Flooding hydrology and mixture dynamics of lake water derived from multiple sources in an Amazon floodplain lake

Lance F. W. Lesack

Departments of Geography and Biological Sciences, Simon Fraser University, Burnaby
British Columbia, Canada

John M. Melack

Department of Biological Sciences and Marine Science Institute, University of California, Santa Barbara

Abstract. The temporal dynamics in the lake water mixture derived from river flooding, local rainfall and runoff, exchange with an adjacent lake, groundwater exchange, and evaporation was computed from detailed measurements over an annual cycle for a permanent lake on the central Amazon floodplain. River water invaded the lake at the start of rising water, but by mid-rising water, lake water steadily flowed out from the lake and into the river, while river levels continued to rise. Despite a relatively low ratio of catchment area to water surface area, the lake exported to the river three times the amount of water originally received from the river. The lake water mixture was dominated by river water early during the water year, increasing to 70% of the mixture. When lake water began flowing out from the lake, the river water fraction was steadily diluted as runoff became dominant. By the end of the water year, runoff contributed 57% of the total water input, river inflow 21%, rainfall 11%, inflow from an adjacent lake 6%, and seepage inflow 4%. However, local runoff, river inflow, and lake water carried over between water years have the potential for considerable interannual variation. The finding that runoff from relatively small local catchments can be sufficiently large to prevent flood waters from entering the lakes during periods of rising water may critically limit our ability to characterize the long-term frequency and duration of flooding in lakes on the floodplains of large rivers in the wet tropics.

Introduction

Numerous economic justifications can be made for studying large rivers or the lakes associated with the floodplains of large rivers [see *Welcomme*, 1979]. Consequently, our understanding of the fundamental ecosystem processes operating in large rivers must be improved, because projecting their ecological behavior from that of small rivers is inadequate [*Richey et al.*, 1990; *Sedell et al.*, 1989]. Flooding and annual delivery of nutrient-rich sediments are commonly perceived to control the productivity of lacustrine and wetland ecosystems associated with the floodplains and deltas of major world rivers [*Junk et al.*, 1989; *Fisher and Parsley*, 1979]. As early as 1931, however, *Svensson* [1933] recognized during the course of his studies on fishes of the Gambia River, West Africa, that productive floodplain swamps could form either from backflooding by the river, accumulation of local rain and runoff, or a combination. Multiple sources of water may be a particularly important characteristic of river floodplains in the wet tropics. Yet to date, the mixture dynamics of waters on the floodplains of rivers in the wet tropics have remained virtually unstudied, despite the importance of understanding how hydrological and biogeochemical processes may interact to control the cycling of nutrients between large rivers and lakes within their floodplains and to control the productivity of the associated aquatic ecosystems.

The information available on the hydrologic characteristics of floodplain lakes has thus far consisted of only three studies in tropical locations and one study in an arctic location. First, among the studies in the tropics, *Hamilton and Lewis* [1987] conducted a solute mass balance study for Lake Tineo, a floodplain lake adjacent to the Orinoco River in Venezuela. Average rainfall at Lake Tineo is less than half the average rainfall found in wet tropical locations, such as the central Amazon, and the lake undergoes flushing and then an isolation phase during distinct wet and dry seasons. Since the catchment area of the lake is small, there are no streams draining into the lake, and groundwater was assumed to be negligible, the only sources of water to the lake are exchange with the adjacent river and rainfall onto the lake surface. Second, some work has been conducted on Lake Cameleão, a lake on an island (Ilha de Marchantaria) in the middle of the Solimões River located just upstream of the confluence of the Solimões and the Negro Rivers in the central Amazon basin [*Furch et al.*, 1983]. At high water the river floods across the entire island; at low water the lake may virtually dry up; hence the only sources of water are from the Solimões and rainfall. Third, significant limnological studies have been conducted on a series of lakes in the Paraná River floodplain in Argentina [*Carignan and Neiff*, 1992], where river flooding, local runoff, and groundwater exchange may all be important, but detailed information on the hydrology of these lakes is not thus far available. Fourth, representing an arctic location, a substantial amount of work has been conducted on a series of lakes in the Mackenzie River delta. This work has included careful measurements of river flooding, exchange with adjacent lakes, local runoff, rain and snowfall

Paper number 94WR02271.
0043-1397/95/94WR-02271$05.00

inputs, and evaporation [*Marsh*, 1986; *Marsh and Bigras*, 1988]. However, river flooding completely dominates the water budgets in these lakes because of very low amounts of local precipitation, and it is rare to find a situation where lakes contain water from more than one dominant source for a significant duration of time [see *Marsh and Hey*, 1989].

Although it was not a direct evaluation of the hydrologic characteristics of floodplain lakes, *Forsberg et al.* [1988] conducted an analysis that attempted to discriminate the influence of water originating from the main stem Amazon on the nutrient concentrations in 51 lakes that were sampled as part of a study of carbon flux in the Amazon [see *Richey et al.*, 1990]. On the assumptions that HCO_3^- is a conservative tracer, that the major source of HCO_3^- is weathering of sedimentary rocks in the Andes, and that sources within the lakes are negligible, they concluded that lakes with a catchment area (BA) to water surface (LA) ratio of less than 20 are expected to have a mixture of river water and local water during the annual low-water period, while lakes with a ratio greater than 20 are expected to be primarily local water.

As part of a study of the biogeochemistry of the central Amazon floodplain [*Lesack*, 1988; *Melack and Fisher*, 1990; *Engle and Melack*, 1993], the sources of water input and the export of lake water were measured over an annual cycle in Lake Calado, a permanent floodplain lake in the central Amazon. The measurements were subsequently used to develop an empirically derived model of the water balance for the lake with components that include river flooding, local rainfall and runoff, exchange with adjacent lakes, groundwater exchange, and evaporation. A detailed evaluation of the exchange of water between the lake and the surrounding groundwater system is reported by *Lesack* [1994]. We present here the water balance and hydrologic characteristics of Lake Calado, with emphasis on the mixture dynamics of the lake water sources over the annual cycle. We evaluate the potential for interannual variation in the dynamics during years that are wetter or dryer than normal. These results represent the first detailed analysis available for floodplain lakes associated with large tropical rivers.

Study Area

The site of measurement was Lake Calado (3°15′S, 60°34′W), a floodplain lake in the central Amazon basin, which is located about 80 km upriver from the confluence of the Solimões (main stem) and Negro Rivers. The lake is connected to the Solimões River year-round, and a significant portion of the network of rainforest streams draining into the lake submerges and reemerges as lake levels rise and fall in response to a combination of river flooding and accumulations of local runoff over the annual cycle (Figure 1). For a portion of the year Lake Calado becomes connected to two adjacent lakes: to Lake Miriti for a significant period during mid- to high-water conditions and to Lake Paru for only a brief period at high water (Figure 2). The interface between local terra firme drainage, which is never flooded, and the Solimões floodplain consists of a transition zone where flood waters can extend beyond what might be considered the geomorphological limit of the floodplain as reflected by sediment composition [see *Junk et al.*, 1989]. The floodplain levee, for example, is primarily depositional silts and clays. The portion of the lake which permanently contains water (about 2 km^2) and which lies between the levee and the transition zone is mainly depositional clays and

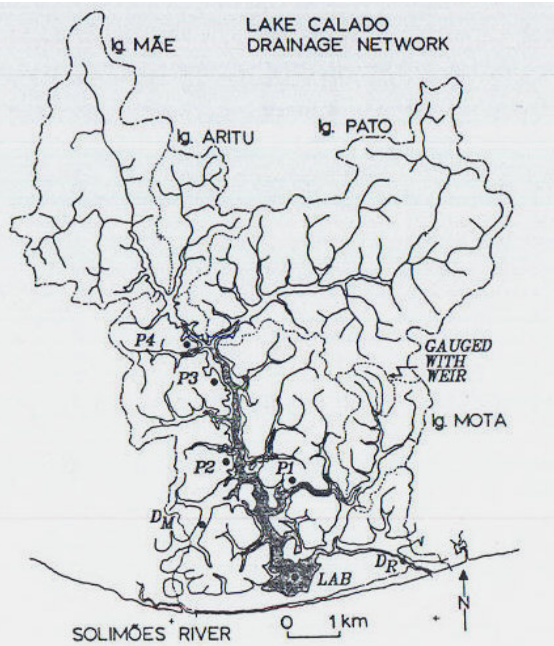

Figure 1. Map of the network of streams draining into Lake Calado. The lake is shown as a low- and high-water outline to indicate the extent to which the lower reaches of the streams become submerged as the water level in the lake changes. The broken line denotes the drainage divide for the entire catchment of the lake, while the dotted line denotes subcatchments for the four largest stream systems draining into the lake. Also shown is the first-order subcatchment of the Igarape (Ig.) Mota stream system, which was gauged; the locations of nonrecording rain gauges (P1 to P4); and the stations within the channels connecting Lake Calado to Lake Miriti and to the Solimões River, where current velocities were measured with drogues (D_M and D_R, respectively).

organics. By contrast, stream valleys and hillslopes surrounding the main basin of the lake, which are in the transition zone (up to 8 km^2) and are flooded part of the year, have distinctly different soils, ranging from loamy clays to loamy sands, which are derived from local parent material and are deeply weathered [*Lesack*, 1988]. Texturally, these soils are similar to soils within the nonflooded area around Lake Calado which have been classified on the soil map published by *Projeto Radambrasil* [1978] as yellow latosols. The nonflooded soils probably fall within the oxisol category of the *U.S. Department of Agriculture* [1975] soil taxonomy system and are similar to the forest soils of Reserva Ducke, located to the north of the Negro River and about 25 km northeast of the city of Manaus, which previously have been studied by *Nortcliff and Thornes* [1978, 1981], *Nortcliff et al.* [1979], and *Brinkmann and Santos* [1973]. These deeply weathered soils situated beyond the depositional silts and clays immediately adjacent to the main stem river, in combination with a large annual change in lake stage, may represent conditions which are particularly conducive to relatively large exchanges of seepage between the lake and surrounding groundwater system [*Lesack*, 1994].

The central Amazon region has a wet tropical climate, but there is a distinct dryer period from June through November, in correspondence with the northward annual movement of the intertropical convergence zone. Mean annual rainfall at

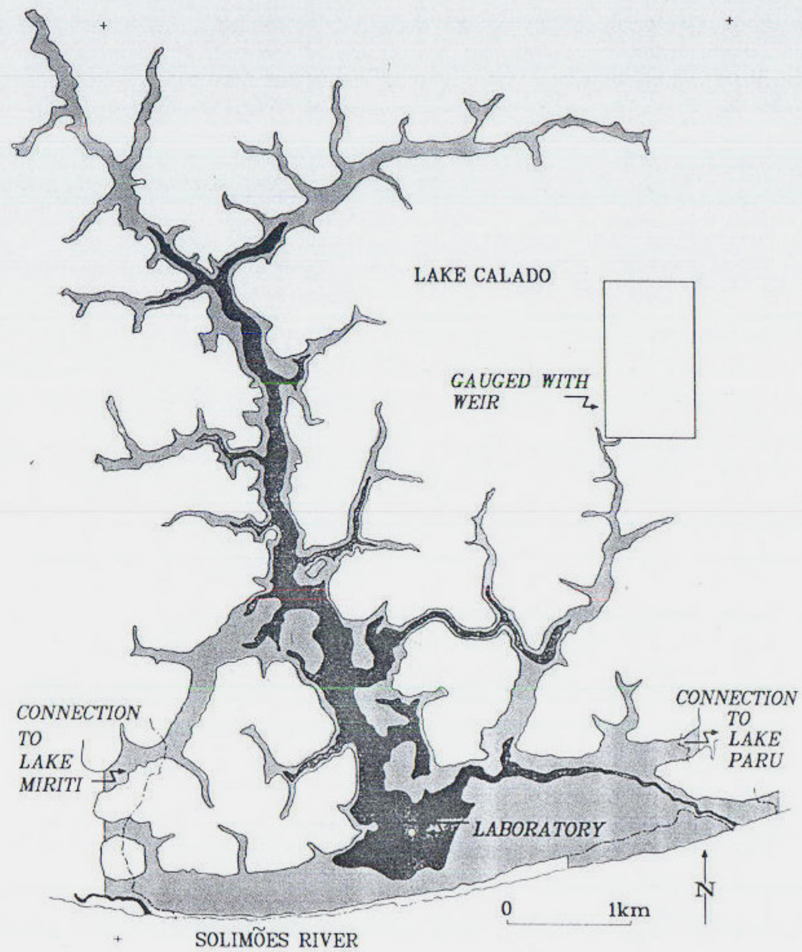

Figure 2. Map of Lake Calado showing greater detail of the high-water (light gray) and low-water (dark gray) outlines and where connection is made to Lake Miriti (inflowing) at mid- to high-water conditions and to Lake Paru (outflowing) at high water.

Reserva Ducke was 2410 mm over the period 1966–1983, with the 10 year return period for wet and dry years corresponding to rainfall of about 2800 mm and 2000 mm, respectively [*Lesack and Melack*, 1991]. Although long-term rainfall records are not available for the Lake Calado site, there are no compelling reasons to believe that its rainfall characteristics should be substantially different from those of Reserva Ducke.

More complete descriptions of the study area, climate, and hydrology of the rainforest streams draining into the lake and of the seasonal changes in seepage exchange between the lake and the surrounding groundwater system are provided by *Lesack* [1993] and *Lesack* [1994], respectively.

Methods

Components of the Water Balance

For any period of time, the water balance of a floodplain lake can be represented by the mass balance equation:

$$\delta S = P + R \pm L \pm H \pm G - E \qquad (1)$$

where δS is the change in the volume of water stored in the lake, P is rainfall onto the surface of the lake, R is inflowing surface runoff, L is exchange of water in either direction through connection to adjacent lakes, H is exchange of water in either direction through connection to the adjacent river channel, G is exchange of water with the surrounding groundwater system through seepage either in or out of the lake, and E is evaporation from the surface of the lake.

If a water year is defined as the period from low water to low water, the volume of water stored in the lake at any given time t during the year is

$$S(t) = S_I + \delta S(t) \qquad (2)$$

where the constant S_I is the initial volume of water in the lake at the beginning of the water year and $\delta S(t)$ is the change in lake volume from the beginning of the water year as a function of time. During the course of the year, each of the inflow and outflow quantities that constitute $\delta S(t)$, as defined in (1), also change as a function of time in response to temporal variations in weather and interactions with concurrent changes in the stage, surface area, and volume of the lake.

The volume of water contributed as rainfall onto the surface of the lake was calculated as

$$P(t) = \int Q_p(t) \cdot a(t)\, dt \tag{3}$$

where $Q_p(t)$ is the rate of rainfall per unit area and $a(t)$ is the area of the lake, both changing as functions of time. A tipping-bucket rain gauge mounted in the middle of the lake provided the temporal distribution of rainfall needed to determine $Q_p(t)$, while $a(t)$ was determined from daily measurements of lake stage and the known morphometry of the lake.

The volume of water contributed as runoff from the local catchment was calculated as

$$R(t) = \int Q_r(t) \cdot [A - a(t)]\, dt \tag{4}$$

where $Q_r(t)$ is the rate of runoff per unit area of drainage, changing as a function of time, and the constant A is the total catchment area of the lake. A calibrated weir and water level recorder installed in a 23.4-ha terra firme subcatchment, located a short distance above the maximum extent of lake flooding, provided a continuous record of the temporal distribution of runoff needed to determine $Q_r(t)$. The quantity $[A - a(t)]$ gives the effective drainage area within the catchment by subtracting the surface area of the lake as it changes as a function of time.

In the case of Lake Calado, connections are made to two adjacent lakes, but the volume is only significant in the connection to Lake Miriti, and the water movement is exclusively inflowing. The volume of water contributed from connection to Lake Miriti was calculated as

$$L(t) = \int u_L(t) \cdot x_L(t)\, dt \tag{5}$$

where $u_L(t)$ is the mean current velocity, $x_L(t)$ is the cross-sectional area in the connecting channel, and their product is the rate of discharge, which changes as a function of time. Daily measurements of lake stage height and known morphometry of the channel enabled $x_L(t)$ to be determined. Measurements of current velocity were fit with regression techniques against stage height to provide a solution for $u_L(t)$.

The volume of water seeping out of the lake and into temporary groundwater storage during rising lake levels and subsequently recovered during falling levels was calculated as

$$G(t) = \int\int Q_g(z, t) \cdot w(z, t)\, dz\, dt \tag{6}$$

where $Q_g(z, t)$ is the rate of seepage per unit area and $w(z, t)$ is the area of lake bottom, both of which change as functions of depth below the surface of the lake z and time. Empirical measurements of seepage rates were fit with multiple regression techniques against the combination of depth below the lake surface and rate of stage change to provide a solution for $Q_g(z, t)$, while $w(z, t)$ was determined from daily measurements of stage and the known bathymetry of the lake.

The volume of water lost from the lake as evaporation was calculated as

$$E(t) = \int Q_e(t) \cdot a(t)\, dt \tag{7}$$

where $Q_e(t)$, the evaporation rate per unit area, was assumed to change over the seasonal cycle as described by *Shuttleworth* [1988].

In the case of Lake Calado, there is a permanent channelized connection to the Solimões River, with water movement occurring in either direction depending on the differences in the rates of change in water level that occur independently in the river and in the lake. Since the volume of water stored in the lake $S(t)$ is known for any given time from the bathymetry of the lake and daily measurements of stage height, the volume of water exchanged with the Solimões River was calculated by rearranging (1) and (2) as

$$H(t) = S(t) - S_I - P(t) - R(t) - L(t) - G(t) + E(t) \tag{8}$$

This is the only term in the water balance that was not independently estimated and therefore also contains the residual error from each of the other terms.

To evaluate the role of the different sources of water in the overall water balance of the lake and the net effect on the transport of water down the adjacent river channel, two coefficients of water retention were defined. The exchange of water between the lake and the Solimões River actually consists of an inflow and outflow component:

$$H(t) = H_{\text{in}}(t) - H_{\text{out}}(t) \tag{9}$$

where $H_{\text{in}}(t)$ is the volume of inflowing water from the Solimões River and $H_{\text{out}}(t)$ is the volume of outflowing lake water that is being exported to the Solimões River. The annual retention of water originating from the Solimões River would hence be the value of $H(t)$ at the end of the water year and can be defined as the quantity H_F. From (9) and (1), the total annual input of water to the lake from all sources can be defined as

$$W_{F\text{in}} = P_F + R_F + L_F + H_{F\text{in}} + G_F \tag{10}$$

and the total annual outflow of water from the lake can be defined as

$$W_{F\text{out}} = H_{F\text{out}} + E_F \tag{11}$$

which in the present case is limited to evaporation and export of water to the Solimões River. In the more general case, a substantial volume of water could also be exported to adjacent lakes. The annual retention of water from all sources can hence be defined as

$$\delta S_F = W_{F\text{in}} - W_{F\text{out}} \tag{12}$$

and is equivalent to the value of (1) at the end of the water year.

Measurement Techniques

Lake bathymetry and lakewater storage. The outline of Lake Calado was prepared from 1:30,000 scale stereo pairs of black and white aerial photos taken during 1955. The images were interpreted and drafted as 1:7500 scale projections through a Bausch and Lomb stereo zoom transfer scope. Reference elevations were established in four ways. First, the 1955 photos established boundaries for the edge of the lake at the time that the photos were taken (taken to be 3 m deep in the south basin). From the stereo image, the boundary of normal high water (taken to be 11 m deep in the south basin) was estimated from the change in shoreline slope and vegetation type. Second, a set of oblique aerial photos taken from a low-flying aircraft during October 1983 established virtually complete coverage of the south basin, most mouths of the major streams, and other reference points at low water (depth <1 m in the south basin). Additional oblique photos were taken when the lake was 4.5 m deep (September 1984) and at normal high water (11 m deep; June 1984). Third, depth measurements were made to establish transects of known elevations. One continuous transect was measured along the mid-longitudinal axis of the lake from the middle of the south basin to the end of the lake. Four transects were measured across the long axis north of the south basin. Nine transects were measured across the south basin. Topographic ground surveys established the height of reference points such as the levee and high-water marks along shoreline. A longitudinal traverse grid was surveyed during low water along a representative stream channel to link the elevation of the lake water edge to the high-water mark and to the elevation of the catchment divide in the terra firme drainage. Fourth, during high water, point depth soundings were made at the mouths of all major streams and connections to neighboring lakes and to determine the height of all islands that are submerged for part of the year. Several types of interpolation were performed between reference points of transects of known elevation, including curve fitting, equal increment calculations, and interpolations by eye under the stereo transfer scope. More complete details on map production are provided by *Lesack* [1988].

To obtain areas and derive volumes for different portions of the lake that could be used for various purposes, the areas within each depth contour were determined by planimetry. Since the level of the lake undergoes such large seasonal fluctuation, the hypsometric relations were defined in terms of stage height rather than depth contours or elevations. Stage height was defined as the height of the water surface above the lake bottom in the center of the southern basin of the lake, which conveniently corresponded to a bank-full height of 11 m. Fitting a 3° polynomial between lake volume and stage and between surface area and stage provided very tight ($R^2 > 0.999$) and highly significant relations that were critically needed to provide estimates of the volume of water stored in the lake at any given time $S(t)$. The area of the lake as a function of stage height s is

$$a(s) = 0.00156 \cdot s^3 - 0.06410 \cdot s^2 + 1.17982 \cdot s + 0.75958 \tag{13}$$

and the volume of the lake as a function of stage is

$$S(s) = -0.01279 \cdot s^3 + 0.53139 \cdot s^2 + 0.87900 \cdot s + 0.41646 \tag{14}$$

where lake area is expressed in square kilometers, lake volume in 10^6 m^3, and stage in meters above the lake bottom.

Local rainfall, runoff, and evaporation. To obtain the temporal distribution of rainfall onto the surface of the lake, a tipping-bucket rain gauge was maintained on a floating laboratory anchored in the middle of the southern basin of the lake (further details provided by *Lesack and Melack* [1991]). To determine whether there was appreciable spatial variability in rainfall, especially with respect to distance from the Solimões River, a transect of four nonrecording gauges was installed up the longitudinal axis of the lake (Figure 1). A replicate pair of nonrecording gauges and a tipping-bucket rain gauge were also installed at the gauged terra firme subcatchment (further details provided by *Lesack* [1993]). The orifice of all nonrecording gauges was 10 cm in diameter. They were installed at standard height above the ground and within forest clearings of sufficient size to prevent interception or shadowing affects [*Hamon et al.*, 1979].

To obtain rates of unit runoff (Q_r), a 23.4-ha first-order subcatchment located a short distance above the maximum extent of lake flooding, and draining undisturbed terra firme rainforest, was gauged by using a calibrated V-notch weir and water level recorder (Figure 1). The drainage area of this subcatchment was determined by extensive ground survey and interpretation of the available 1:30,000 scale stereo aerial photographs. More complete details of the methods and hydrological techniques applied during investigations of the terra firme subcatchment are provided by *Lesack* [1993]. The drainage divide for the complete catchment of the lake was mapped, using a stereo zoom transfer scope, from the same set of 1:30,000 scale stereo aerial photos that were used in obtaining the lake bathymetry. The drainage area of the lake (A), as determined from planimetry of the 1:30,000 scale map that was produced from the photos, is 56.1 km^2.

Stream flow measurements at the weir site ran from February 21, 1984, to February 24, 1985. However, values for the low-water period from November 1, 1983, to February 20, 1984, were needed to provide a complete set of matching hydrological measurements that could be used in the water balance calculations for the 1983–1984 water year. To fill the missing period, values for the corresponding days during 1984 and 1985 were taken, and a constant was subtracted to properly position the spliced curve. The constant was determined as the value which lowered the curve so that the lowest part of the curve was made equal to 0.29 mm of runoff per day. The value 0.29 mm was chosen because it corresponded to a discharge measurement at the weir on December 10, 1983 (47 L/min). This was the approximate date of the annual minimum and corresponds well with the minimum during 1984, which occurred on December 15. The extrapolated curve is reasonably representative of what actually occurred during late 1983. On the basis of uncalibrated stage heights at the weir between mid-December 1983 and February 1984, there was an initial rise and subsequent fall in base flow discharge during January, similar to what occurred during 1984. Further evidence that the extrapolation is reasonable is that the runoff values for February 20 and 21 are well aligned even though the criterion for adjusting the spliced curve was the annual minimum value, not the end value where the fitted curve connects to the real data.

Estimates of instantaneous evaporation rate Q_e from the surface of Lake Calado during January and February 1983 were provided by S. MacIntyre (University of California, Santa Barbara, personal communication, 1988). The estimates are based on the mass transfer method, derived from measurements of wind speed and the vapor pressure and temperature gradients between the air and lake surface [*Brutsaert*, 1982]. The instrumentation and field methods are given by *MacIntyre and Melack* [1984, 1988]. The temporal distribution of evaporation rates $Q_e(t)$ from the lake surface over the seasonal cycle was assumed to change as described by *Shuttleworth* [1988].

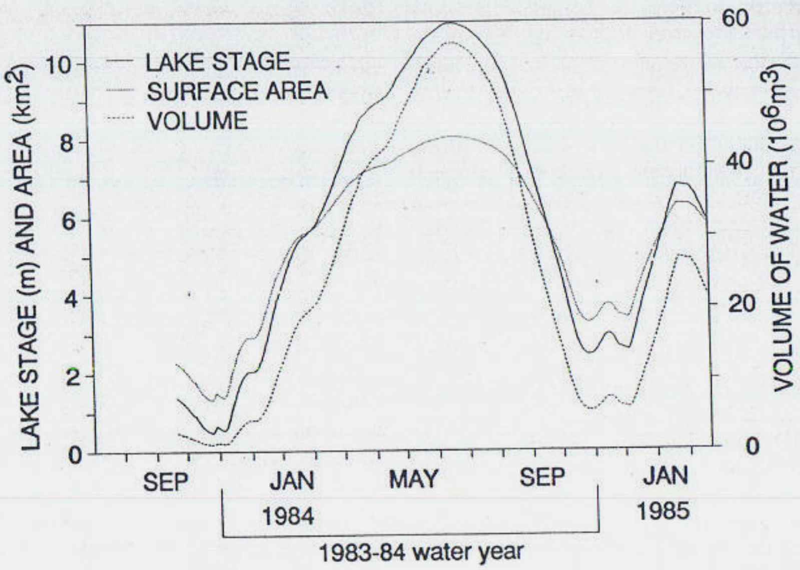

Figure 3. Stage height, surface area, and volume of Lake Calado from September 1983 to February 1985.

Lake seepage. Detailed information on the techniques employed to measure seepage rates (Q_g) and to estimate the flux of seepage exchange between the lake and the groundwater system (G) are provided by *Lesack* [1994]. Seepage between the lake and the Solimões River was considered to be negligible for water balance purposes because this would have to occur through the floodplain levee, which is composed of a mixture of silts and clays. Moreover, seepage meter measurements indicated that the rate of water movement into and out from these sediments was very slow relative to that for the other soils surrounding the lake. To evaluate the exchange of groundwater between the lake and the deeply weathered tropical soils adjacent to the shore along the main axis of the lake, seepage meters were installed at various representative locations during the falling-water periods of 1984 and 1985. The seepage rates encountered were sufficiently high that the seepage meters could be significantly smaller than the original design of *Lee* [1977]. The seepage meters were constructed from polycarbonate plastic water jugs 20 cm in diameter and had a sloping upper surface and neck which allowed gas to move into the collection bag that was attached during the period of measurement. Since the devices were transparent, it was possible to observe whether appreciable gas had become stuck inside the device and whether the installation had caused undue disturbance. Rubber bungs, with a small length of rigid plastic tubing running through them to serve as an outlet, were inserted into the necks of the jugs, and polyethylene bags were attached to the outlet tubes during periods of measurement.

The devices were typically installed 60 to 80 cm underwater, which was as deep as possible. Seepage measurements were taken each day until the water depth reached about 10 cm, at which point the devices began to emerge from the water. Since the rate of decline in lake stage that occurred during the period in which the measurements were made ranged from 5 to 15 cm/day, a sampling period consisting of at least 4 days of measurements was possible for each device, from the time it was installed until the time it had to be reinstalled. For each sampling period during which seepage measurements were made, 10 of the devices were installed at points around the southern basin of the lake. For initial analysis of the results [*Lesack*, 1994], each seepage measurement from each device (10 devices, one measurement each per day) from each sample period (several days long) was treated as a replicate for a total of 240 individual measurements during the period of study (falling-water periods during each of 1984 and 1985).

Flow measurement in connecting channels. Measurements of current velocity in the channels which connect Lake Calado to Lake Miriti and to the Solimões River (Figure 1) were obtained with light-weight drogues. The drogues were V-shaped, 20 cm in height, and each was suspended from a polystyrene float (12-cm diameter) with line of appropriate length to reach the depth of interest within the water column. To correct for the integrated drag of the float and line when current velocities at depths above the drogue were appreciably different, drogues at each particular depth were paired with a float and weighted line that had no drogue attached to it. Current measurements were typically performed in replicate at depths of 0.25, 0.5, and 1.0 m and at 1-m intervals thereafter, depending on the depth of the water. Measurements were limited to periods when there was little wind available to generate waves and surface currents.

Results and Discussion

Temporal changes in stage height and the volume and surface area of Lake Calado during the period of study are shown in Figure 3. The water year used in the water balance calculations was previously defined as the period from low water to low water, starting from the beginning of November 1983 through the end of October 1984. Maximum stage was reached during late June, and over the course of the water year the stage height ranged from <1 to 11 m, the surface area of the lake ranged from <2 to 8 km^2, and the lake volume ranged from $<1 \times 10^6$ to 57×10^6 m^3.

Table 1. Amount of Rainfall at Various Locations in the Catchment of Lake Calado

Station	Rainfall, mm		
	July 16, 1984–February 24, 1985	April 12, 1984–February 19, 1985	February 15, 1984–February 24, 1985
Floating laboratory	1382	2022	2673
Gauged subcatchment	1463	2135	2870
P1		2175	
P2	1377	2091	2724
P3	1407	2129	2845
P4	1410		

Since all the rain gauges were not always in operation, three different periods that permit comparison are shown. These periods overlap but do not directly correspond to the period of the water balance calculations. Station positions are shown in Figure 1.

Water Flux Components

Rain. Measurements of rainfall, among the network of nonrecording gauges that were established in the catchment of Lake Calado, show little evidence of significant spatial variation (Table 1). Rainfall on the surface of the lake during the 1983–1984 water year was about 2590 mm. Compared with the 18-year record of rainfall at Reserva Ducke, this particular water year at Lake Calado would be considered in the range of normal to somewhat wet. The period from January through March 1984, however, was quite wet. From February 1984 to February 1985, which was the period defining the water budget for the gauged terra firme subcatchment, rainfall at the stream gauging station was 2870 mm [*Lesack*, 1993] (about equivalent to a wet year with a 10 year return period).

Runoff. Streamflow measured at the weir site from February 21, 1984, to February 24, 1985, was 1650 mm [*Lesack*, 1993]. The runoff derived for the water year utilized during the present study is about 1470 mm, and its distribution is shown in Figure 4. This quantity is comparable to values derived by *Forsberg et al.* [1988] (1460 mm) from measurements of discharge on a variety of different reaches and major tributaries of the main stem Amazon River over the period 1982 through 1984. Much lower runoff values (~400 to 500 mm) have been reported by *Franken and Leopoldo* [1984] for two other small catchments within the region (Barro Branco at Reserva Ducke and Bacia Modelo), but the rainfall amounts during their investigations were equivalent to two of the driest years among 18 years of rainfall records for Reserva Ducke. Detailed treatment of the hydrologic characteristics of the runoff and comparison with previous investigations are provided by *Lesack* [1993].

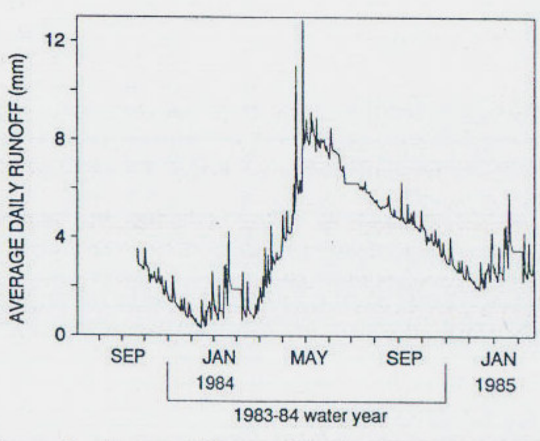

Figure 4. Average daily runoff during the period of study. The period from February 1984 to February 1985 is based on stream discharge measured at the weir, normalized to its upstream drainage area. Values for the period from October 1983 to February 1984 were extrapolated from the relative distribution observed during the same period during 1984–1985.

Connection to Lake Miriti. Connection between Lakes Calado and Miriti is established when the stage of Lake Calado reaches about 8 m. The connection was maintained during 1984 from about March 10 to August 28, during which time the stage of the lake ranged from 8 to 11 m. Measurements of current velocity and discharge were made in the connecting channel on 12 dates during the period of connection. *Lesack* [1988] found virtually no relation between discharge and stage height, but did find a roughly inverse linear relation between current velocity and stage height ($R^2 = 0.60$). Since it was possible to fit a relation between stage height and current velocity, and the approximate area of the gauging station is known, daily average current velocity and discharge were calculated for the period during which the connection existed (Figure 5). The depth in the center of the channel where the discharge measurements were performed was about 1 m deeper than the elevation at the point where the connection was first established. Hence, while the connection is maintained through only 3 m of range in stage, the channel depth at the gauging station ranged from about 1.4 through 4.3 m over the period when sustained flow was occurring. During the same period, the cross-sectional area of the channel ranged from about 2.7 to 10.1 m^2, while the width of the channel ranged from about 1.9 to 2.5 m. The discharge of water did not vary greatly, ranging from 0.36 to 0.62 m^3/s. Over the rising-water period, flow velocity declined by a factor of about 2.4 (0.134 to 0.056 m/s), while the cross-sectional area of the channel increased by a factor of about 3.7 (2.7 to 10.1 m^2). As a result, the discharge increased, but only by about 56% (0.36 to 0.56 m^3/s) over the period from when sustained flow first began in the channel until peak water level was reached. A consequence of the inverse relation between channel depth and flow velocity, in combination with the positive relation between channel depth and channel cross section, is that maximum discharge (0.62 m^3/s) occurred when the channel depth was approaching (3.38 m) but not at its maximum depth (4.33 m).

There are two relatively large sources of error associated with these discharge measurements. First, the effective cross-

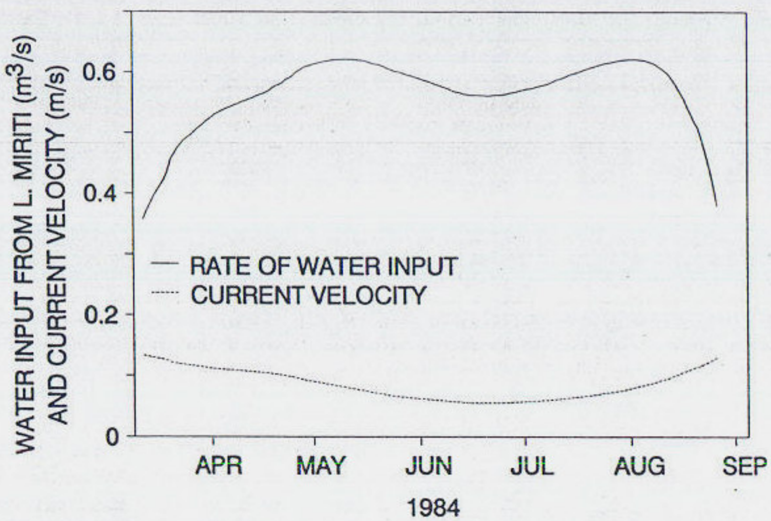

Figure 5. Estimated rate of water input from Lake Miriti to Lake Calado and the current velocity in the connecting channel during the period that the connection existed.

sectional area of the channel was probably underestimated because it was choked with vegetation. Second, the current velocities that were measured were probably overestimated. The currents were measured with a drogue suspended in the center of the channel at 0.25 m below the surface of the water, which is within the region where the maximum current velocities will be encountered. Because drogues were used to measure the currents, it was not possible to obtain lateral profiles of the current across the channel or to obtain measurements near the bottom without snagging the drogue on the submerged vegetation. Given that flow velocities were typically slow (4 to 10 cm/s) and that velocities measured with drogues may not necessarily represent true point velocities, we did not attempt to correct the measured velocities to a mean velocity that would be representative of the channel as a whole. The net result of overestimating current speeds and underestimating the channel cross section, however, is that the errors contributing to uncertainty in the estimated discharge rates should have a compensating tendency.

Groundwater. The evidence indicates that groundwater interacts with the lake in a rather complicated way, and a detailed treatment of these results has been provided by *Lesack* [1994]. Measurements of hydraulic head, derived from piezometer nests installed in the soil matrix within the near-shore margin of the lake, show that the water moves out of the lake during the rising-water period and back into the lake during the falling-water period. The results of the lake seepage model [*Lesack*, 1994], which is based on seepage meter measurements, indicate that about 1.5 times as much water seeped into the lake during falling water as originally seeped out the lake during rising water (Figure 6). This result is consistent with a conceptual model of water exchange between the lake and the surrounding groundwater system provided by *Lesack* [1994]. Moreover, the volume of seepage outflow over the rising-water period (4×10^6 m^3) is plausible if the rate of groundwater flow is constrained by the hydraulic conductivities expected for the soils.

Evaporation. The estimates of evaporation rate from the surface of the lake Q_e that were provided by S. MacIntyre (personal communication, 1988) spanned 13 days in January and 15 days in February of 1983. Instantaneous values ranged from 1.5 mm/day during cool windless periods at night to greater than 10 mm/day on bright windy days. The time-weighted average for all of the instantaneous values was 4.2 mm/day. Since the measurements were performed in the middle of the southern basin of the lake, where the average wind fetch is longer than for the rest of the lake, they should overestimate the spatially averaged rate for the lake. Conversely, since the measurements were performed during the middle of the rainy season, they probably underestimate the temporally averaged evaporation rates for the year. Furthermore, the seasonal influence of floating macrophytes and flooded forest on evaporation rates complicates the methods which may be used and the interpretation of the rates obtained.

A value of 4 mm/day was selected as an estimate for average Q_e in the water balance calculations for the lake. This is equivalent to an annual evaporation of 1460 mm and is comparable to annual rates of potential evapotranspiration from the surrounding rainforest that have been reported previously [see *Shuttleworth*, 1988; *Bruijnzeel*, 1990]. Given the potential for interaction between the spatial and temporal variance in Q_e and the temporal differences in the area of the lake, substantially more effort would be required to accurately estimate evaporative flux from the surface of the lake, $E(t)$ (equation (7)), over relatively short intervals of time (one to several months). As a first approximation for a budget over an annual period, however, the approach taken in this study was judged to be reasonable.

Connection to the Solimões River. The flux of water through the connection between the lake and the Solimões River is the most difficult component of the water balance to estimate because the current velocities are usually too slow and variable for conventional gauging techniques. The temporal distribution of water flux through the connection during the period of study (Figure 7), calculated as the water balance residual, clearly shows that the direction of flow should change from inflowing to outflowing in the middle of March. The profiles of current velocities in the channel connecting the lake to the river that were measured during this period (Table 2) confirm the timing of this change in flow direction. On March 15 the current velocities were slow and variable, but all were inflowing except for that at one depth which was indeterminate. On March 21 all of the measurements were indeterminate, but the color of the water in the channel had clearly changed to that of the lake water, which is visually distinguishable from that of the Solimões River. On April 2 the currents were clearly flowing out of the lake.

The water movement is driven by small differences in the rate of rise occurring between the lake and the Solimões River, and the data in Figure 7 imply that both the direction and discharge rate are quite variable. The exceptionally high spike during March appears to be the result of an anomalously high rate of rise in the Solimões River. The large spikes of water out of the lake during April are associated with conspicuously large amounts of rainfall directly onto the surface of the lake. Although the manner by which sources of water are discriminated in the water balance model is actually more sophisticated, the manner by which direct rainfall is able to overshadow volumes of storm flow entering the lake from streams during the same storms can be rationalized as follows. Detailed analysis of storm flow hydrographs from the stream draining the gauged subcatchment [*Lesack*, 1993] demonstrated that if all storm flow volumes are assumed to originate as saturation overland flow, the contributing drainage area needed to account for the volumes was at most only 4% of the catchment. More commonly, values were in the range of 1–3%. In the present case, the surface area of Lake Calado at high water is about 8 km^2, and the total drainage area (including the lake surface) is about 56 km^2. Hence the area contributing stream flow to the lake is about 48 km^2, but the area which could contribute saturation overland flow to the lake is only 4% (conservatively) of 48 km^2, or about 2 km^2. Thus if, for example, 100 mm of rainfall occurred during a storm, the volume of storm flow added to the lake by streams will be the product of 2 km^2 and 100 mm (after appropriate unit conversions). By contrast, the volume of rainfall added directly to the lake surface will be the product of 8 km^2 and 100 mm.

The amount of water potentially pushed out of the lake during a local storm (the magnitude of the outflow spikes in Figure 7) is also related to seasonal timing, since a given storm would not have the same effect in the water exchange calculation if it occurred when the surface area of the lake is smaller. Spatial effects may also be important, given that some of the storms can be somewhat localized when they occur, whereas the rain and runoff distributions that are being used in the model are each derived from only one location. However, there is probably a tendency for the very large magnitude storms to have a relatively uniform effect over the scale of the lake basin. Hence, while there may be some inaccuracy in the effect of individual storms, the longer term average should be representative. It seems clear that water movement in the channel is quite sensitive to large storms, unusually wet periods, and high rates of stage change in the river.

Temporal Dynamics of the Lake Water Mixture

To illustrate the importance of the various sources of water in the water balance and the degree to which the lake flushes at different times of the year, the water balance during the 1983–1984 water year is plotted in Figure 8. The water fluxes, which can be either inflowing or outflowing, and the volume of water stored in the lake are plotted temporally from the initial

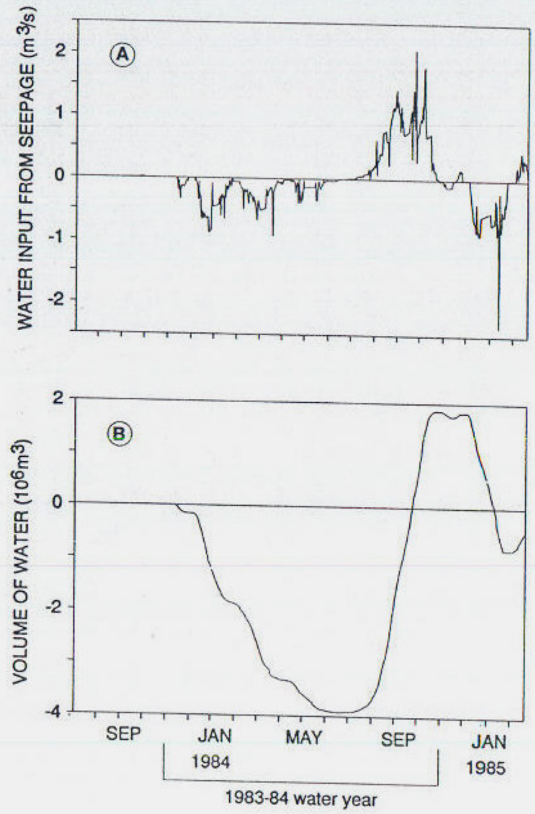

Figure 6. (a) Rate (dG/dt) of water exchange and (b) cumulative flux of water $G(t)$ exchanged between the lake and the surrounding groundwater system during the period of study.

rise in lake level on November 7, 1983, to the initial rise in water the following year on December 6, 1984. The plots clearly show that the two dominant components in the water balance are exchange of water with the Solimões River and local runoff.

From the beginning of the water year until the middle of March, the only removal of water from the lake was through evaporation, which is a permanent loss, and seepage into the groundwater system, which is recovered later in the year. Both of these fluxes are relatively small. Until the middle of March, there was a relatively steady inflow of water from the Solimões River. After the middle of March, water began to steadily flow out from the lake and into the Solimões River, while the level of the lake continued to rise in response to the other inflows. This is the time when the connection to Lake Miriti was first made and the inflow of water from Lake Miriti to Lake Calado first began. By the middle of July, the volume of water that originally entered the lake from the Solimões River was equal to the volume of lake water that had been exported to the Solimões River, and thereafter, the lake was making a net contribution of water to the river. By the middle of August, the volume of runoff that had entered the lake since the beginning of the water year was equal to the maximum volume that the lake had reached this year, which is also roughly equivalent to the bank-full volume (57.4×10^6 m^3) of the lake. By the end of the water year, the volume of runoff that had entered the lake had reached about 1.5 times the bank-full volume of the lake,

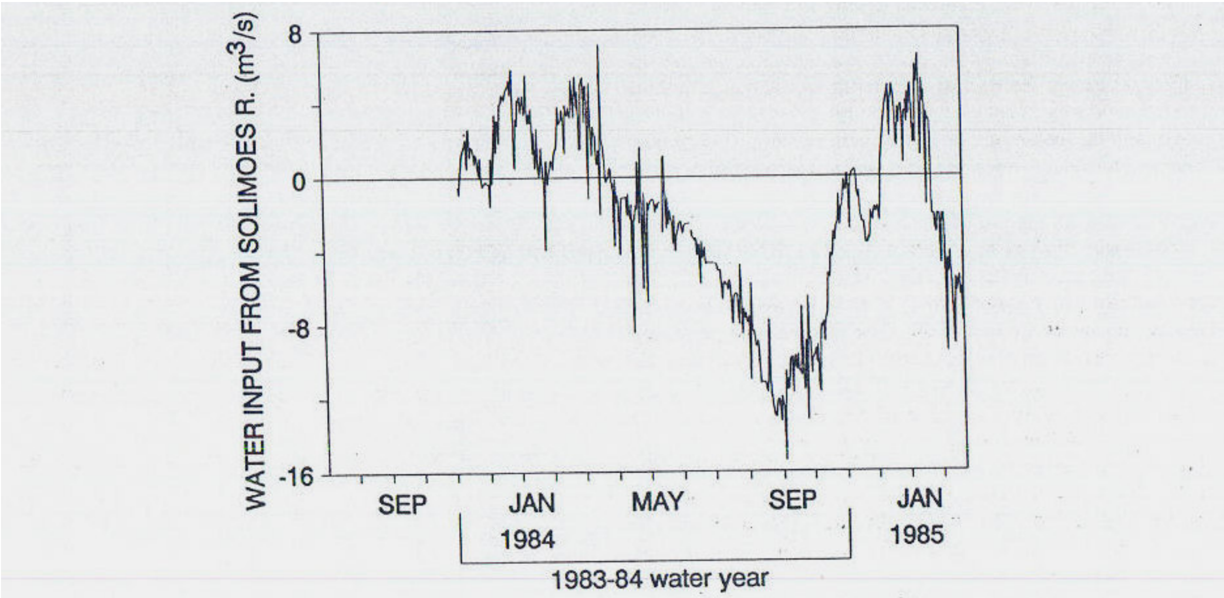

Figure 7. Rate of water input from the Solimões River into Lake Calado during the period of study.

and the lake had exported about 2 times its bank-full volume to the river (120×10^6 m^3).

Given that 27×10^6 m^3 of water from the Solimões River entered the lake from the beginning of the water year until the middle of March, when the direction of flow switched, the net annual export of water was about 90×10^6 m^3, or 3.3 times the volume of water that was originally received from the river. These calculations exclude the fact that some of the 8×10^6 m^3 of water received from Lake Miriti also originated from the Solimões River. Even if all of this water had come from the Solimões River, the amount would moderately increase the total volume received from the Solimões to 35×10^6 m^3, but this new volume would result in only a minor change in the ratio of the inflow volume to the volume exported.

For each of the components in the water balance, the water which is being transported is from a different source and will have a different chemical composition, depending on the direction of the flux. For example, in the case of the connection to the Solimões River, river water is being transported when the direction of the flux is into the lake, while lake water, which is a composite mixture of water from all sources, is being transported when the flux is out from the lake. Similarly, when seepage out of the lake occurs, the water is presumed to be the composite mixture of lake water, whereas when seepage into

Table 2. Measurements of Current Velocity in the Channel Connecting Lake Calado to the Solimões River

		Current Velocity, cm/s, at Depth From Surface:								
Date	Direction	0.25 m	0.5 m	1 m	2 m	3 m	4 m	5 m	6 m	7 m
1983										
Sept 22	Out	−13.4	−13.7	−14.2						
Sept 25	Out	−15.3	−14.9	−12.9						
Oct 9	Out	−40.3	−40.3							
Oct 17	Out	−35.5	−33.3							
Oct 27	Out	−39.3	−35.8							
Nov 7	In	4.7	2.9							
Nov 20	In	11.7	12.1	11.4						
Nov 27	In	12.9	12.8	12.3						
Dec 7	In	7.9	9.2	10.9						
Dec 15	In	12.6	12.2	11.6	10.4					
1984										
Jan 8	In	5.0	3.7	3.3	2.6	1.8				
Jan 19	In	···	···	···	7.6	8.0	7.6			
Feb 7	In and out	1.4	···	···	···	−1.6	−1.9	−1.6		
Feb 23	In	···	5.5	6.7	7.2	7.3	4.9	4.8		
March 15	In and out	7.1	4.3	3.3	···	1.8	1.9	3.0	2.9	
March 21	In and out	···	···	···	···	···	···	···	···	···
April 2	Out	···	···	···	−2.5	−5.0	−5.3	−7.6	−5.3	−6.4
April 13	Out	−5.7	−5.2	−3.6	−3.8	−3.2	−1.9	−2.2	−4.0	−5.0

Velocities were measured at 1-m depth intervals from the water surface with drogues. Three center dots indicate indeterminate data.

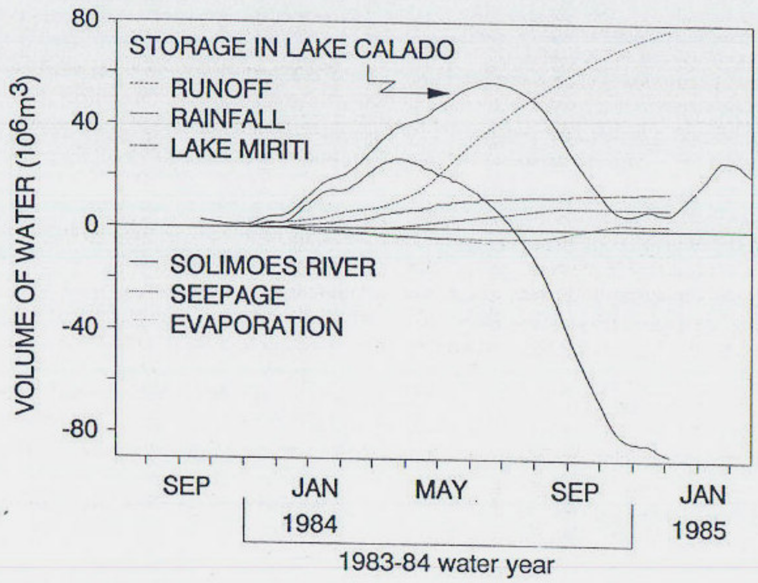

Figure 8. Components of the water balance for Lake Calado during the period of study. The input and outflow components are plotted as the cumulative flux of water relative to the start of the 1983–1984 water year at the beginning of November 1983. The plotted storage component $S(t)$ is the instantaneous sum of the input and outflow terms.

the lake occurs, it is presumed to have an identity of its own from mixing with groundwater derived from infiltration of rainfall, being subjected to evapotranspiration, and potentially undergoing geochemical diagenesis. To evaluate the mixture of water types present in the lake over the course of the water year, the fraction of the total input of water contributed by a given source since the beginning of the water year is plotted over the period of study in Figure 9. This representation is based on the assumption that only inflows change the mixture of the lake water, while export to the Solimões River, removal into temporary groundwater storage, and evaporation remove all the components of the mixture uniformly, without differentially altering the mixture.

Initially, the only water in the lake was the mixture from the end of the previous year, which in this particular year quickly became a small fraction of the mixture. Solimões River water

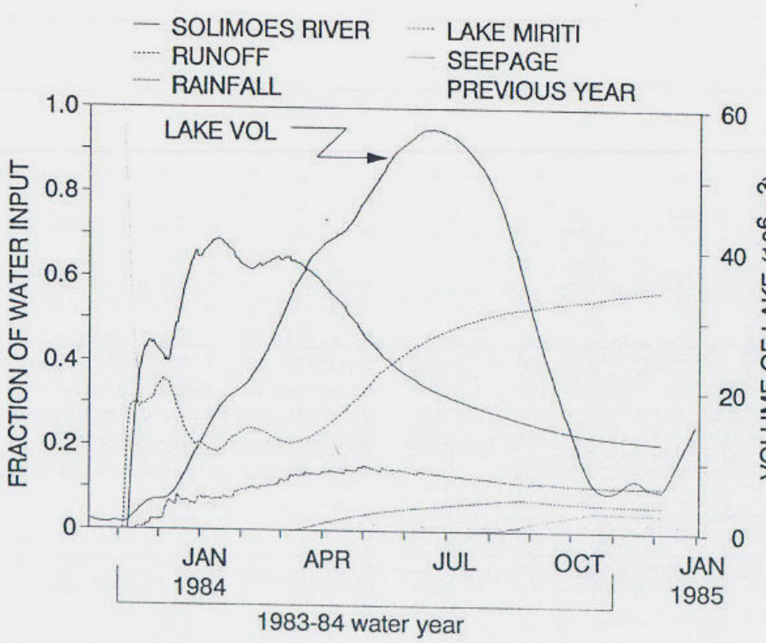

Figure 9. Fraction of total water input contributed by each source of water during the period of study. Also shown is the volume of the total water mixture in storage, as represented by the volume of the lake.

dominates the mixture during the early part of the year, increasing to as much as 70% of the mixture in January and February. As the direction of flow in the connection to the Solimões River reversed and lake water began flowing out of the lake near the end of March, the fraction of water in the lake contributed by runoff steadily increased, while the portion constituted by the Solimões River water declined as a result of being diluted. The runoff and Solimões River fractions became equivalent in the middle of May, when each constituted about 40%. By the end of the water year, runoff had contributed 57% of water input and the Solimões River 21%, and the remaining 22% was split among rainfall onto the surface of the lake (11%), inflow from Lake Miriti (6%), inflowing seepage (4%), and lake water carried over from the previous water year (less than 1%).

We have collected water samples to discriminate the stable isotope composition (^{18}O and ^{2}H) of the sources of water. However, the signatures are not sufficiently distinct over the time scales discussed in this paper (weeks to months) that they would be able to improve our analysis. We believe that a conservative tracer mixing model could potentially be useful for generalizing the results from a lake such as Calado, but only after it is validated with real hydrologic data. This is a complex undertaking which will be pursued in future research.

Sources of Error in the Water Balance

In evaluating the potential sources of error in the water balance for Lake Calado, we have drawn upon a review by *Winter* [1981] and case studies by *LaBaugh* [1985] and *LaBaugh and Winter* [1984]. Measurement of changes in lake storage is dependent on measurements of stage height and the accuracy of the bathymetric map. Stage height was derived from staff gauge readings that were continuously cross checked against bench marks established at 2-m intervals in elevation. The elevations of the bench marks were established with standard leveling techniques while the shoreline was exposed at low water. Given that the common rate of stage change in the lake is 5–10 cm/day, the estimates of change in stage over periods of weeks to months should have an associated error of less than 5%. A good quality bathymetric map should be in error by less than 5%. *Hakanson* [1978] has shown that the error in lake bathymetry increases substantially as the density of survey control points decreases. There were some problems in determining the high-water shoreline boundary for Lake Calado, but this problem occurs mostly in the lower reaches of stream channels, which become rias during high water. This area of flooded stream channel is not a very large portion of the total area of the lake. Furthermore, through most of the range in stage change, the interface between the hillslopes and water's edge, which forms most of the shoreline along the main basin of the lake, is relatively steep. The error in the estimates of change in lake volume over a several-meter range in stage may be as good as 5–7% and are likely no worse than 15%.

Rainfall measurements were usually obtained from rain gauges at five different sites around the lake and for some periods from six sites. Since the maximum area of the lake is about 8 km^2, this represents a sampling density of about 1.6 km^2/gauge. Moreover, at three of the sites replicate gauges were used, and at two of these three, recording rain gauges were also present. *Winter* [1981] cites estimates of error for daily areal mean rainfall of 10% at a gauge density of 21 km^2/gauge and of 4% at a density of 2.6 km^2/gauge. For point estimates of precipitation, error decreases as the length of time over which values are averaged increases. Winter cites errors of 10–20% for monthly means and about 5% for annual estimates. At Lake Calado the expected error in annual rainfall onto the surface of the lake would therefore be about 5%. Since the gauges were not shielded, the upper limit of the error estimate was taken as 15%.

Measurement of runoff with a calibrated weir and water level recorder is considered the best available method. Error can usually be limited to less than 5% [see *Lesack*, 1993]. On the basis of a study of long-term mean areal runoff in a temperate region, *Winter* [1981] cites an average error of 18% in regionalizing unit runoff values from gauged areas to similar ungauged areas. However, Winter also cites evidence that in most humid regions stream flow is closely related to the drainage area and the annual rainfall, presumably because soil water deficits do not develop and evapotranspiration is able to occur at near its full potential. Winter cites a study by *Riggs* [1973] which contends that the standard error of regressions relating drainage area and rainfall to stream discharge in humid areas is in the range of 10–15%. This implies that regionalization of unit runoff values in a rainforest may have less potential error than the 18% value that was reported in the temperate region study. The gauged subcatchment at Lake Calado was undisturbed forest, while significant forest area had been cut or was in regrowth in the ungauged greater catchment of the lake [*Lesack*, 1993]. Given also that the instantaneous surface area of the lake $a(t)$ was subtracted from the catchment area of the lake, the estimate of error expected in the runoff flux to the lake was taken as 15–30%.

The estimates of inflow from Lake Miriti are dependent on the measurements of lake stage, the accuracy of the relation between stage height and current velocity, and the relation between stage height and the cross-sectional area of the channel. The error in the measurement of stage height is small. The error associated with long-term estimates of discharge based on measurements of stage and a good rating curve, but without installation of a control structure, should be about 5–10%. In these situations the common sources of error in most rating curves are hysteresis effects as current velocities change rapidly during the passage of flood waves associated with storms. The water flux in the channel that connects Lake Miriti to Lake Calado did not appear to be appreciably influenced by storms. The current velocities were always relatively slow. There were, however, problems associated with estimating both the cross-sectional area of the channel and the mean current velocity. Moreover, the rating curve between stage and current velocity has a negative slope. The Manning equation shows that as the channel becomes deeper, current velocity should increase if channel roughness or the slope of the water surface does not change. Since the channel is choked with vegetation, the most likely implication is that roughness does change, although it is possible that the difference in elevation of the water surface between the lakes also changes. Given the above, the estimate of water flux over the period that the lakes were connected may be in error by 30–50%.

The energy budget, which is known to be the best method of estimating evaporation, has an associated error of about 10% over annual periods. The mass transfer method is based on an aerodynamic approach and empirical coefficients. Error can be limited to about 15% with this method when the coefficients are calibrated against the energy budget method. Given that the mass transfer equation used by S. MacIntyre (personal communication, 1988) to provide the estimates of average evaporation rate Q_e for Lake Calado was not independently

Table 3. Annual Flux of Each Water Balance Component and Range of Error Associated With Each

	Annual flux, 10^6 m^3	Lower Limit		Upper Limit	
		ξ, fraction	ξ, 10^6 m^3	ξ, fraction	ξ, 10^6 m^3
δS_{in}	56.02	0.06	3.36	0.15	8.40
δS_{out}	51.11	0.06	3.07	0.15	7.67
P	14.44	0.05	0.72	0.15	2.17
R	77.69	0.15	11.65	0.30	23.31
L	8.27	0.30	2.48	0.50	4.14
G_{in}	5.86	0.50	2.93	1.00	5.86
G_{out}	4.03	0.50	2.02	1.00	4.03
E	9.30	0.20	1.86	0.30	2.79
			Residual Terms		
H	−88.02	0.15	13.39	0.31	27.44
H_{in}	28.60				
H_{out}	−116.63				

Water balance components are defined in (15). The error assigned to each component (ξ), expressed as a fraction, was multiplied by the annual flux of the given component to convert the error to a volume of water. The residual of the water balance has been assigned to the exchange flux between the lake and the Solimões River, and the error associated with the residual was calculated from (16).

calibrated, that the seasonal distribution in evaporation rate $Q_e(t)$ was derived from measurements of monthly evaporation reported by *Shuttleworth* [1988], and that the evaporative flux $E(t)$ was actually the product of $Q_e(t)$ with the instantaneous surface area of the lake $a(t)$, the error estimate selected for error propagation calculations was 20–30%.

We are not aware of any studies that have been able to directly estimate the error associated with groundwater fluxes based on seepage meter measurements. *Winter* [1981] believes that even hydraulic conductivities cannot be estimated any closer than within about 50% and that in many cases the error is closer to 100%. In the present study, measurements of average hydraulic conductivity [*Lesack*, 1994] in the deeply weathered soils adjacent to the lake, which extend beyond the silt and clay deposits near the main stem river channel, ranged from 1.5×10^{-3} cm/s to 8.2×10^{-4} cm/s. These values are consistent with hydraulic conductivities expected for the lower range of clean sands and the upper range of silty sands [*Freeze and Cherry*, 1979] and with the expected magnitude of variability suggested by *Winter* [1981]. Moreover, *Lesack* [1994] has been able to demonstrate that the estimated volume of seepage outflow from Lake Calado to the surrounding soils over the rising-water period is plausible if constrained by this range of hydraulic conductivities. Consequently, the range of 50–100% was used as an estimate of uncertainty for the total seepage outflow from the lake [*Lesack*, 1994] and has subsequently been used in the error propagation calculations.

Water exchange between the Solimões River and the lake (H) is the only term in the water balance of the lake which was not independently estimated and represents the residual of the water balance as given by (8). This can be rewritten as

$$H = (\delta S_{in} - \delta S_{out}) - P - R - L - (G_{in} - G_{out}) + E \qquad (15)$$

where the changes in lake storage (δS) and groundwater (G) are each broken into inflow and outflow components. Assuming that the measurement errors associated with each term are independent and random, and given that each term in the water balance equation has been estimated with an uncertainty ξ and variance ξ^2, the error associated with the water balance residual H can be estimated by the quadratic sum of the expected errors for each of the measured terms [*Taylor*, 1982]:

$$\xi_H = [(\xi_{\delta S in})^2 + (\xi_{\delta S out})^2 + (\xi_P)^2 + (\xi_R)^2 + (\xi_L)^2 + (\xi_{G in})^2 + (\xi_{G out})^2 + (\xi_E)^2]^{1/2} \qquad (16)$$

The annual flux of each of the water balance components defined in (15), the range of associated error selected for propagation calculations, and the estimate of residual error as calculated from (16) for each end of the range of selected errors are compared in Table 3. The estimated error of the annual water balance residual is equivalent to 15–31% of the net water flux exchanged between the lake and the Solimões River. The assumption that the measurement errors among the lake water balance terms are independent and random is generally considered acceptable [*Winter*, 1981]. However, it is possible that the measurement errors associated with the δS_{in} and δS_{out} terms are systematic and correlated with each other, and likewise with the G_{in} and G_{out} terms, although we have no substantive reason to expect that this is the case. In the worst case, where these errors are fully correlated, the component errors would be additive rather than quadratically additive, and the estimated range of error in the water balance residual would then increase to 17–35% of the net water flux exchanged between the lake and the Solimões River.

Given that the estimates of error for each of the water balance components were selected from accepted literature values for the methods employed, the tendency should be to overestimate some of the errors and underestimate others, thereby causing a compensating effect. The most probable magnitude of the error in the residual term may therefore be somewhere in between the range of error listed above. The volume of water represented by this range of error is 13×10^6 to 27×10^6 m^3, which is larger than the estimates of annual water flux contributed by inflow from Lake Miriti, seepage exchange, or evaporation from the lake surface. It is clear, however, that the magnitude of the residual error in the water balance for Lake Calado is most sensitive to the error in estimating changes in lake volume (δS) and in estimating the

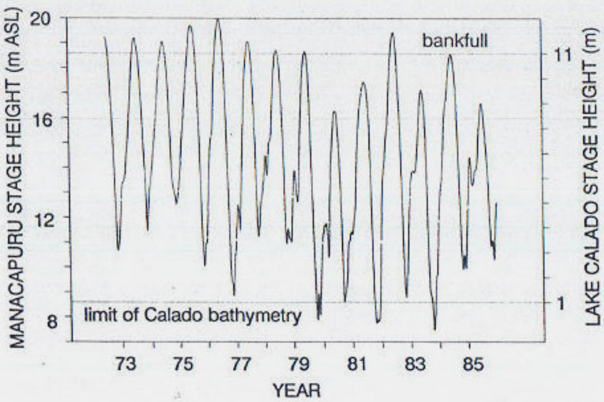

Figure 10. Stage height at the town of Manacapuru (in meters above sea level) and inferred stage height at Lake Calado from 1972 through 1985. Manacapuru is a gauging station on the Solimões River within about 5 km of Lake Calado.

influx of runoff (R). We believe that both of these components have been measured well [see *Lesack*, 1993]. Moreover, this belief is supported by our drogue measurements of current velocities in the channel connecting the lake and the Solimões River (Table 2), which clearly show that the currents reversed direction when the water balance model predicted that they should. The assignment of quite large errors to the inflow from Lake Miriti and to seepage exchange had little effect on the size of the total residual error.

Potential Interannual Variation

From the water balance model it is clear that the components which play the largest role in determining the mixture of water found in the lake and the amount of water that is exported to the river as the lake flushes during falling water are the amount of local runoff and the rate and range of stage change that occurs in the river. A second issue, however, is that during the present study, the lake was virtually empty (conveniently) at the beginning of the water year ($S_I \approx 0$). A more general discussion of the mixture dynamics would have to allow for the possibility that a significant amount of water may be carried over to subsequent water years. Given three significant sources of variation, a key question is whether the amounts of local runoff are coupled to particular patterns of stage change and minimum levels of lake volume or whether they are independent. If they are coupled, the variability that will occur in the lake will be much less than if each operates independently. Analysis of the long-term water level records for the Solimões River and of the hydrologic characteristics of the local catchment around Lake Calado provides some insight on the amount of variability that might be expected for each of the above water balance components.

Carryover component from previous water years. The historical record of daily stage height at the town of Manacapuru, a gauging station on the Solimões River within about 5 km of Lake Calado, covers the period from 1972 through 1985. Analysis of the record provides a means of evaluating the variability in the volume of lake water carried over from one water year to the next (S_I) in Lake Calado and whether patterns of stage change may be coupled to local weather. First, *Lesack* [1988] has established a tight linear relation ($R^2 > 0.999$) between stage heights ranging from 2 to 11 m at Lake Calado and stage heights at Manacapuru during the period of study. Below 2 m the relation breaks down, with a large hysteresis effect between rising and falling water. A plot of the Manacapuru stage record with a scale of inferred stage height at Lake Calado superimposed (Figure 10) shows a number of important features. With the exception of one year, the falling-water period is monotonic, while during rising water there is often a brief decline in level early in the water year. Bank-full stage was exceeded in 9 years of the 14-year record but in only 1 of the last 6 years. The low-water stage during 1983, which begins the water year during the present study, is the lowest stage level of the record, and this level is approached in only two other years. Figure 11 shows the inferred minimum annual volumes for Lake Calado that correspond to the minimum stage levels shown in Figure 10. These low-water lake volumes potentially represent lake water which is carried over to the subsequent water years and range from less than 1×10^6 to 16×10^6 m^3. Relative to the annual water fluxes obtained during the present study, the amount of water carried over to subsequent years could range from the least important term in the budget to the third most important (see Table 3).

Second, if it is assumed that the long-term variation in rainfall at Reserva Ducke is representative of rainfall variation within the Lake Calado area, analysis of the concurrent records (1972 through 1983) of Manacapuru stage and Reserva Ducke rainfall may provide evidence of potential linkage between local weather at Lake Calado and patterns of stage change in the adjacent Solimões River. *Lesack* [1988] has established that there is no evidence of a relation between Reserva Ducke rainfall and minimum stage heights at Manacapuru in corresponding water years. There is, however, a weak relation (Figure 12) between Reserva Ducke rainfall and maximum stage heights at Manacapuru and a moderate relation between rainfall and inferred annual changes in the volume of Lake Calado (derived from annual decline in Manacapuru stage and the hypsometric characteristics of the lake). Although this clearly should be investigated further, a potential interpretation of these patterns is that the regional-scale weather, which drives the change in stage of the Solimões River, is only weakly coupled to local weather.

Interaction between local runoff and flooding from river. Assuming weak linkage between stage change patterns and local weather, two critical factors that control whether water

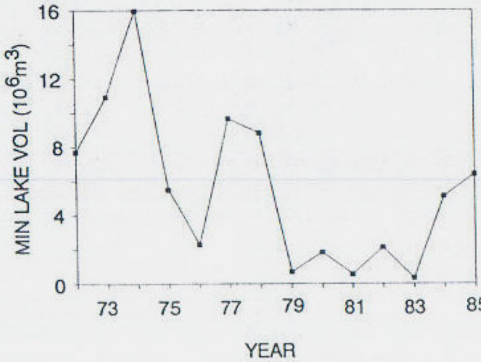

Figure 11. Minimum annual volume of Lake Calado from 1972 through 1985, as inferred from minimum annual stage levels observed at Manacapuru.

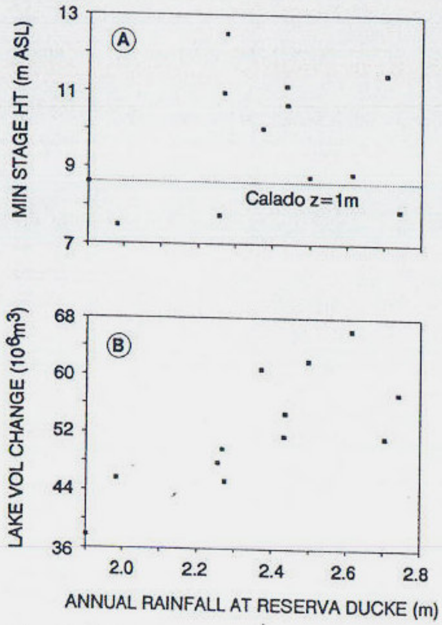

Figure 12. (a) Minimum annual stage height at Manacapuru (in meters above sea level) and (b) inferred annual changes in the volume of Lake Calado versus annual rainfall at Reserva Ducke from 1972 through 1983.

enters the lake from the river during all or only some portion of the rising-water period are the amount of runoff from the lake catchment and the timing of the increased rates of runoff associated with the annual period of higher rainfall. First, given that the rates of evapotranspiration (ET) in the central Amazon are high in magnitude and do not vary by much, interannual variation in runoff at the spatial scale of small catchments could vary by at least a factor of 2 [*Lesack*, 1993]. For example, assuming that actual ET could vary from about 1300 to 1400 mm in rough correspondence with dry and wet years, respectively [see *Shuttleworth*, 1988; *Bruijnzeel*, 1990], if rainfall varies with a 10 year return period from about 2000 to 2800 mm among dry and wet years [*Lesack and Melack*, 1991], the amount of water potentially available for runoff should range at least from 700 to 1400 mm for dry and wet years, respectively. Second, the inferred annual rise in the stage of Lake Calado ranges from 6 to 11 m, which can span an annual change in lake volume ranging from 36×10^6 to 68×10^6 m^3. The above clearly shows that both local runoff amounts and the range of stage change in the adjacent river have considerable interannual variation. However, the nature of the interaction between potential flooding by river water and the potential of local runoff to prevent river water from entering the lake cannot be directly addressed from the information presently available. Nevertheless, the information required to establish a relation between the intra-annual distribution of rainfall and the timing of runoff does exist. Such a relation could be used to simulate the timing of runoff, as a function of rainfall distribution from the Reserva Ducke records, and various combinations of runoff volume, runoff distribution, and stage change could be incorporated into our water balance model to simulate the potential interactions. This type of analysis is beyond the scope of the present paper but will be pursued in future investigations.

From the above analysis it is clear that there is considerable potential for variation in each of local runoff, Solimões flooding, and the carryover component. However, the degree to which the variation may be coupled requires further investigation.

Representativeness of the Results

The measure of success attributed to the present study might ultimately depend on whether obtaining a water balance value with a potential residual error of ±15–30% is considered an acceptable level of accuracy. The main point of the discussion of errors was to give a sense of which components of the water balance contain the most error and whether the errors matter. It is difficult to assess, for example, the error involved in regionalizing unit runoff values from a small catchment of undisturbed rainforest to a larger catchment that has been subjected to a variety of land use practices. Conversely, we find it difficult to envision how obtaining a water budget to within a few percent for one lake out of more than 8000 on the Amazon floodplain [see *Melack*, 1984] would have general utility except for other lakes that are in precisely the same state of disturbance. Given the level of effort required to obtain credible water budgets either for terrestial catchments or for floodplain lakes, we are skeptical that evaluating land use effects can be approached with brute force as a random sampling problem. A broader objective of the present research, however, is to develop a more general model of the water balance dynamics for floodplain lakes associated with major world rivers. Given this goal, the results obtained for Lake Calado are more useful because the water balance model is based on runoff from an undisturbed catchment. We consider incorporating the effect of different land use patterns on landscape water balances to be a refinement, where the necessary additional measurements could be guided by modeling efforts based on the undisturbed situation.

Considering the importance of local runoff in the dynamics of the water mixture in Lake Calado, a major question is the abundance of lakes with local catchments that are sufficiently large to prevent river water from entering lakes during rising-water periods on the Amazon floodplain. The enumeration of Amazon floodplain lakes performed by *Melack* [1984] included classification of the lakes into various morphotypes. Within the classification scheme, ria lakes (sum of composite and dendritic morphotypes) represent a morphology associated with relatively large lakes that have significant local catchment areas, and they constitute about 1000 of the 8000 lakes that were counted. Subsequent analysis by *Sippel et al.* [1992] has demonstrated that although only 10% of the lakes on the Amazon floodplain are larger than 2 km^2 (mostly ria lakes), they constitute more than half of the total lake area on the floodplain. We hence conclude that Lake Calado represents an important class of lake among the greater population on the Amazon floodplain and that our results have implications for a meaningful portion of them.

Implications for Hydrology of Floodplain Lakes in the Wet Tropics

Setting the water balance and hydrologic characteristics for Lake Calado into a broader context is somewhat difficult. Although there is little doubt that comparable floodplain lakes are abundant and widespread in the wet tropics, there is very little information available on lakes which are sufficiently analogous in a hydrologic sense for comparisons to be meaningful. The analysis of *Forsberg et al.* [1988] indicated that floodplain lakes in the central Amazon would require a catchment area (BA) to water surface (LA) ratio of greater than 20 in order to contain primarily local water during the annual low-water period, while lakes with a ratio of less than 20 would be expected to contain a mixture of river water and local water. The BA:LA ratio of Lake Calado is 7, and during the 1983–1984 water year, 3 times the water volume received from the river had been exported back to the river by the time that low water had been reached. This amount of water is equivalent to 2 times the bank-full volume of Lake Calado over the period from mid-rising water to low water. However, if the flushing rate is calculated in an instantaneous sense, which would be equivalent to the instantaneous volume of the lake (continually changing) divided by the instantaneous rate of outflow to the river, the lake could be considered to have flushed many times by the time that low water was reached. Over the annual cycle we have distinguished periods of flushing without flooding (low water), progressive flooding without flushing (early rising water), progressive flooding with flushing (later rising water), and regressive flooding with flushing (falling water). Given the above findings, the notion of being able to predict which lakes will contain primarily local water during "the annual low-water period" based on BA:LA represents an oversimplification of what may actually occur hydrologically. Moreover, the critical value of BA:LA will depend on whether it is a dry year or a wet year, whether the change in river stage is large and occurs rapidly or is smaller and slower, and whether water has been carried over from the previous year. Although the hypothesis of Forsberg et al. represents an important initial step, it would now benefit from a reevaluation which incorporates improved hydrologic data and potential sources of interannual variability.

The fact that water is capable of flowing out from lakes with relatively small catchment areas and into the river during rising-water periods and the fact that such lakes are abundant [*Melack*, 1984; *Sippel et al.*, 1992] may critically limit our ability to directly characterize the regional flooding hydrology of floodplain lakes in the wet tropics. For a prime example, *Marsh and Hey* [1989, 1991] were able to perform a regional analysis of the frequency and duration of flooding in lakes in the Mackenzie Delta (Canadian arctic) by using airborne observational techniques to determine the timing of lake flooding during the spring flood. Concurrently measured water levels in the adjacent river channels allowed them to obtain the threshold elevation that needed to be reached (sill elevation) before flooding of a given lake would commence. Subsequent analysis of long-term water level records then allowed reconstruction of the historical frequency and duration of flooding that would be expected for lakes of a given elevation among their populations of lakes. This entire analysis, however, was only possible because Marsh and Hey could assume that when water levels in the river channels exceeded the lake sill elevation, during the rising-water period, flooding of the lake would indeed commence. This assumption has been validated for Mackenzie Delta lakes by careful water balance studies [*Marsh*, 1986; *Marsh and Bigras*, 1988]. Our work clearly shows that this assumption is not valid for the Amazon floodplain and probably is not valid for floodplain lakes in other areas of the wet tropics.

Conclusions

This study represents the first evaluation over an annual cycle of the mixture dynamics of lake water derived from multiple sources for a lake on the central Amazon floodplain and the first such evaluation available for lakes associated with the floodplains of large tropical rivers in general. Lake Calado represents an important class of lake which accounts for a substantial amount of the total lake area among the greater population on the Amazon floodplain. The dominant components of the annual water budget are local runoff, inflow from the adjacent river, and water carried over from previous years. All three components are capable of considerable independent variation. This provides an important base of information needed for the development of a more general model of the water balance dynamics in floodplain lakes.

Despite a relatively low ratio, 7, of catchment area (BA) to water surface area (LA), Lake Calado exported back to the river about 3 times the amount of water originally received from the river. This resulted in substantial flushing of the lake before the arrival of the low-water period and is inconsistent with previous work which concluded that floodplain lakes in the central Amazon would require a BA:LA ratio of greater than 20 in order to contain primarily local water during the annual low-water period.

In contrast to a general paradigm in which flooding from river channels is perceived to play a dominant role in the annual flushing, nutrient replenishment, and reinitialization of lacustrine ecosystems on the floodplains of major world rivers, our work clearly shows that local rainfall and runoff can be sufficiently large components of lacustrine water balances in the wet tropics to prevent flood waters from entering lakes with open connection to the river, even during periods of rising river levels. This may critically limit our ability to directly characterize the long-term frequency and duration of flooding in floodplain lakes of the wet tropics.

Acknowledgments. We thank E. Salati, F. Bergamin, and H. O. R. Schubart, directors of the Instituto Nacional de Pesquisas da Amazônia (INPA), and O. Fonseca, G. Vazzoler, A. dos Santos, B. Robertson, and J. Salomão Ribeiro for support and encouragement in Brazil. We appreciate the field assistance provided by Antonio Adalberto da Silva and Leslie Smith. This analysis benefitted from the suggestions of S. Cooper and J. Dozier and discussions with P. Marsh and D. Moore. The manuscript was improved by the suggestions of R. Stallard and two anonymous reviewers. Financial support was received from NSF grants DEB81-11398 and BSR-8706643 to T. R. Fisher and J. M. Melack.

References

Brinkmann, W. L. F., and A. Santos, Natural waters in Amazonia, VI, Insoluble calcium properties, *Acta Amazonica*, *3*, 33–40, 1973.

Bruijnzeel, L. A., Hydrology of moist tropical forests and effects of conversion: A state of knowledge review, report, Humid Trop. Programme, Int. Hydrol. Programme, UNESCO, Paris, 1990.

Brutsaert, W. H., *Evaporation into the Atmosphere*, 299 pp., D. Reidel, Norwell, Mass., 1982.

Carignan, R., and J. J. Neiff, Nutrient dynamics in the floodplain ponds of the Parana River (Argentina) dominated by the water hyacinth *Eichornia crassipes*, *Biogeochemistry*, *17*, 85–121, 1992.

Engle, D. L., and J. M. Melack, Consequences of riverine flooding for seston and the periphyton of floating meadows in an Amazon floodplain lake, *Limnol. Oceanogr.*, *38*, 1500–1520, 1993.

Fisher, T. F., and P. E. Parsley, Amazon lakes: Water storage and nutrient stripping by algae, *Limnol. Oceanogr.*, *24*, 547–553, 1979.

Forsberg, B. R., A. H. Devol, J. E. Richey, L. A. Martinelli, and H. Dos