TITLE: Seasonal and spatial variability of CO$_2$ emission from a large floodplain lake in the lower Amazon

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RUNNING TITLE: CO$_2$ variability in a large Amazon lake
Abstract

The inundation status of the Amazon floodplain affects biogenic gas production and evasion. We analyzed spatial variability of dissolved CO$_2$ concentration and gas evasion in a large floodplain lake in the lower reach of the Amazon River in four hydrological phases. We calculated surficial CO$_2$ concentrations from measurements of pH, dissolved inorganic carbon, temperature, and conductivity and use meteorological data to calculate gas transfer coefficients to estimate CO$_2$ evasion. Gas transfer coefficients which take into account both wind and heating and cooling at the lake’s surface are on the order of four times higher than values previously used in regional estimates of gas evasion from lakes on the Amazon floodplain. Supersaturation of CO$_2$ occurred throughout the lake and was higher in the littoral zone and in regions receiving Amazon River inflows. CO$_2$ concentration was reduced in regions with phytoplankton blooms. The range of CO$_2$ concentrations was least at low water, 47 µM to 233 µM, and largest at high water, 4 µM to 1,591 µM. The average annual value was 172 µM and is typical of lakes in the downstream reaches of the Amazon. Based on seasonality in flooding extent, surface water CO$_2$ concentrations, and gas transfer coefficients, we estimate Lake Curuaí outgassed from its open-water to the atmosphere 2.3 Tg C yr$^{-1}$. 
1. Introduction

The Amazon basin stores and cycles a large quantity of carbon. Different approaches
used to determine the region’s carbon balance suggest it has been, on average, nearly neutral
over the last decade [Houghton et al., 2009]. However, there remain uncertainties regarding
changes in land use and in the balance between photosynthesis and respiration due to variations
in climate. The fluvial system, recently recognized as an important component of the carbon
budget, releases to the atmosphere an amount of CO$_2$ roughly equal to the carbon sequestered by
the forest. The basin-wide outgassing of CO$_2$ from rivers and floodplains was estimated by
Richey et al. [2002] as 500 Tg C yr$^{-1}$, but this is considered conservative and is currently under
revision [Richey et al., 2009]. Refined analyses of spatial and temporal patterns of fluxes are
required for uncertainties to be resolved.

Carbon dioxide is supersaturated in most Amazonian rivers and floodplain waters [Richey
et al., 1988; Richey et al., 2002] and its exchange with the atmosphere occurs via diffusive and
turbulent transfer across the air-water interface [MacIntyre et al., 1995]. On the scale of an
individual lake, Melack and Engle [2009] reported that CO$_2$ evasion represented the majority of
carbon loss. The production of dissolved CO$_2$ in Amazon floodplains and other tropical wetlands
includes respiratory releases from aquatic plants and animals, but is dominated by microbial
metabolism [Hamilton et al., 1995; Waichman, 1996] which consumes labile organic matter of
young age released from seasonally flooded forest, aquatic macrophytes and upland forests
[Quay et al., 1992; Mayorga et al., 2005; Engle et al., 2008; Melack et al., 2009; Richey et al.,
2009].

The annual flood of the Amazon River influences the ecology and biogeochemistry
within its floodplain and plays a major role in retention and transformation of organic matter and
exchange of gases between land, water and the atmosphere [Junk, 1997, Melack et al., 2009].

Richey et al. [1988] measured CO₂ concentration in lakes as part of cruises along the Amazon River, but within lake spatial and temporal variability has yet to be examined. Hence, we determined spatial distributions of CO₂ concentration and efflux from open-water of Lago Grande de Curuaí (Lake Curuaí) for four hydrological phases. We calculate surficial CO₂ concentrations from measurements of pH, dissolved inorganic carbon (DIC), temperature, and conductivity and use meteorological data to calculate gas transfer coefficients to estimate CO₂ evasion. The results from our study complement earlier estimates of CO₂ evasion for Amazon floodplains in three important ways: (i) Large shallow lakes, such as Lake Curuaí, are representative of the downstream reach of the Amazon River floodplain [Sippel et al., 1992] and have not been included in previous studies. (ii) The large number of samples permits examination of within-lake spatial variability. (iii) For the first time, samples spanning four hydrological phases are combined with meteorological measurements in the same lake to calculate gas exchanges.

2. Site description

Lago Grande de Curuaí lies in the floodplain of the Amazon River to the south of Óbidos, Pará, Brazil (Figure 1). It is composed of several interconnected lakes temporarily or permanently connected to the Amazon R. by channels and overbank flow. The stage of the Amazon R. varies ca. 6 m annually in the nearby reach with a peak in June and minimum in November (Figure 2). The lake’s flooded area ranges from about 850 to 2274 km²; at flood peak open-water covers about 65% of flooded area with flooded vegetation covering the remaining area (Table 1). Water movements and mixing in L. Curuaí are complex, and the effects of physical forces (e.g., wind, solar radiation) are modified by the lake’s surface area and depth,
varying seasonally and spatially (Figure 3a-d and Table 1). During low water the lake becomes sufficiently shallow for sediments to be resuspended by wind-induced turbulence. During high river inflows the residence time of water in the lake is reduced.

Bonnet et al. [2008] determined the hydrologic balance for the lake, and reported the lake and its catchment to be, on an annual basis, a source of water to the Amazon R., which accounts for about three quarters of annual inputs of water to the lake. Runoff from the local catchment, direct rainfall, and groundwater seepage were estimated to represent ca. 10%, 9%, and 4% of the annual total inputs, respectively. Each year, the storage phase of L. Curuaí starts between November and January and lasts until May-June; the drainage phase starts in July and lasts until November.

Barbosa et al. [2010] described large seasonal and spatial variations in total suspended sediment (TSS), chlorophyll-a, and pH throughout L. Curuaí. Average surface water concentrations of TSS changed from 14.5 mg L\(^{-1}\) during high water to 463 mg L\(^{-1}\) during low water. Average chlorophyll-a varied from 8 \(\mu\)g L\(^{-1}\) during rising water to 70 \(\mu\)g L\(^{-1}\) during receding water. While pH in the Amazon R. was relatively constant throughout the year (ca. 6.5), pH in the lake varied from an average of 6.7 during low water to 7.7 during receding water. High pH, up to 9.2, was found during receding water in areas with algal blooms of up to 338 \(\mu\)g L\(^{-1}\) of chlorophyll-a.

3. Methods

Field measurements: Four campaigns were conducted to sample the four phases of inundation [Barbosa et al., 2010]: receding (Sep. 2003), low (Nov. 2003), rising (Feb. 2004), and high water (June 2004). Surface water (~0.5 m) was sampled at approximately 72 spatially distributed stations over 10-day periods (Table 2). Transparency was measured using a 20-cm
diameter Secchi disk. Temperature, dissolved oxygen, conductivity and pH were measured with a Horiba U-10 water quality meter. The pH probe was calibrated at the beginning of each day using reference solutions with pH 4 and 7. Water was collected, filtered under vacuum through glass fiber filters (Whatman GF/C, 47 mm diameter), and frozen in the field for subsequent laboratory analysis. These manipulations may change slightly DIC concentrations. In the laboratory, concentrations of DIC were determined using a Shimadzu TOC-5000A carbon analyzer calibrated with potassium biphthalate.

Measurements of meteorological parameters and water temperatures, required for computation of gas transfer coefficients, became available in April, 2004, from an instrumented buoy moored in the southeastern region of the lake (Figure 1). Hence, the high water campaign in June 2004 was the only one with concurrent data. In order to approximate conditions during the three other seasons, measurements acquired in the corresponding days of the following year were used. A few gaps, due to telemetry signal losses, were filled with data from the Vila Franca meteorological station, located 20 km to the southeast (2° 20' 55" S; 55° 1' 44" W) [Fitzjarrald et al., 2009].

Meteorological parameters monitored included water column temperature at 1 m and air temperature (shielded thermistors: MEAS 44033RC; accuracy of ±0.1 °C), relative humidity (Rotronic MP103A-CG-0200; accuracy of ±1.5 % at 23 °C), down- and upward shortwave radiation (wavelength range of 400-1100 nm; sensitivity of 490 µV/W/m²; accuracy of ±16%), wind (R. M. Young 05106; wind speed, accuracy of ±0.3 m s⁻¹ and wind direction, accuracy of ±3 degrees) at 3 m above the lake, and atmospheric pressure (Vaisala PTB100-A; accuracy of ±0.3 mb at 20 °C). Longwave radiation emitted from the lake surface was estimated from water temperature according to the Stefan-Boltzmann law with an emission coefficient of 0.97. The
downward flux of longwave radiation was obtained as hourly averages from two years (2002 and 2003) measured at the Tapajos National Forest (3° 1' 5" S; 54° 58' 8" W) [Miller et al., 2009].

Depth of the active mixed layer ($z_{AML}$) is required for computation of the gas transfer coefficients. Values of $z_{AML}$ were assumed to be 0.5 m or the seasonal lake mean depth ($z_{mean}$), because temperature profiles were not measured. This decision was based on our examination of the frequency and extent of vertical mixing in Lake Calado, determined from multi-year time series [Melack and Fisher, 1983; MacIntyre and Melack, 1984, 1988], and the seasonal morphometric changes of L. Curuaí (Figure 3a-d and Table 1). Seasonal stratification persisted once the depth of L. Calado reached about 6 m. During stratification, the mixed layer deepened at night due to cooling and stratified in the day due to heating. Similar patterns are expected in L. Curuaí although wind induced mixing may cause greater mixed layer depths in the larger lake. However, computed attenuation coefficients reach 18 m$^{-1}$ during low water (see Table 2); hence daily stratification is likely. At high water, regions deeper than 6 m represent only 16% of open-water, and this portion is insignificant during the other hydrological phases. Based on similarities and differences between L. Curuaí and L. Calado, the water column in L. Curuaí during the low and medium (rising and receding) water stages is likely to be fully mixed at sunrise, develop a shallow thermocline (~0.5 m) towards mid-day at low wind conditions, and fully mix again at night. During the high water stage, the lake may develop periods of stable stratification under low wind conditions with significant diel variation of the mixed layer depth (1 to 6 m, depth permitting).

**Calculation of CO$_2$ concentrations:** Dissociation constants from Plummer and Busenberg [1982] and inorganic carbon species as a function of pH from Stumm and Morgan [1996] were used for conversion of DIC to dissolved carbon dioxide concentrations in surface waters (CO$_{2aq}$).
Adjustments were made for ionic strength; without complete major ion data, we estimated ionic strength from conductivity [Butler, 1992]. Atmospheric partial pressure of CO$_2$ of 379 µatm, the global value reported in IPCC [2007], was used, and seasonal variations were not included. The concentration of dissolved CO$_2$ in equilibrium with the atmosphere (CO$_{2eq}$) was calculated using Henry's law. For comparison in our discussion, concentrations presented in µatm in Richey et al. [2002] were converted to mM, using Henry’s constant for CO$_2$ at 25°C (0.0339 M atm$^{-1}$) and correcting for the partial pressure of water (0.0313 atm).

**Gas exchange at the air-water interface:** The magnitude and direction of the gas flux is dependent on the concentration difference between the atmosphere and water and upon transport processes at the interface. CO$_2$ fluxes were calculated using the dissolved CO$_2$ concentration departure from equilibrium and a gas transfer coefficient ($k$).

$$F = k (CO_{2aq} - CO_{2eq})$$

Due to the uncertainty in parameterizing the gas transfer coefficient, we used three models to determine a range of $k$ values.

The first model used is the wind-based approach of Cole and Caraco [1998]:

$$k_{600, C&C} = 2.07 + 0.215 \ U_{10}^{1.7}$$

where $k_{600}$ is the gas exchange coefficient normalized to the Schmidt number of CO$_2$ at 20°C in freshwater ($Sc = 600$) and $U_{10}$ is the wind speed at a 10 m height above the water surface. Wind measured at 3 m was converted to $U_{10}$ assuming law of the wall scaling and taking into account atmospheric stability.

The second model used is a small eddy version of the surface renewal model, which takes into account turbulence generated by wind action and heat losses that cause convective motions in the water column [MacIntyre et al., 1995].
where $v$ is the kinematic viscosity. We let $c_1 = 0.56$ [Isenogle, 1985]. The coefficient has not yet been fully established and current values range from 0.419 to 1.2 (Zappa et al., 2007; MacIntyre et al., 2010). The dissipation of turbulent kinetic energy ($\varepsilon$) in the active mixed layer was computed from the water friction velocity ($u_w^*$) and buoyancy flux ($\beta$).

\[
\varepsilon = 0.84[0.58(-\beta) + 1.76u_w^* 3/ k_{AML}]
\] (4)

where the coefficients were determined empirically [Lombardo and Gregg, 1989] and $k$ is the von Karman constant. $u_w^*$ is computed from shear stress at the air-water interface assuming it is equal on either side of the interface, giving $\tau_w = \rho_w u_w^* 2 = \tau_a = \rho_a C_d U^2$ where $\tau_a$ and $\tau_w$ are shear stress on air and water side of the air-water interface; $U$ is wind velocity at 3 m height and $C_d$ is a drag coefficient computed for that height following Amoroch and DeVries [1980]; $\rho_a$ and $\rho_w$ are density of air and water, respectively. The buoyancy flux is calculated as

\[
\beta = g \alpha H / \rho_w C_p
\] (5)

where $g$ is gravity, $\alpha$ is the coefficient of thermal expansion, $C_p$ is heat capacity, and $H$ is the effective heat flux.

\[
H = S + q(0) + q(z_{AML}) - \frac{2}{z_{AML}} \int_{z_{AML}}^0 q(z) dz
\] (6)

Where $S = \rho_w C_p <w'T'(0)>$ is the surface heat flux, obtained by the sum of latent heat flux, sensible heat flux, and longwave back radiation; $q$ is shortwave radiation; and $z$ is depth. Light attenuation coefficient ($k_d$) was estimated based on Secchi disk (SD) measurements using the conversion factor for turbid lakes $k_d = 1.3 / SD$ suggested by Koenings and Emundson [1991]. The penetration of radiation into the water column was parameterized using Beer’s Law following Jellison and Melack [1993]; angle of refraction was included in the calculation.
Following MacIntyre et al. [2010], a damping coefficient \( (c_2/z_{AML}) \) was introduced to equation 4 to account for the effect of dampening of turbulence during periods of heat gain. Dissipation (equation 4) was then computed as 
\[
\varepsilon = 0.84[0.58(-\beta) + (c_2/z_{AML})1.76u_*^3/k z_{AML}] \text{ when } L_w/z_{AML} \geq 0 \text{ and } U_{10} < 3 \text{ m s}^{-1}; \text{ where } L_w \text{ is the water-side Monin-Obukhov length scale. Since data are lacking to determine } z_{AML}, \text{ the computation was run twice with } z_{AML} \text{ fixed at } 0.5 \text{ m and at the seasonal lake mean depth (} z_{\text{mean}} \text{) which represent the approximate range of diurnal variation expected for L. Curuaí (as described above).}
\]

The third model used is a wind-based model which includes diel heating and cooling [MacIntyre et al., 2010].

\[
\beta > 0 \text{ (cooling)} \quad k_{600, EC} = -0.15 + 1.74 U_{10} \quad (7)
\]

\[
\beta < 0 \text{ (heating)} \quad k_{600, EC} = 2.0 + 2.04 U_{10} \quad (8)
\]

CO\(_2\) exchange coefficients \((k)\) were converted as a function of field conditions based on the Schmidt number \((Sc)\) dependency, assuming \(n = 0.5\) which corresponds to the case of turbulent flow below a free surface [Jähne et al., 1987].

\[
k_i = k_{600, i} (Sc/600)^{-n} \quad (9)
\]

where \(i\) is an index representing the different models (C&C; SR; EC).

4. Results

CO\(_2\) concentrations: L. Curuaí and the Amazon R. near Óbidos showed a seasonal pattern of mean CO\(_2\) concentration peaking during high water and decreasing at receding water (Figure 2). Lake CO\(_2\) concentrations ranged from 0.3 \(\mu\)M to 608 \(\mu\)M during receding water with low values in pelagic regions and high values towards the littoral zone (Figure 3e). The concentration in the Amazon R. at receding water was 225 \(\mu\)M but did not influence lake waters during this phase of floodplain drainage. At low water, CO\(_2\) concentrations ranged from 47 \(\mu\)M to 233 \(\mu\)M.
and were fairly uniform throughout the lake (Figure 3f). CO₂ concentrations increased toward the eastern channels due to the influence of Amazon R. inflow with CO₂ concentration estimated as 495 µM. During rising water, CO₂ concentrations ranged from 36 µM to 560 µM and were spatially heterogeneous with higher concentrations observed in littoral regions (Figure 3g). Amazon R. inflow from the east continued to increase CO₂ concentration in the floodplain. During high water, surface concentrations of dissolved CO₂ were higher than in the other periods and ranged from 4 µM to 1,591 µM (Figure 3h). Based on two samples, surface concentrations in the Amazon R. were 880 µM and 991 µM. Amazon R. inflows from western channels and northern overbank flow increased CO₂ concentration in receiving waters of the floodplain. Low CO₂ concentration was observed in interior pelagic regions.

**Meteorological conditions:** Ten-day time series of meteorological measurements and computed energy fluxes for the four hydrological phases are shown in Figure 4. Wind generally intensified during daytime and weakened during nighttime, reaching speeds above 5 m s⁻¹ on most days (Figure 4a-d). Ranges of diel variation of water temperature measured at 1 m varied seasonally (Figure 4e-h). Greater diel variation of water temperature was observed at low water. Toward the end of the rising water period and during high water, air temperature remained lower than water temperature. Maximum net shortwave radiation typically exceeded 1000 W m⁻² each day, except during June (Figure 4i-l). Sensible heat flux remained negative throughout June (Figure 4m), and ranged from about 20 W m⁻² during nighttime to -15 W m⁻² during daytime in September (Figure 4n); 50 W m⁻² during nighttime to -15 W m⁻² during daytime in November (Figure 4o); and 40 W m⁻² during nighttime to -30 W m⁻² during daytime in February (Figure 4p). Net longwave radiation ranged typically from -30 W m⁻² during nighttime to -70 W m⁻² during daytime in all periods. Evaporative cooling was substantial and dipped to values lower
than -200 W m$^{-2}$ during windy periods and when surface water temperatures were higher than the air temperatures.

The effective heat flux, which indicates whether the actively mixing layer will gain or lose heat to the atmosphere, was positive during daytime, indicating that the upper water column tended to stratify at those times, and negative at night indicating mixed layer deepening would occur. The effective heat flux was slightly higher during daytime when $z_{AML}$ was assumed to be $z_{mean}$, compared to when $z_{AML} = z_{0.5}$. due to the greater volume present to absorb and distribute the heat (Figure 4q-t). However, the timing of the shift from positive to negative values was independent of the value of $z_{AML}$, hence the sign of buoyancy flux, which is computed from effective heat flux and is needed for equation 4, was independent of $z_{AML}$.

**Gas transfer coefficients:** Figure 5 shows $k_{600}$ computed using the different approaches. The lowest estimates were obtained using Cole and Caraco’s [1998] wind based model, $k_{600, C&C}$. Values were on average 85% higher when computed using the regression equations based on wind and buoyancy flux, $k_{600, EC}$. Estimates of $k_{600, SR}$ with the surface renewal model depended upon the assumed $z_{AML}$. When using $z_{AML} = z_{0.5}$, $k_{600, SR}$ was similar to $k_{600, EC}$. When assuming $z_{AML} = z_{mean}$, seasonality was evident with proportionally lower estimates obtained during high water.

**Evasion of CO$_2$:** CO$_2$ evasion from open-water was estimated for each hydrological phase (Figure 6b) using the mean values of CO$_2$ departure from saturation (Table 2; $CO_{2aq} - CO_{2eq}$), and gas transfer coefficients averaged over the 10-day time series (Figure 6a) of each field campaign.

By combining CO$_2$ flux with the areal extent of open-water for each phase (Table 1), the highest rate of evasion was observed during high water (Figure 6c), corresponding to highest
mean CO$_2$ concentration and lake areal extent, while little seasonal variation was observed among mean $k_{600}$ values (Figure 6a). Integrated over the year, by considering the fluxes computed based on $k_{EC}$ for each period representative for a three months period, L. Curuaí outgassed 2.3 Tg C yr$^{-1}$ to the atmosphere.

5. Discussion

Open-water surficial concentrations of CO$_2$, and hence gas evasion, vary with the phase of the Amazon River and distance from shore. Seasonal stratification is likely to develop over pelagic regions during high water [Melack and Fisher, 1983; MacIntyre and Melack, 1984, 1988] and, hence, limit mixing of hypolimnetic water enriched in CO$_2$ with depleted water in the upper mixed layer. This effect helps to explain the low values observed. In addition, phytoplankton became abundant during high and receding water stages [Barbosa et al., 2010]. As a result of the increased photosynthesis, pH increased with concomitant decreases in CO$_2$. For example, during receding waters at one offshore station, CO$_2$ concentration was 0.3 µM, chlorophyll-$a$ concentration was 338 µg L$^{-1}$, and pH was 9.2. CO$_2$ concentrations were higher in littoral regions likely due to increased root respiration associated with floating macrophytes [Hamilton et al., 1995] and decomposition of grasses in the aquatic-terrestrial transition zone during rising water. Flooding of land containing excreta from buffalo and cattle, commonly grazed in this area [Sheikh et al., 2006], also contributed labile organic matter.

CO$_2$ concentrations in the Amazon R. remained at least two-fold higher than lake averages in each of the hydrological phases and, therefore, concentrations in the lake increased where and when river inflow occurred. However, since samples were collected from surficial water during mid-day heating, lake values are likely to be lower than at other times of day [Crill et al., 1988]. In particular, higher surficial CO$_2$ concentrations would be expected during
nighttime due to deep convective mixing. Turbulence in river flow results in less vertical variability.

*Richey et al. [2002]* observed CO$_2$ concentrations in the Amazon R. floodplain ranging from 97 to 1,445 µM. Concentration of CO$_2$ in the floodplain lakes decreased toward downstream reaches, averaging 463±335 µM in the upstream reach between Vargen Grande and Itapeua, where a scroll-bar topography produces mostly long narrow lakes, to 207±138 µM in the downstream reach between São José do Amatari and Óbidos, where a wide floodplain allows the dominance of large shallow lakes.

Our annual average estimate for L. Curuaí obtained from all samples was 172±212 µM and falls within the range reported for the downstream reach [Richey et al., 2002]. Longer residence times in larger lakes increases the chances of a more complete decay of organic matter within the lake before carbon is lost by outflow to the mainstem. Additionally, shallow lakes with larger open-water areas exposed to wind represent greater potential for aquatic CO$_2$ evasion. These attributes apply to large floodplain lakes that are more abundant in the downstream reach and may explain their lower mean CO$_2$ concentration. Given the high spatial variability in these downstream lakes, adequate spatial and temporal sampling is required to approximate the mean value of CO$_2$ concentration.

Few measurements of dissolved CO$_2$ are available from other lakes and wetlands in tropical South America. In shallow lakes of the Pantanal wetland, median dissolved CO$_2$ concentrations are about 100 µM with a range from near 0 to 280 µM [Hamilton et al., 1995]. In a shallow interfluvial wetland near the upper Negro River, dissolved CO$_2$ concentrations averaged 390 µM with standard deviation of 213 µM [Belger et al., 2010]. In Balbina Reservoir on the Uatumã R. in central Brazil, concentrations of dissolved CO$_2$ in surface waters ranged
from 42 to 180 μM over a year of sampling [Kemenes, 2006]. Though the upper Negro basin values average somewhat higher, these values from large lakes and a reservoir are similar to those in L. Curuáí.

Based on eddy covariance (EC) measurements Jonsson et al. [2008] have demonstrated that conservative estimates are produced using the widely adopted Cole and Caraco [1998] model and suggested another wind-based regression. In a reanalysis of the Jonsson et al. [2008] data, MacIntyre et al. [2010] demonstrated the additional dependence of $k_{600}$ on buoyancy flux and the applicability of the surface renewal model as a mechanistic approach to use in lakes. $k_{600,EC}$ computed in this study are almost twice as high as those computed following Cole and Caraco [1998]. Estimates of $k_{600,SR(zAML = z 0.5 m)}$ were essentially identical to $k_{600,EC}$, which implies that the coefficient $c_1$ we used applies for the case of a shallow mixing depth. In the absence of actual $z_{AML}$ measurements, our analysis demonstrates the approximate seasonal ranges for $k_{600,SR}$ are likely to vary for L. Curuai: 50% during high water, when $z_{AML}$ undergoes larger diel variations; 30% during midstage; and 4% during low water. However, additional effort is required to understand controls on the coefficient, $c_1$.

The gas exchange coefficients based on Cole and Caraco’s [1998] wind-based model and those computed following MacIntyre et al. [2010] are two and four times higher, respectively, than the conservative value ($k = 2.7 ± 1.0 \text{ cm hr}^{-1}$) used for regional assessments of floodplain environments [Richey et al., 2002]. Therefore, it is likely that outgassing of CO$_2$ from lakes in the lower Amazon floodplain is at least two to four times higher than previously estimated. Richey et al. [2002] estimated a rate of outgassing of CO$_2$ from rivers and wetlands of the central Amazon basin equal to 8.4±2.4 Mg C ha$^{-1}$ yr$^{-1}$. We estimate L. Curuáí outgassed from its open-water environment to the atmosphere $2.3±1.1$ Tg C yr$^{-1}$ or $18.4±7.9$ Mg C ha$^{-1}$ yr$^{-1}$. Estimates
would be even higher if the increased concentrations at the air-water interface resulting from
nocturnal mixing \([Crill et al., 1988; Aberg et al., 2010; Laurion et al., 2010]\) were included.

The annual water level oscillations affect the dynamics of CO\(_2\) evasion from Amazon
floodplains, and predictable biological response to fairly regular hydrological seasonality is
expected. However, inter-year variation in river discharge and intra-year variability in seasonal
timing of peak and minimum stages may cause variations not captured by a one year study. The
use of autonomous CO\(_2\) sensors \([e.g., Lynch et al., 2010]\) and thermistor chains would produce
valuable time series data and detect important episodic events for further understanding of CO\(_2\)
emission from the Amazon floodplain. Additionally, this study focused on the open-water
environments of L. Curuá, which accounts for about 72 \%, 63 \%, and 65 \% of the total flooded
area of this floodplain during low, medium (receding and rising), and high water stages,
respectively (Table 1). In order to appraise the total aquatic CO\(_2\) evasion of this floodplain,
additional adequate measurements would need to be undertaken in the vegetated flooded
environments.

6. Conclusions

The open-water environment of large floodplain lakes has high spatial variability in CO\(_2\)
concentration with higher concentrations in the littoral and at locations influenced by incoming
Amazon River water. Based on results from this study, CO\(_2\) evasion from large lakes in the lower
Amazon floodplain will be at least two to four times higher than estimated in current regional
calculations. The effects of turbulence induced by wind-mixing and convective cooling,
morphology, water level change, habitat type and habitat inundation status, and lake metabolism
are all important factors that determine the variations on biogenic gas production and emission
rates. On-going development of coupled models of lake inundation hydrology, hydrodynamics
and biogeochemistry will offer the spatial-temporal resolution needed for resolving carbon dioxide emissions, and the other fluxes of carbon within the floodplain ecosystem.

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References


Butler, J. N. (1992), *Carbon Dioxide Equilibria and Their Applications*, Addison-Wesley, Reading, MA.


Table 1. Morphological variables during hydrological phases: flooded area \( (A_{\text{flooded}}) \); open-water area \( (A_{\text{open-water}}) \); mean depth \( (z_{\text{mean}}) \) within open-water; maximum depth \( (z_{\text{max}}) \). Standard deviations are in parenthesis.

<table>
<thead>
<tr>
<th>Hydrological phase</th>
<th>High</th>
<th>Receding/Rising</th>
<th>Low</th>
</tr>
</thead>
<tbody>
<tr>
<td>( A_{\text{flooded}} ) (km(^2))</td>
<td>2274</td>
<td>2015</td>
<td>850</td>
</tr>
<tr>
<td>( A_{\text{open-water}} ) ** (km(^2))</td>
<td>1479</td>
<td>1278</td>
<td>612</td>
</tr>
<tr>
<td>( z_{\text{mean}} ) (m)</td>
<td>3.2 (2.5)</td>
<td>1.6 (1.5)</td>
<td>0.6 (0.4)</td>
</tr>
<tr>
<td>( z_{\text{max}} ) (m)</td>
<td>15.0</td>
<td>12.4</td>
<td>9.7</td>
</tr>
</tbody>
</table>

* Estimated from digital elevation model obtained by integrated interpolation of bathymetry and Shuttle Radar Topography Mission data (C. Barbosa and C. Rudorff, unpublished data) and water stage gauged at Curuáí.

** Estimated from classified Landsat images (C. Barbosa and C. Rudorff, unpublished data).

Table 2. Field campaign dates and number of samples. Mean transparency from Secchi disk \((SD)\), light attenuation coefficient \((k_d)\), pH, dissolved inorganic carbon \((DIC)\), dissolved \( CO_2 \) concentration in surface waters \((CO_{2aq})\), and dissolved \( CO_2 \) concentration in equilibrium with the atmosphere \((CO_{2eq})\). Standard deviations are in parenthesis.

<table>
<thead>
<tr>
<th>Hydrological phase</th>
<th>Receding</th>
<th>Low</th>
<th>Rising</th>
<th>High</th>
</tr>
</thead>
<tbody>
<tr>
<td>Samples (n)</td>
<td>71</td>
<td>73</td>
<td>72</td>
<td>74</td>
</tr>
<tr>
<td>( SD ) (m)</td>
<td>0.23 (0.11)</td>
<td>0.08 (0.04)</td>
<td>0.01 (0.03)</td>
<td>0.65 (0.05)</td>
</tr>
<tr>
<td>( k_d ) ((m^{-1}))</td>
<td>6.6 (2.7)</td>
<td>18.1 (4)</td>
<td>14.7 (6.3)</td>
<td>2.0 (0.2)</td>
</tr>
<tr>
<td>pH</td>
<td>7.68 (0.75)</td>
<td>6.69 (0.58)</td>
<td>7.25 (0.33)</td>
<td>7.51 (0.76)</td>
</tr>
<tr>
<td>DIC (ppm)</td>
<td>11.8 (2.7)</td>
<td>7.5 (4.0)</td>
<td>16.6 (4.1)</td>
<td>32.3 (3.6)</td>
</tr>
<tr>
<td>( CO_{2aq} ) (µM)</td>
<td>92.2 (114.2)</td>
<td>125.2 (34)</td>
<td>148 (85.2)</td>
<td>313.9 (354.3)</td>
</tr>
<tr>
<td>( CO_{2eq} ) (µM)</td>
<td>10.5 (0.4)</td>
<td>10.8 (0.2)</td>
<td>11.1 (0.4)</td>
<td>11.2 (0.2)</td>
</tr>
</tbody>
</table>

* Estimated using the conversion factor for turbid lakes \( k_d = 1.3 / SD \) [Koenings and Emundson, 1991].
Figure legends

Figure 1. Map of the study site and location of meteorological buoy.

Figure 2. Seasonal variation in CO$_2$ mean concentration for four hydrological phases in Lake Curuaí and Amazon River. Mean and 2003-2004 annual stage hydrograph of the Amazon River at Óbidos, Pará, Brazil.

Figure 3. (a-d) Lake bathymetry for four hydrological phases. Images are masked (gray) according to inundation extent. Spatial distribution of dissolved concentration of CO$_2$ at (e) receding (Sep 2003), (f) low (Nov 2003), (g) rising (Feb 2004), and (h) high water (Jun 2004). Samples were interpolated using a biharmonic spline [Sandwell, 1987]. Images are masked according to open-water extent obtained from Landsat image classification. Dark gray regions are differences between inundation and open-water areas and represent flooded vegetation.

Figure 4. Hourly average values of (a-d) Wind speed at 10 m, (e-h) air and 1 m water temperatures, (i-l) net shortwave radiation (400-1100 nm), (m-p) sensible heat, latent heat, and longwave radiation fluxes, (q-t) surface heat flux and effective heat flux, assuming active mixing layer depth equivalent to 0.5 m ($z_{AML} = 0.5$ m) and the seasonal lake mean depth ($z_{AML} = z_{mean}$).

Figure 5. Gas transfer coefficients for (a) high, (b) receding, (c) low, and (d) rising water, displayed as 3 hour averages, calculated using: wind-based model following Cole and Caraco [1998] (C&C), a wind and buoyancy flux based model (EC), and surface renewal model following MacIntyre et al. [2010] assuming active mixing layer depth equivalent to 0.5 m (SR ($z_{AML} = 0.5$ m)) and the seasonal lake mean depth (SR($z_{AML} = z_{mean}$)).

Figure 6. (a) Mean exchange coefficients for CO$_2$ in Lake Curuaí. (b) Estimates of CO$_2$ evasion from lake computed using mean values of CO$_2$ departure from saturation and gas transfer
coefficient. (c) Mean lake seasonal fluxes accounting for the variation in open-water surface area for the four hydrological phases. Legends for a and c are the same as presented in b.