kuerten and Assine (2011)
	Bodoquena	~3800 cal yr BP to present (U-series on speleothems)	Speleothem growth bands, geochemistry indicate more frequent instances of drought from ~3800 to 2500 cal yr BP	Bertaux et al. (2002)
	Rio Aquidaban	~3700–1900 cal yr BP (^{14}C on tufa)	Tufas linked to dry climate in late Holocene	Boggiani et al. (2002)
	Bodoquena	~6500–2700 cal yr BP (^{14}C on tufa)	Elevated fluvial dam-type tufas linked to strong riverine baseflow from ~6500 to 2700 cal yr BP	Sallun Filho et al. (2009)
	Nhecolândia	~11,400 cal yr BP to present (^{14}C on soil OM)	Carbon isotopes in soils suggest transition to savanna landscape between ~11,400 and 4600 cal yr BP	Victoria et al. (1995)
	Nhecolândia	Pleistocene (relative)	Deflation pans, lunettes linked to dry Pleistocene	Klammer (1982)

our reconstruction, as *O. navicella* and *C. seckti* would have become readily available resources around 2600 cal yr BP, once strong seasonal PR flooding resumed following Holocene drought.

Resolving the nature and timing of major hydrologic changes in the Pantanal is vital, not only for improving understanding of the evolution of this sensitive wetland system, but also because these changes have implications for atmospheric methane (CH_4). Notably, Nahlik and Mitsch (2011) showed the importance of seasonal moisture contrasts and flood pulsing on CH_4 emissions using experiments from tropical wetlands in Costa Rica. Estimates of current dry-season CH_4 emissions from the Pantanal based on extrapolations from floodplain lakes suggest a globally significant annual flux (at least 1.9 Tg CH_4 yr^{-1}; Marani and Alvalá, 2007; Bastviken et al., 2010). Recent research suggests that Southern Hemisphere wetlands were an important source of CH_4 after 5000 cal yr BP (Singarayer et al., 2011). Data from LG support this hypothesis, with the environmental conditions known to be favorable for wetland methanogenesis, including warm air and soil temperatures, shallow inundation driven by the seasonal PR flood pulse, and relatively high rates of organic matter burial after ~2600 cal yr BP.

Conclusions

1. In the Pantanal wetlands, assessing Holocene changes in PR flooding benefitted from detailed microstratigraphic observations, assessment of multiple biogeochemical and sedimentary indicators, and bedding-unit scale radiocarbon dating on floodplain lake cores.
2. Unit-scale radiocarbon and stratigraphic variability defined an interval of punctuated sedimentation from ~11,000 to 6200 cal yr BP in LG, with erosion of the lake floor by crevasse-derived flows. A weak SASM produced low water levels in LG during this interval, spurring oxidation of mixed-source OM. What PR flooding did occur prior to ~2600 cal yr BP was likely episodic and weak compared to the modern flood pulse. Flooding appears to have become more frequent in the central Pantanal from ~6200 to 5300 cal yr BP, which is consistent with evidence of higher P:E in the southern Pantanal and other areas of Brazil. An episode of Holocene drought and confinement of the PR in its channel was especially pronounced in the Pantanal, as shown by the development of a hiatus in LG (~5300–2600 cal yr BP) and the desiccation of the southern end of LM at ~4700 cal yr BP.
3. The late Holocene witnessed a transgression at LG after ~2600 cal yr BP spurred by strong PR flooding, most likely in response to a strong SASM and steepened SST gradients in the tropical Pacific. Widespread inundation had important implications for global CH_4 flux and indigenous populations living along lake margins. Highstand conditions at LG and a true PR flood pulse appears to have occurred ~440 cal yr BP, coeval with higher P:E during the Little Ice Age.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.yqres.2012.05.015.

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