ELSEVIER

Journal of Hydrology 183 (1996) 191–204

Journal of Hydrology

Using stable isotopes to determine sources of evaporated water to the atmosphere in the Amazon basin

Luiz Antonio Martinelli[a,*], Reynaldo Luiz Victoria[a,c], Leonel Silveira Lobo Sternberg[b], Aristides Ribeiro[a], Marcelo Zacharias Moreira[a,b]

[a]*Centro de Energia Nuclear na Agricultura, Av. Centenário, 303, Cep 13416-000, Piracicaba, SP, Brazil*
[b]*Department of Biology, University of Miami, Coral Gable, FL 33134, USA*
[c]*Departamento de Física e Meteorologia, ESALQ-USP, Caixa Postal 9, 13400, Piracicaba, SP, Brazil*

Received 21 April 1995; accepted 29 October 1995

Abstract

The return of water in vapor form from the land to the atmosphere, via plant transpiration and evaporation, is fundamental for the maintenance of the regional water cycle in the Amazon basin. Whereas transpiration, the dominant process, has the extensive vegetation cover as a large single source, evaporation can have several sources, and their relative importance and location are poorly known. The isotopic composition ($\delta^{18}O$ and δD) of water from various sources was used to see whether or not specific sources of water vapor to the atmosphere could be determined. It is well established that natural waters fall on a line called the meteoric water line (MWL; the regression of $\delta^{18}O \times \delta D$), with slope equal to eight and an intercept equal to ten. When a water body loses water via evaporation the slope become smaller than eight, typically 5–6. We estimated the slope of the regression of $\delta^{18}O \times \delta D$ for several potential sources. We analyzed 1273 samples: 500 of rainfall, 409 of river water, 134 of lake water, 164 of soil water, 40 of throughfall and stemflow water, and 26 of shallow ground-water. We found that large rivers and lakes are likely contributors of evaporated water to the atmosphere. However, as they cover only a small area of the basin, other sources are needed. Probably, evaporated water originates from several small sources that were not detected by the isotopic composition of our data.

1. Introduction

The recycling of water to the atmosphere via evapotranspiration has been shown to be very important for the maintenance of the water cycle in the Amazon region. The estimates

* Corresponding author.

SSDI 0022-1694(95)02974-5

for the contribution of evapotranspiration to local rainfall range from 48% to 80% (Villa Nova et al., 1976; Molion, 1979; Salati et al., 1979; Marques et al., 1980; Jordan and Heuneldop, 1981; Ribeiro et al., 1982; Leopoldo et al., 1982a,b; Shuttleworth et al., 1984). Transpiration is the largest component of evapotranspirative flux, and has a clearly recognizable source: the dense vegetation of the region. Evaporation, on the other hand, may be responsible for up to 40% of the evapotranspiration flux (Gat and Matsui, 1991; Victoria et al., 1992), but very little is known about the possible loci of the process.

In this study, we first analyzed a data set of precipitation samples, collected at several locations throughout the Amazon basin, to confirm the importance of water evaporation to the regional hydrologic cycle. Second, we attempted to identify the possible sources of evaporated water, through analyzes of the stable isotopic composition of the water in rivers and lakes, soil water, water intercepted by the canopy, and other sources inside the forest.

The movement of water is the main driving force for various other biogeochemical cycles in the region (Richey et al., 1990). Its influence can extend from regional to global scales, as the condensational energy released by convective precipitation can be of sufficient magnitude to affect global climate patterns (Paegle, 1987). In addition, as recently demonstrated by site-specific studies, substituting forest for pasture alters the energy balance and consequently the water balance in the Amazon region (Wright et al., 1992). Results from global circulation models (CGMs) show that replacement of the rain forest by pasture will decrease the rainfall and evaporation (Shukla et al., 1990). As the extent of deforestation in the last decade is approximately 20 000 km^2 (Fearnside et al., 1990; Skole and Tucker, 1993), to properly address the consequences of the land-cover and land-use changes in the region it is necessary to understand the undisturbed water cycle.

2. Materials and methods

2.1. The isotopic method: units of measurement and background

The variations in natural abundance of oxygen and hydrogen isotopes in water samples are measured as deviations from an international standard, which is a composite sample of water from several oceans called SMOW (Standard Mean Ocean Water). The results are given in δ per mil units, defined as

$$\delta‰ = \frac{R_{\text{sample}} - R_{\text{SMOW}}}{R_{\text{SMOW}}} \times 10^3$$

where R is the ratio $^{18}O/^{16}O$ or D/H, and the values are represented by $\delta^{18}O‰$ and $\delta D‰$, respectively.

The isotopic composition (^{18}O and deuterium) of samples of natural waters collected all over the world falls on a line which yields the following equation: $\delta D = 8\delta^{18}O+10$ (Craig, 1961). This line is called the meteoric water line (MWL). Fig. 1 shows the isotopic composition of the water vapor phase (V_0) in the atmosphere, as it generates a rainfall with the isotopic composition (R_0). The rain becomes enriched in ^{18}O and D, owing to the

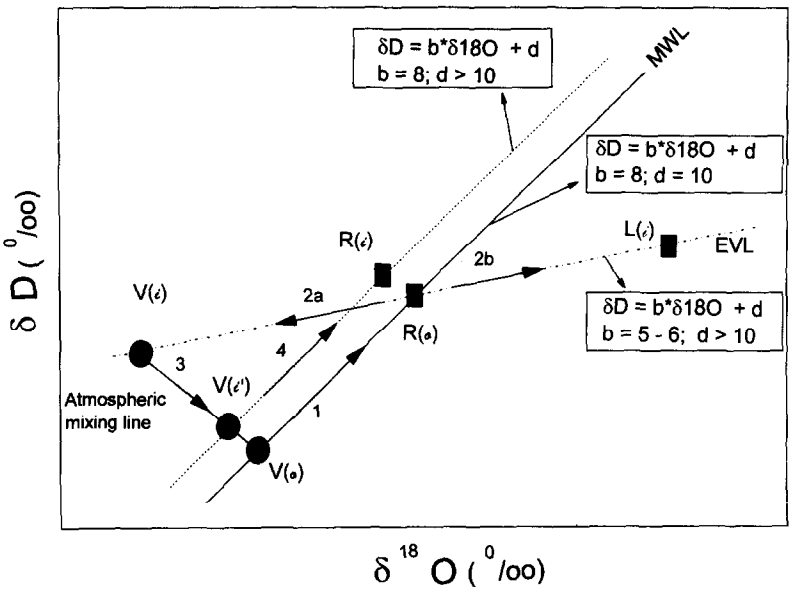

Fig. 1. Theoretical plot of $\delta^{18}O$ vs. δD showing the evolution of the water isotopic composition of a hypothetical rainfall event (R_o) that was formed by atmospheric water vapor with isotopic composition V_o (Arrow 1), both following the meteoric water line (MWL). After undergoing evaporation, the remaining liquid phase (lake, soil water, water trapped on leaves, etc.) became enriched (Arrow 2b), and the vapor phase (V_i) became depleted (Arrow 2a), following a typical evaporation line (EVL). After reaching a new equilibrium with the vapor already present in the atmosphere (Arrow 3), the vapor will have a new isotopic composition (V_i'), which in turn will generate a new rainfall event (R_i) that follows a line parallel to the MWL, but with a d value higher than ten (Arrow 4). (Adapted from Gat and Matsui (1991).)

isotopic fractionation that occurs during the condensation process (represented by Arrow 1 in Fig. 1).

Isotopic enrichment occurs in a water body during the process of evaporation because of the differences in the vapor pressure between DHO and $H_2^{18}O$ and $H_2^{16}O$ (represented by Arrows 2a and 2b in Fig. 1). Further, because of equilibrium and kinetic isotopic fractionation factors, the relationship between δD and $\delta^{18}O$ values of the remaining water pool and of the vapor phase will be described by a line with slope less than eight and ranging between five and six (Craig, 1961; Gat, 1971). The δD and $\delta^{18}O$ values of the remaining water pool (L_i) will lie to the right of the MWL, and the δD and $\delta^{18}O$ values of the vapor phase (V_i) will lie to the left of the MWL, defining a line called the evaporation line (Fig. 1). The water vapor (V_i) generated by this process, at equilibrium, will have a new isotopic composition (V_i'). Arrow 3 in Fig. 1 represents the atmospheric mixing line, where V_i' will lie. The relationship between δD and $\delta^{18}D$ of precipitation (R_i), which originates from V_i', is described by a line parallel to the MWL (i.e. with slope equal to eight), but with intercept higher than ten (called the deuterium excess). Thus, the deviations from MWL produced during evaporation and condensation of vapor provide useful isotopic markers to trace the source of vapor in the Amazon basin. First, we can measure the slope described by the δD and $\delta^{18}O$ values of water bodies which are potential sources of vapor. If the

slopes are significantly lower than that of the MWL, then we can conclude that a particular water body has undergone considerable evaporation. If a given water body is large, this is one way to determine that it is an important source of vapor in the Amazon basin. Second, we can calculate the deuterium excess of precipitation and the vapor phase, by forcing a line with a slope of eight through the δD and $\delta^{18}O$ values of these waters, and determine whether they were produced by evaporation (Daansgard, 1964). If the deuterium excess is significantly greater than ten, we can conclude that the measured precipitation or vapor was produced by evaporation (Gat and Matsui, 1991).

2.2. The database

The stable isotopic composition of three types of samples were determined: (1) rainfall samples before any interaction with vegetation; (2) samples after some kind of interaction with the forest vegetation, i.e. throughfall, soil water, shallow ground-water, and vapor generated by plant and soil evaporation and plant transpiration; (3) samples of rivers, lakes, and other water bodies.

The precipitation samples were collected at 2 week intervals, from January 1989 to January 1991, at 11 locations spread throughout the Amazon basin. These locations are part of the IAEA network for studying stable isotopes in precipitation.

To test the effects of evaporation from rivers, we used water samples collected during 1971 and 1972 from several rivers throughout the basin (Matsui et al., 1980), and from river water samples collected from 1982 to 1985 during the CAMREX project (Richey et al., 1990).

In addition, two small creeks (igarapés) in central Amazon, and one in eastern Amazon, were also sampled. The first two (Tarumã and Ladeira Grande) are both located approximately 150 km north of Manaus, at the ZF-2 Forest Reserve, and were sampled by Leopoldo (1981) and Senna (1992), respectively. The eastern Amazon site was at Fazenda Vitória, an experimental station located 300 km south of Belém, and the samples were taken at 2 week intervals from April 1992 to April 1993.

Water samples from lakes were collected during the CAMREX cruises. In total, 82 samples were taken from 41 lakes. Lake Calado was more intensively sampled than other lakes. The main sources of water for this lake are the rio Solimões and igarapés draining the highlands (locally called 'terra-firme') (Lesack and Melack, 1995). Three sampling points were established: P1, near the linking channel to the rio Solimões; P2, at the center of the lake; P3, near an igarapé mouth. The samples were collected weekly during a rising water period when the major flow of water is from the Solimões to the lake. The sampling period was 19 May–23 August 1988. Lake Calado and five additional lakes near Manaus were sampled monthly during 1990. Four of these lakes were linked to the rio Solimões–Amazon (Brito, Calado, Castanho and Marchantaria) and the other two were linked to the rio Negro (Piranha and Cacau).

Samples of precipitation, throughfall, stemflow, and soil water were collected at the ZF-2 Reserve (central Amazon) by Leopoldo (1981), and samples of precipitation, shallow ground-water and deep soil water were sampled at Fazenda Vitória (eastern Amazon). The samples from Fazenda Vitória were taken at two sites: the dense forest and a 20-year-old pasture. Samples of rainfall were collected from one pluviometer above the canopy and

one in the pasture. Soil water was mainly collected during the rainy season using soil solution extractors (vacuum tensiometers).

Finally, other water sources, which include water evaporation from litter, trunk and leaves, and internal water evaporation and transpiration by plants, were analyzed by Ribeiro et al. (1996). The water vapor samples were collected at four heights from the ground: 0.5, 10, 28 and 52 m in a typical terra-firme forest located at the Biological Reserve of Jaru (Rondônia State; 10°05′S, 61°56′W).

2.3. Analysis

The analysis were performed at the Centro de Energia Nuclear na Agricultura using a Micromass 602E (Micromass, Winsford, UK) for ^{18}O and a Matt (Matt, Bremen, Germany) GD150 for deuterium. Most of samples were analyzed in replicates. A maximum deviation of 0.2‰ and 2.0‰ was permitted for oxygen and deuterium, respectively.

2.4. Statistics

The Student's *t*-test, with *P* values of 1 and 5%, was used to test the significance of *d* (deuterium excess) values larger or smaller than ten, and slope smaller than eight. In the case of Fazenda Vitória, the slope for shallow ground-water and soil water was compared with that for local precipitation, instead of the general value of eight.

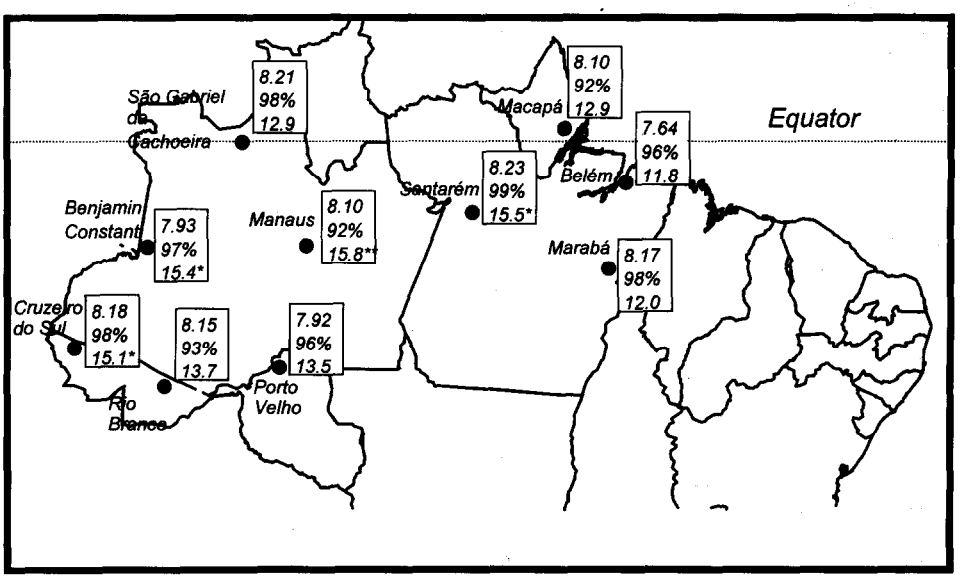

Fig. 2. Map of Amazon basin showing the location of the precipitation sampling network. For each station, from the top to the bottom, the slope and correlation coefficient of the regression line for $\delta^{18}O$ vs. δD and the average *d* values are shown.

3. Results and discussion

To determine the possible contributions of evaporated water to the rainfall, we used d values. The distribution of d values shown in Fig. 2 confirms earlier findings about the importance of a significant evaporative flux in the Amazon hydrological cycle. The geographical distribution, with higher d values in the central area of the basin, suggests that direct evaporation processes are more significant in this area. This is also the main axis of the Solimões–Amazon river, where approximately 150 000 km^2 of floodplain exist. About 11% of this area is covered with lakes (Sippel et al., 1992). The volume of some of these lakes can be significantly reduced during the dry season, indicating that the evaporative flux is very important at this time. In addition, the rivers can be a significant source of evaporated water. Using techniques similar to those used in this study, Ramesh and Sarin (1992) found that the Ganges river suffered significant losses by evaporation, especially during the dry season. It is thus reasonable to hypothesize that the central area of the Amazon could be an important source of evaporated water to the atmosphere.

For rivers and lakes we first examined the slope of the line $\delta^{18}O \times \delta D$. For the 1971 and 1972 data sets, the regression analysis was done in two different ways: (1) grouping the

Table 1
Values of slope with respective standard errors, coefficient of regression, and number of data points for the regression of $\delta^{18}O \times \delta D$ in river water samples for the 1971 and 1972 data

	Slope	r^2 (%)	n
1971 data			
Longitude (deg)			
48–55	7.25 ± 0.62	90	18
55–70	7.03 ± 0.48	89	27
70–80	8.51 ± 0.13	99	29
Month			
Sept.	7.21 ± 0.79	83	19
Oct.	8.06 ± 0.16	98	38
Nov.	8.04 ± 0.24	99	16
Dec.	8.86 ± 0.44	98	9
1972 data			
Longitude (deg)			
48–55	7.74 ± 0.41	90	43
55–60	7.82+038	88	58
60–65	7.36 ± 0.27	92	66
65–80	8.45 ± 0.16	99	31
Month			
Jan.	6.83 ± 0.38*	90	38
Feb.	6.61 ± 0.38*	92	24
Mar.	7.11+0.53	89	24
Apr.–May–June	8.10 ± 0.13	99	27
July	7.58 ± 0.28	98	20
Aug.	8.29 ± 0.32	97	19
Sept.–Oct.–Nov.	8.29 ± 0.32	97	23
Dec.–Jan.	6.44 ± 0.55*	95	9

* The slope is statistically smaller than eight at the 1% level.

Table 2
Values of slope with respective standard errors, coefficient of regression, and number of data points for the regression of $\delta^{18}O \times \delta D$ for samples collected exclusively along the rio Solimões–Amazon, for the 1971 and 1972 data and for CAMREX cruises

	Slope	r^2 (%)	n
1971			
Sep.–Oct.–Nov.	5.64 ± 0.66**	83	14
1972			
Jan.–Feb.	6.21 ± 0.36**	88	41
Mar.	5.95 ± 0.92**	78	14
July–Aug.	6.47 ± 0.50	84	34
Oct.–Nov.	7.70 ± 0.90	81	19
CAMREX cruises			
Cruise 4	6.72 ± 0.52**	89	19
Cruise 6	5.56 ± 0.8 7**	76	15
Cruise 12	6.79 ± 0.67*	83	23
Igarapé			
Tarumã	6.38 ± 0.38**	25	26
Ladeira Grande	6.69 ± 0.60**	82	30
Faz. Vitória	6.16 ± 1.63	37	26

* The slope is statistically smaller than eight at the 5% level.
** The slope is statistically smaller than eight at the 1% level.

river data by month, and by latitude and longitude; (2) examining only the samples from the Solimões–Amazon, grouping them by month where data were available. When different rivers are analyzed together, as for the 1971 and 1972 data, only the samples collected in December, January and February yielded slopes that indicated the existence of evaporation losses (Table 1). However, when only the samples collected along the main stream were analyzed, slopes significantly smaller than eight were found most of the time. This suggests that as water travels down-river some is lost by evaporation (Table 2). Samples from the main channel and mouth of tributaries collected during the CAMREX cruises also showed the same pattern, with slopes smaller than eight. Therefore, at least the main channel of the basin appears to be a significant source of water to the atmosphere (Table 2).

The analysis shows that the isotopic behavior in várzea lakes is much more complex than that in rivers. The slopes for lakes were statistically smaller than eight only in CAMREX cruises 9 and 12 and the month of December in the annual sampling around Manaus (Table 3). In general, várzea lakes are fed by two water sources: a major river, such as the Solimões–Amazon in the case of Lake Calado, or one or more igarapés that drain the local basin or terra-firme forest (Lesack and Melack, 1995). These two sources have different isotopic composition, owing to their different origins (Martinelli et al., 1989). Therefore, during some periods of the year, a lake may experience spatial variations in its stable isotopic composition as a function of the different water sources. A good example can be seen in Fig. 3, which shows the variation of $\delta^{18}O$ values in water samples collected at three points in Lake Calado as a function of time and height of the Solimões river. Lesack and Melack (1995) showed that, in late May, the local runoff and Solimões river were the main water sources to Lake Calado, contributing similar volumes of water to

Table 3
Values of slope with respective standard errors, coefficient of regression, and number of data points for the regression of $\delta^{18}O \times \delta D$ for lakes sampled during CAMREX cruises, for the monthly sampling around the city of Manaus (central Amazon), and for Lake Calado

	Slope	r^2 (%)	n
CAMREX cruises			
Cruise 7 (Feb.)	5.92 ± 1.60^{NS}	74	7
Cruise 8 (July)	7.95 ± 1.25	80	12
Cruise 9 (Aug.)	$6.46 \pm 0.66^{**}$	81	25
Cruise 10 (Apr.)	6.93 ± 1.01	73	19
Cruise 12 (May)	$6.61 \pm 0.61^{*}$	85	20
Monthly sampling			
Jan.	7.95 ± 0.41	99	5
Feb.	6.44 ± 1.54	85	5
Apr.	7.11 ± 0.77	95	6
May	8.04 ± 0.31	99	6
June	6.44 ± 1.25	86	6
July	7.77 ± 1.39	91	5
Aug.	$10.11 \pm .53$	92	6
Sept.	7.14 ± 0.91	94	6
Dec.	$3.22 \pm 1.34^{**}$	59	6
Lake Calado			
Point 1 (P1)	$5.39 \pm 1.09^{**}$	65	15
Point 2 (P2)	5.55 ± 1.80^{NS}	44	12
Point 3 (P3)	$2.17 \pm 1.72^{**}$	11	24
Igarapé	0.80 ± 3.46^{NS}	4	15
Rainfall	7.84 ± 0.36	97	16

* The slope is statistically smaller than eight at the 5% level.
** The slope is statistically smaller than eight at the 1% level.
NS, Although the slope is smaller than six, it is not statistically different from eight.

the lake. From this point on, the contribution from local runoff decreased and that from rio Solimões increased. Furthermore, the isotopic composition of the water along the lake clearly shows a gradient, indicating the different degree of mixture between the two water sources. At P1, near the channel that links the river to the lake, the water was isotopically lighter, showing the influence of the Solimões river. At the other extreme, P3 shows the isotopically heaviest values, as influenced by the heavier feeding igarapé. At the center of the lake (P2) the values were intermediate between P1 and P3. As the lake drainage proceeds, the contribution of the local runoff increases (Lesack and Melack, 1995). The predominance of one single source is reflected in the isotopic composition of the water at all sampling points, which becomes heavier and more uniform (Fig. 3). As a consequence of this water mixing, the slopes, and even the correlation coefficient yield for the $\delta^{18}O \times \delta D$ line, were distinct for each water type (Table 3).

Besides the spatial difference discussed above, there is also a temporal variability as a function of the river hydrography (Fig. 3). We discuss this variability using P1, which 'feels' the influence of the feeding river, as an example. In general, the lake water was isotopically heavier than the feeding river throughout the sampling period (average value for the period was $-7.2 \pm 0.3‰$), indicating the larger contribution of the local basin water

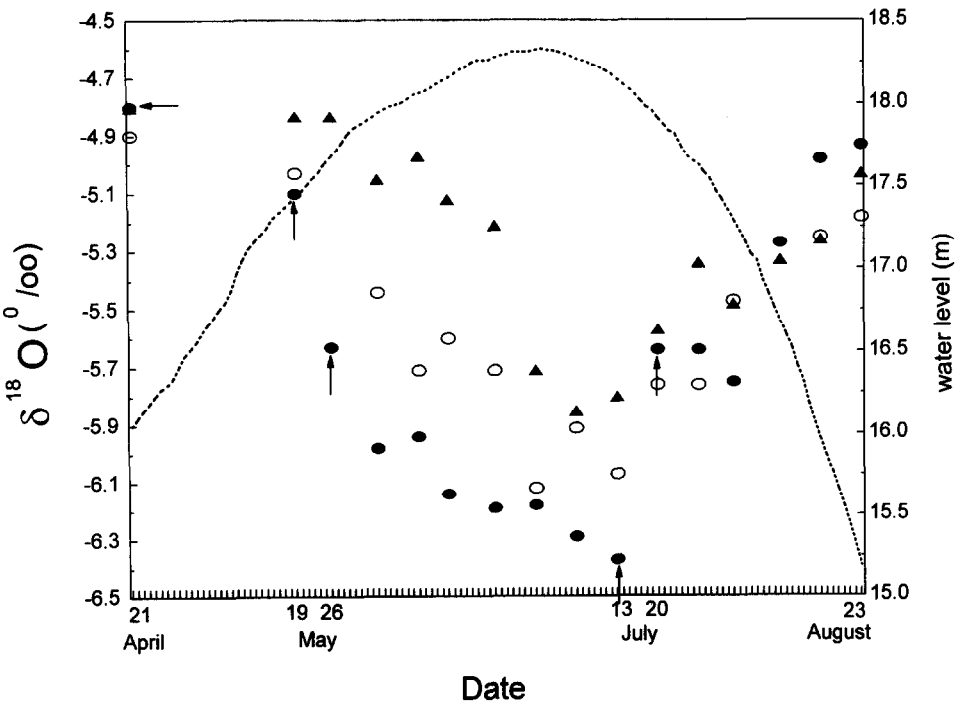

Fig. 3. Temporal variability of $\delta^{18}O$ values at three points in Lake Calado: near the linking channel to the rio Solimões (●, P1); at the center of the lake (○, P2); near the mouth of the feeding igarapé (▲, P3). The dashed line represents the rio Solimões water level at Manacapuru, and arrows indicate points of change in $\delta^{18}O$ values referred to in the text.

(average value for the period was −5.0 ± 0.20‰). However, after the river flooding, the isotopic composition of the lake tended towards that of the feeding river, clearly showing the mixture of the two water types. With the falling of the river hydrography, the isotopic composition of the lake reversed towards that of the local basin water. The first major change in the isotopic composition of Lake Calado occurred during the river flooding, between 19 and 26 May, when the river water level rose from 17.4 to 17.7 m. The isotopic value decreased from −5.1 to −5.7‰. The $\delta^{18}O$ values continued decreasing until 13 July, when the lake reached a minimum of −6.4‰ at a river stage of 18.15 m. At the next sampling (20 July), the river stage decreased to 17.80 m, with an increase in the lake isotopic composition to −5.7‰. The isotopic composition increased until the end of the sampling period, again reaching values similar to the local basin.

Várzea lakes are not homogeneous water bodies prone to evaporation, as is needed for the ideal situation for classical evaporation studies using isotopes. On the contrary, they represent mixtures of water from different sources, where the relative contributions vary as a function of space and time (Lesack and Melack, 1995). Therefore, if lakes have water with isotope ratios falling on a line with a slope significantly less than eight this does not necessarily mean that they are losing water by evaporation.

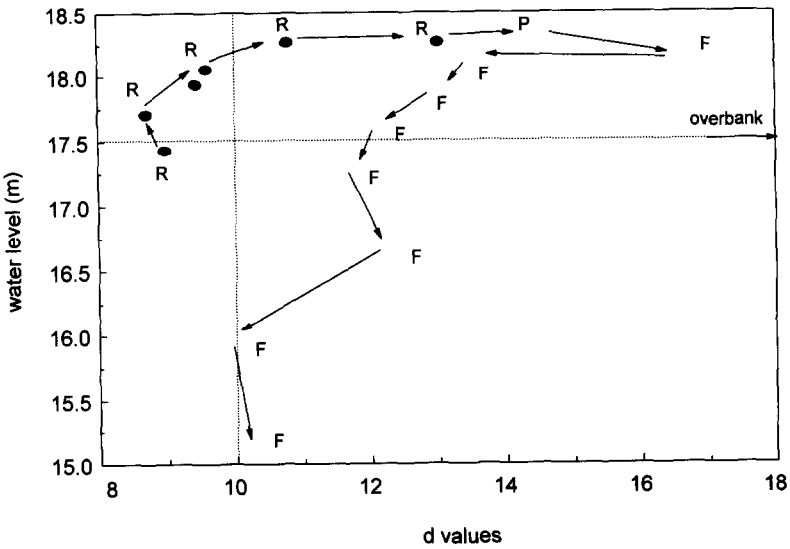

Fig. 4. Variation of d values of Lake Calado water in relation to the rio Solimões water level at Manacapuru, around 20 km upstream from the entrance of Lake Calado, which was used to infer the lake water level. R denotes the rising water period, P the peak of flooding and F the falling water period.

As an alternative to the use of the slopes of the $\delta^{18}O \times \delta D$ correlation, which was not suitable in this case, the d value can be used to indicate water evaporation. In general, the d values for rainfall in the Amazon area are higher than ten, reaching 15 in the central part of the basin (Fig. 2). Values smaller than ten should be produced in the residual water (Fig. 1). One advantage of using d values is that lakes are analyzed individually. Any individual value falling below the MWL is an indication that the associated lake lost water by evaporation. It should be borne in mind that the regression line of all values may fall below the MWL but remain parallel to it, still yielding slopes equal or similar to eight, although the lakes underwent evaporation (Fig. 1). On the other hand, the errors involved in the estimation of the d values are generally large, as the error involved in the analyses of $\delta^{18}O$ is multiplied by eight (J. Gat, personal communication, 1994). The plot of river water level, as a proxy of Lake Calado water level, against d values for P1 is shown in Fig. 4. During the rising water period, at the early flooding stage, the d values were consistently smaller than ten, suggesting that the water in the lake was subjected to evaporation. As the water level rose, owing to river flooding (Fig. 4), the d values increased towards those of the feeding river ($d > 10$), reaching the highest value ($d = 17$) at the beginning of lake drainage. From this point on, the d values progressively decreased towards the value of ten. This trend suggests that várzea lakes can be an important source of evaporated water to the atmosphere, from the falling to the early flooding stage. Outside this period, the d values were always higher than ten, indicating the absence of a significant source of evaporated water. For the CAMREX data set, from 83 lake samples, only 22 showed $d < 10$; of those, 16 were from lakes sampled during Cruise 9, a falling water cruise.

It is clear, from the discussion above, that várzea lakes are very complex water bodies, and that more site-specific studies are necessary, if the isotope technique is to be used to

Table 4
Values of slope coefficient of regression, and number of data points for the regression of $\delta^{18}O \times \delta D$ for precipitation, throughfall, stemflow, shallow ground-water and soil water collected in the central and eastern Amazon

	Slope	r^2 (%)	n
Central Amazon			
Precipitation	8.13	97	20
Throughfall	8.68	98	20
Stemflow	8.30	98	28
Soil water	8.03	97	50
Eastern Amazon			
Precipitation—forest	7.37 ± 0.50	90	26
Precipitation—pasture	7.86 ± 0.18	99	23
Shallow ground-water	7.51 ± 0.60	87	26
Soil water—MC (0.5 m)	5.94 ± 0.87**	78	15
Soil water—MD (0.5 m)	7.15 ± 0.67	88	17
Soil water—PA (0.5 m)	6.81+0.69*	67	17
Soil water—PE (0.5 m)	7.88 ± 0.80	91	12
Soil water—PA (3–6 m)	7.67+0.54	88	28
Soil water—PE (3–6 m)	7.65 ± 0.63	88	22

* The slope is statistically smaller than eight at the 5% level.
** The slope is statistically smaller than eight at the 1% level.

clarify the effects of evaporation from them. Such studies could certainly lead to a better estimate of the real contribution of evaporated water from a specific lake to the atmosphere, but extrapolation of the results to the entire Amazon várzea would be very difficult and unrealistic.

Besides water from river and lakes, water in the forest itself may be an important source of evaporated water to the atmosphere. The most important pools in the forest are the canopy interception and soil water.

The reason to examine throughfall and stemflow resides in the fact that a significant proportion of the rainfall can be trapped in the forest leaves, to be later returned to the atmosphere by evaporation (Franken et al., 1982a,b; Lloyd and Marques, 1988). If this intercepted rainfall is not completely evaporated, the fraction left behind falls on a typical evaporation line. Throughfall and stemflow can be the vehicle for such water. The slope obtained with the data from the Forest Reserve ZF-2 does not comply with the above, but shows values higher than eight, similar to the value found for the local precipitation (Table 4). Another possible source of evaporated water could be forest soil water. Owing to the large amount of litter, the infiltration of water into forest soils is slow (D. Nepstad, personal communication, 1993), enhancing the possibility of water loss by evaporation. This hypothesis was also not confirmed by our data. From soil water sampled at six points, only one sample showed a slope significantly smaller than the local precipitation (Table 4).

We further investigated the importance of the forest as a source of vapor to the atmosphere, by determining the relationship between $\delta^{18}O$ and δD of the vapor phase, as well as comparing the values of this phase with that of the soil water (Ribeiro et al., 1996).

The rationale here is that transpiration by plants does not fractionate after steady-state

equilibrium is reached (1–2 h). Therefore the water vapor generated by this process should be isotopically similar to the soil water taken up by plants. On the other hand, evaporation from a free water surface, when the process does not go to completion, causes fractionation, and the remaining liquid will be enriched in heavy isotopes. For mass balance purposes, the generated vapor will be depleted. Therefore, the δ values of vapor generated by evaporation will be smaller (isotopically lighter) than those for vapor generated by transpiration. The analysis of the daily variation of the isotopic composition of the water vapor inside the forest should therefore be indicative of the relative importance of each of these processes.

A clear diurnal cycle in the δD values of the vapor inside the forest was observed (Ribeiro et al., 1996). The isotopic values were smaller (lighter) during the night than during the day for all forest strata, with the exception of that nearest the ground. Condensation of the vapor caused by a drop in the temperature during the night is the probable cause for this cycle. The increase observed during the day was attributed to the mixture of the remaining vapor from the night with vapor generated by transpiration, plus the re-evaporation of the previously condensed vapor during the day.

4. Conclusions

1. The high d values found in the central area of the Amazon basin indirectly confirm the importance of evaporation in this area as a source of water vapor to the atmosphere.
2. Major rivers such as the Amazon were found to be one of the important sources of water vapor to the atmosphere. Our data suggest that, as water travels down-river, part may be lost by evaporation.
3. Várzea lakes also appear to be an important source, especially during the late falling water and early rising water periods.
4. The rain forest has an important role in the hydrological cycle of the region, returning water to the atmosphere through evapotranspiration, although we could not find a strong signal that could indicate the contribution of various evaporation sources. However, we want to emphasize that only a few locations were sampled in this case. Further research, with more sampling locations and a larger number of samples, should be carried out to obtain a definite answer to this question.

Acknowledgements

We are indebted to the skill of the crew of the 'L.M. Amanai'. J. Gat, K. Rozanski and J. Jouzel made helpful comments on an earlier version of this manuscript. John Melack and his team, especially Diane Eagle, collected water samples from Lake Calado. Financial support for this study was provided by the National Science Foundation, Grant DEB 81-07552. This paper is Contribution 78 of the CAMREX project.

References

Craig, H., 1961. Isotopic variations in meteoric waters. Science, 113: 1702–1703.

Daansgard, W. 1964. Stable isotopes in precipitation. Tellus, 16: 436–468.

Fearnside, P.M., Tardin, T.A. and Meira Filho, L.G., 1990. Deforestation rate in Brazilian Amazonia. PR/SCT, Instituto de Pesquisas Espaciais, São José dos Campos.

Franken, W., Leopoldo, P.R., Matsui, E. and Ribeiro, M.N.G., 1982a. Estudo da interceptação da água de chuva em cobertura florestal Amazônica do tipo Terra-Firme. Acta Amazônica, 12: 327–332.

Franken, W., Leopoldo, P.R., Matsui, E. and Ribeiro, M.N.G., 1982b. Interceptação das precipitates em florestas Amazônica de Terra-Firme. Acta Amazônica, 12: 15–22.

Gat, J., 1971. Comments on stable isotope method in regional groundwater investigations. Water Resour. Res., 7: 980–993.

Gat, J.R. and Matsui, E., 1991. Atmospheric water balance in the Amazon Basin: an isotopic evapotranspiration model. J. Geophys. Res., 96: 13179–13188.

Jordan, C.Y. and Heuneldop, L., 1981. The water budget of an Amazonian rain forest. Acta Amazônica, 11: 87–92.

Leopoldo, P.R., 1981. Aspectos Hidrológicos de Floresta Amazônica Densa na Região de Manaus: Repartição da Chuva e Respectivas Composições em Isótopos Estáveis, 1981. Tese de Livre Docencia, UNESP-Botucatu.

Leopoldo, P.R., Franken, W. and Salati, E., 1982a. Balanço hídrico de pequena bacia hidrográfica em floresta Amazônica de terra-firme. Acta Amazônica, 12: 333–338.

Leopoldo, P.R., Franken, W., Matsui, E. and Salati, E., 1982b. Estimativa da evapotranspiração de floresta Amaz ônica dc Terra-Firmc. Acta Amazônica, 12: 23–28.

Lesack, L.F.W. and Melack, J.M., 1995. Flooding hydrology and mixture dynamics of lake water derived from multiple sources in an Amazon floodplain lake. Water Resour. Res., 33: 329–345.

Lloyd, C.R. and Marques, S.O.F., 1988. Spatial variability of throughfall and stemflow measurements in Amazonian rain forest. Agric. For. Meteorol., 42: 63–73.

Marques, J., Salati, E. and Santos, J.M., 1980. Cálculo da evapotranspiração real na bacia Amazônica através do método aerológico. Acta Amazônica, 10: 357–361.

Martinelli, L.A., Victoria, R.L., Matsui, E., Richey, J.E., Forsberg, B.R. and Mortatti, J., 1989. The use of oxygen isotopic composition to study water dynamics in Amazon floodplain lakes. Proceedings of a Regional Seminar for Latin America on the Use of Isotope Techniques in Hydrology, 1987, Mexico City. International Atomic Energy Agency, Vienna, TecDoc-502, pp. 103–112.

Matsui, E., Cervellini, A., Friedman, I. and Salati, E., 1980. Isotope Hydrology in Amazonia, part 1: $\delta^{18}O$ and δD in river waters. Energ. Nucl. Agric., 2: 101–132.

Molion, L.C., 1979. A climatonomic study of the energy and moisture fluxes of the Amazon basin with considerations of deforestation effects. Ph.D. Thesis, University of Wisconsin, Madison.

Paegle, H. 1987. Interactions between convective and large-scale motions over Amazônia. In: R. Dickinson (Editor), The Geophysiology of Amazônia. Wiley, New York, pp. 347–387.

Ramesh, R. and Sarin, M.M., 1992. Stable isotope study of the Ganga (Ganges) river system. J. Hydrol., 139: 49–62.

Ribeiro, M.N.G., Salati, E., Villa Nova, N.A. and Demetrio, C.G.B., 1982. Radiação solar disponivel em Manaus (AM) e sua duração com o brilho solar. Acta Amazônica, 12: 339–346.

Ribeiro, A., Victoria, R.L., Martinelli, L.A., Moreira, M.Z. and Roberts, J., 1996. The isotopic composition of the atmospheric water vapor inside a canopy in the Amazon forest: vertical and diurnal variation. In: J.H.C. Gash, C.A. Nobre, J. Roberts and R.L. Victoria (Editors), Amazonian Deforestation and Climate. Wiley, Chichester, in press.

Richey, J.E., Hedges, J.I., Devol, A.H., Quay, P.D., Victoria, R.L., Martinelli, L.A. and Forsberg, B.R., 1990. Biogeochemistry of carbon in the Amazon River. Limnol. Oceanogr., 35: 352–371.

Salati, E., Dall'Olio, A., Gat, J. and Matsui, E., 1979. Recycling of water in the Amazon basin: an isotope study. Water Resour. Res., 15: 1250–1258.

Senna, H.B., 1992. Balanço de Macro Elementos e Composicão Isotópica de Uma Bacia Hidrográfica na Amaz ônia Central. Master Thesis, University of São Paulo.

Shukla, J., Nobre, C. and Sellers, P., 1990. Amazon deforestation and climate change. Science, 247: 1322–1325.

Shuttleworth, W.J., Gash, J.H., Lloyd, C.R., Moore, C.L., Roberts, J., Marques, P., Fisch, G., Silva, P.V.P., Ribeiro, M.N.G., Molion, L.C.B., Sá, L.D.A., Nobre, J.C., Cabral, O.M.R. and Moraes, J.C., 1984. Eddy correlation measurements of energy partition for Amazonian forest. Q. J. R. Meteorol. Soc., 110: 1143–1162.

Sippel, S., Hamilton, S.K. and Melack, L.M., 1992. Inundation area and morphometry of lakes on the Amazon River floodplain, Brazil. Arch. Hydrobiol., 124(4): 385–400.

Skole, D. and Tucker, C., 1993. Tropical deforestation and habitat fragmentation in the Amazon: satellite data from 1978 to 1988. Science, 260: 1905–1910.

Victoria, R.L., Martinelli, L.A., Mortatti, J. and Richey, L.E., 1992. Mechanisms of water recycling in the Amazon Basin: isotopic insights. Ambio, 20: 378–391.

Villa Nova, N.A., Salati, E. and Matsui, E., 1976. Estimativa da evapotranspiração na bacia Amazônica. Acta Amazônica, 6: 215–228.

Wright, I.R., Gash, J.H.C., Rocha, H.R., Shuttleworth, W.J., Nobre, C.A., Maitelli, G.T.M., Zamparon, C.A.G.P. and Carvalho, P.R.A., 1992. Dry season micromcteorology of central Amazonian ranchland. Q. J. R. Meteorol. Soc., 118: 1083–1099.