Hydrology and Earth System Sciences, 1(2), *279–290* (1997) © EGS

Hydrology & Earth System Sciences

Soil water storage and groundwater behaviour in a catenary sequence beneath forest in central Amazonia.

II. Floodplain water table behaviour and implications for streamflow generation.

M. G. Hodnett[1], I. Vendrame[2], A. De O. Marques Filho[3], M. D. Oyama[4] and J. Tomasella[4]

[1] Institute of Hydrology, Wallingford, OX10 8BB, UK
[2] CTA, Instituto Tecnico Aeroespacial, 12228-904, São José dos Campos, SP, Brazil
[3] Instituto de Pesquisas da Amazônia, CP478, 69011 Manaus, AM, Brazil.
[4] CPTEC, INPE, CP 01, 12630-000, Cachoeira Paulista, SP, Brazil

Abstract

Valley floor groundwater level data collected during the ABRACOS project (Gash et al. 1996), and published streamflow data from small forested catchments in geomorphologically similar areas nearby have been analysed to improve the understanding of the processes of streamflow generation. Early in the wet season, the floodplain water table is typically at 0.8 m depth, or less, and receives only local, vertical recharge. Large storms may create a groundwater ridge beneath the floodplain, temporarily creating a gradient in the direction of the hillslope. Later in the wet season, floodplain water levels are controlled primarily by the discharge of groundwater which maintains the dry season streamflow. The groundwater is recharged by deep drainage from beneath the plateau and slope areas once the dry season soil water deficit has been overcome. In the late wet season, the water level is almost at the floodplain surface and may create seeps on the lower slopes in very wet years. For the period 1966–1989, the recharge was estimated to range from 290 mm to 1601 mm with a mean of 1087 mm. Published data show that baseflow is 91% of annual runoff. Stormflow is generated on the floodplain, and water table recessions after rainfall events show that the runoff response depends on the depth to the water table. These results are from areas with deeply weathered and permeable soils; in areas of Amazonia with shallower soils, the predominant flow generation processes will differ (Elsenbeer and Lack, 1996).

Introduction

The impacts of Amazonian deforestation on climate are being investigated extensively, but the impacts on the processes of streamflow generation and catchment water yield have barely been examined. Grayson *et al.* (1992) and Bonell (1993) have stressed the importance of process studies as the basis of conceptualisation in hydrological modelling, and baseline studies on the processes of streamflow generation in undisturbed forest are essential to permit the modelling of the impacts of deforestation.

The Amazon basin covers 6.1×10^6 km^2 but few hydrological process and small catchment studies have been carried out. Streamflow data have been published for only limited periods for three small catchments in central Amazonia, all on the same Tertiary sedimentary formation, with similar soils and geomorphology (Leopoldo *et al.*, 1982, 1995, Lesack, 1993). Most of the process studies and streamflow data are from the Barro Branco, a 1.3 km^2 catchment in the Reserva Ducke, a forest reserve near Manaus, for the periods 1976–1977 and 1981–1983 (Franken 1979, Franken and Leopoldo, 1984, 1987, Nortcliff and Thornes, 1978, 1981, 1984, 1988, Leopoldo *et al.*, 1982, 1984, 1985, 1995). Elsewhere in Amazonia, studies have been reported by Ross *et al.* (1990) and Nortcliff *et al.* (1990) for a study area in Roraima in N.Brazil and Elsenbeer *et al.* (1990) for a small (1 ha) basin in steep terrain in Peru.

Leopoldo *et al.* (1995) presented 3 years of monthly rainfall, baseflow and 'direct surface runoff' data from the Barro Branco catchment, which has a well defined floodplain. In each of the 3 years, baseflow was consistently 91% of the total runoff, even though annual runoff (calendar years) varied from 909 mm in 1982 to 459 mm in 1983. On a monthly basis, the baseflow varied from 79% to 100%. The largest monthly total runoff, in April 1982, was 134.8 mm of which direct runoff was 18.5 mm, (13.7%), when the rainfall was 379 mm. The runoff in the wettest month was only 1.9 times greater than that in the driest month, even though the rainfall was almost 16 times higher. Streamflow is sustained throughout the dry season and may exceed rainfall; in August 1992, the rainfall, direct runoff and baseflow were 48 mm, 1 mm and 51.5 mm respectively. The lowest monthly baseflow was 24.8 mm (equivalent to 0.83 mm day^{-1}).

Leopoldo *et al.* (1982) presented data from the 23.5 km^2 Bacia Modelo (or Tarumã-Açu) catchment, about 70 km north of Manaus. It also has a well defined floodplain (Chauvel *et al.*, 1987). The rainfall and total runoff for the period from February 1980 to February 1981 were 2089 mm and 541 mm respectively and the data showed a seasonal pattern similar to that of the Barro Branco. Baseflow was not separated.The rainfall was 353 mm less than the mean (1966–1992) for Reserva Ducke, which, if representative of the area of Bacia Modelo, indicates that the runoff was also below average. Over the driest 3 months (July to September) the rainfall was less than evapotranspiration, but the runoff was sustained, decreasing from 1.27 mm day^{-1} to 1.01 mm day^{-1}. These rates are very similar to the dry season rates observed for the much smaller Barro Branco catchment. The mean flow rate in the wettest month (1.88 mm day^{-1}) was less than twice that of the driest month (1.09 mm day^{-1}), despite a factor of 5 difference in rainfall.

Lesack (1993) studied streamflow and hydrochemistry over a 1 year period in a 23.4 ha catchment flowing into a floodplain lake near Manaus. This catchment has a mainly V-shaped cross-section and a floodplain only in its lowest reaches. Baseflow in the very wet year studied (total 2870 mm) was 95%. At the start of observations in February 1984, baseflow was equivalent to 0.37 mm day^{-1}, reached 7.5 mm day^{-1} at the end of the wet season and then decreased to 1.8 mm day^{-1} at the start of the next recharge period. The latter high rate may reflect the large amount of recharge in the 1984 season. The very low direct surface runoff observed in this catchment, and in the Barro Branco, indicates that most of the throughfall infiltrates. This is confirmed by the observations of Chauvel (1982) for the oxisol soils under forest in the surrounding area and by Nortcliff and Thornes (1984), who measured negligible surface runoff on the slopes within the Barro Branco catchment. The sustained flow from these catchments in the dry season demonstrates the importance of catchment storage (in the saturated and deep unsaturated zones) in the hydrological behaviour of these areas with deeply weathered and relatively permeable soils.

In this paper, the short-term dynamics of the water table response to individual storms are examined. Published conclusions about streamflow generation in this area are reviewed and the streamflow data from the Barro Branco and Bacia Modelo (Leopoldo *et al.*, 1995, 1982) are analysed using a water balance approach. These results are presented and interpreted with the groundwater level data to determine the processes of streamflow generation and the role of the floodplain in this region of central Amazonia.

Methods

SITE, LOCATION AND SOILS

The forest site, location, and soils have been described by Hodnett *et al.* (1995, 1996a, 1997). The floodplain is similar to that described by Nortcliff and Thornes (1988) in the Reserva Ducke: 'floodplains have uneven surfaces with local depressions within which water lies during wetter periods'. The depressions at this site still contained water at the end of the 1990 dry season, when the driest conditions in the study period were observed on the plateau (Hodnett *et al.* 1995).

INSTRUMENTATION

The site and the other instrumentation has been described by Hodnett *et al.*, (1995, 1997). In September 1990, 7 dipwells were installed in a transect from the lower slope onto the valley floor. Each dipwell was paired with a neutron probe access tube. Table 1 shows the depths of the dipwells and the heights of the measuring points above the valley floor datum. In October 1992, when water levels were deeper than at any time previously in the study, 4 new dipwells were drilled to below the water table within 0.4 m of the original dipwells D1–D4. Water levels were measured until July 1993 using pressure transducers connected to a Campbell CR10 data logger set to log at 10 minute intervals.

CALCULATION OF DEEP DRAINAGE

A simple water balance model (Hodnett *et al.*, 1996b) calibrated using soil water storage data (1990–93) from the ABRACOS forest site was used to predict deep drainage from the profile using the daily rainfall record from Reserva Ducke for 1966–1992. Drainage totals were calculated for calendar years and for 'seasons', defined as the period from the onset of drainage in the wet season through to the cessation of drainage in the early dry season, typically from November to June. These seasonal totals will be better correlated with the baseflow discharge because, in Central Amazonia, a drainage figure for a calendar year combines the drainage from the latter part of one period and the start of the next. Although the Barro Branco catchment is within the Reserva Ducke, the rainfall record used to estimate the deep drainage was from a different gauge from that used by Leopoldo *et al.*, (1995). The monthly totals were similar, but in February and April 1982 the rainfall was much lower in the gauge used by Leopoldo *et al.*; for these two months, the deep drainage was adjusted downwards by the difference in rainfall between the gauges.

DATA ANALYSIS—PUBLISHED FLOW DATA

The monthly rainfall and runoff data for the Bacia Modelo and Barro Branco catchments (Leopoldo *et al.*,1982,1995)

Table 1 *Dipwell locations and depths*

Tube number	Adjacent access tube number	Depth (m)	Ground level rel to datum	Remarks	New dipwell, with pressure transducer
1	24	3.95	3.36	Furthest upslope	21/10/92
2	25	2.07	1.42	Slope	21/10/92
3	27	1.92	0.79	Slope	21/10/92
4	26	0.80	0.08	Valley floor, foot of slope	21/10/92
5	28	0.93	0.44	Valley floor	—
6	29	0.65	0.04	Valley floor	—
7	30	0.65	0.0	Valley floor	—

have been examined using the approach applied by Klemes (1983). Rainfall-runoff relationships were plotted and examined for trends in behaviour. Catchment storage has already been identified as playing an important role, and the relationship between catchment storage (S_c) and runoff (Q) was investigated. Monthly values of S_c were derived using the water balance:

$$\Delta S_c = R - E - Q \qquad (1)$$

where R = rainfall and E is the evapotranspiration. For Bacia Modelo it was assumed that the mean evapotranspiration rate of 4.1 mm day^{-1} (including interception), derived by Leopoldo *et al.* (1982), applied to each month. The values of S_c calculated for this catchment include the groundwater, deep unsaturated zone, and soil water storage (accessible to the plants). The latter should not influence the baseflow directly, but it will affect the timing of the onset of recharge.

For the Barro Branco, baseflow (Q_b) and surface runoff (Q_s) data were available. The monthly deep drainage (D_d) was estimated and allowed the deep unsaturated zone/groundwater storage (S_g) to be estimated from the groundwater balance:

$$S_g = D_d - Q_b \qquad (2)$$

It was expected that the baseflow from the catchment (Q_b) would show a better relationship to the groundwater storage (S_g) than to the catchment storage (S_c) which includes the soil water.

Results

WATER LEVELS AT THE STUDY SITE

The weekly monitored water level and gradient data have been presented by Hodnett *et al.*, (1997). Except during 1992, following a late and weak wet season, the water table did not fall more than 0.8 m below the surface during the study period. It was deepest in the early wet season and nearest to the surface (0.1 m) at the beginning of the dry season. During 1992 the water level fell to 1.8 m below the floodplain surface (the lowest in the study); the 10 minute data recording started at this time and followed the recharge throughout the wet season and the early part of the recession in the following dry season.

10 minute water level data

The water levels, and the gradients between the dipwells, are shown in Fig. 1 for the period from 21 October 1992 to 22 July 1993. Subsets of the data are plotted (a) in time series, and in some cases (b) as 'cross-sections' for individual events to illustrate the differences in response to rainfall through this period, as the water table rose. These events are marked in Fig. 1, and are shown in: Fig. 2, the first major event on 3 November 1992 (Day 308), Fig. 3, a small early season event on 15 November (320), Fig. 4, a large mid season event on 31 January 1993 (397) and Fig. 5, a late season event on 6 June (523).

General pattern

The early wet season 10 minute data showed a similar pattern to the weekly recorded data. Large and very rapid rises in water level occurred in response to rainfall, and the gradients remained small (notably between D1 & D2 and D2 and D3), except for a few days after each storm, when, in some cases, the gradients reversed. This pattern continued until about 14 January 1993 (Day 380) when the gradient began to increase steadily between D2 and D3, indicating the onset of recharge beneath the plateau and slope. The gradient increased from about 0.01 m m^{-1} in January to about 0.04 m m^{-1} in early March (about Day 430). Although the D1 water level data are missing for this period, there was a marked rise in water level, and a change in gradient from 0.004 m m^{-1} to 0.02 m m^{-1}.

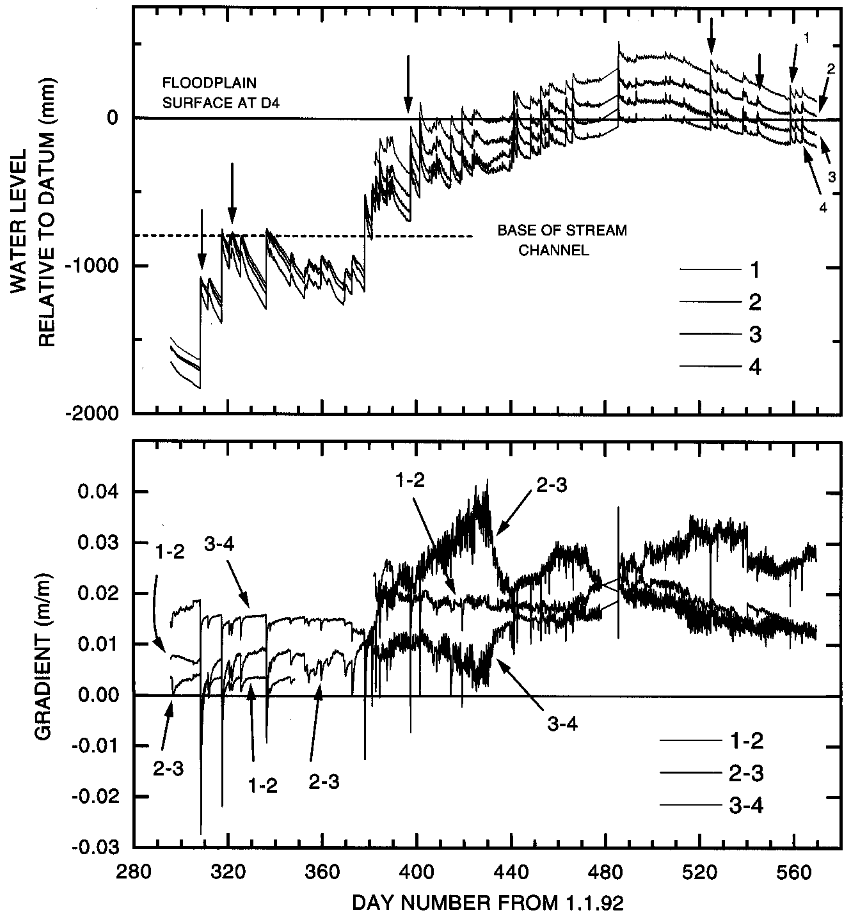

Fig. 1 *Water level data from 4 dipwells logged at 10 minute intervals from 21 October 1992 to 22 July 1993. The tubes are in sequence down the foot of the slope from D1, 2.57 m above the floodplain, to D4, on the floodplain itself. Also shown are the water table gradients between the dipwells.*

Closer to the floodplain, between D3 and D4, the gradient decreased as the water levels rose.

After Day 380, there was a general trend of increasing water levels; subsequent rainfall events led to relatively smaller rises, with a rapid rise to peak, and a more rapid recession than earlier in the season. The recession after each recharge event was usually curtailed by another storm but, after Day 430, a recession of 2–3 days was followed by a period of static water levels, even after 10 days without rain. In this period, the gradient between D2 and D3 decreased sharply, but increased between D3 and D4; it may have reached a less permeable zone causing a greater rise at D3 and a reduction in the gradient upslope.

After 11 April 1993 (Day 467) the levels in all dipwells increased for about 10 days although there were no rainfall inputs large enough to produce significant recharge on the floodplain. This rise must reflect continuing recharge through the deep unsaturated zone from beneath the plateau and slope areas at a rate greater than the rate of discharge beneath the floodplain to the stream. After 1 May (Day 486) the water level at D4 (on the floodplain) reached the surface during rainfall events and remained within 0.1 m of the surface for about 1 month. Levels in the other dipwells also reached a 'ceiling' at the same time, followed by a slow decline of about 0.2 m over the next 65 days. In this period, large rainfall events such as on Day 523 (65 mm) caused only small and short-lived peaks. In late July, the floodplain water table was still within 0.3 m of the surface. The transmissivity of the floodplain deposits must reach a maximum once the water level is at the surface and limit the discharge towards the stream. The discharge can only increase further if there is a significant rise in water level beneath the hillslope.

Individual events—early wet season The water table fell to the lowest level recorded in the study (1.83 m below the valley floor) on 3 November 1992, just before a storm of 135 mm recorded at the adjacent pasture site. The water levels before, during, and after this event are shown in Figs 2a and 2b. Before the storm, there was a gradient from the hillslope toward the floodplain. 35 minutes after the start of the storm, when 55 mm of rain had fallen, the water table began to rise on the floodplain (D4) creating a groundwater 'ridge'. During this period, a mean 15 minute

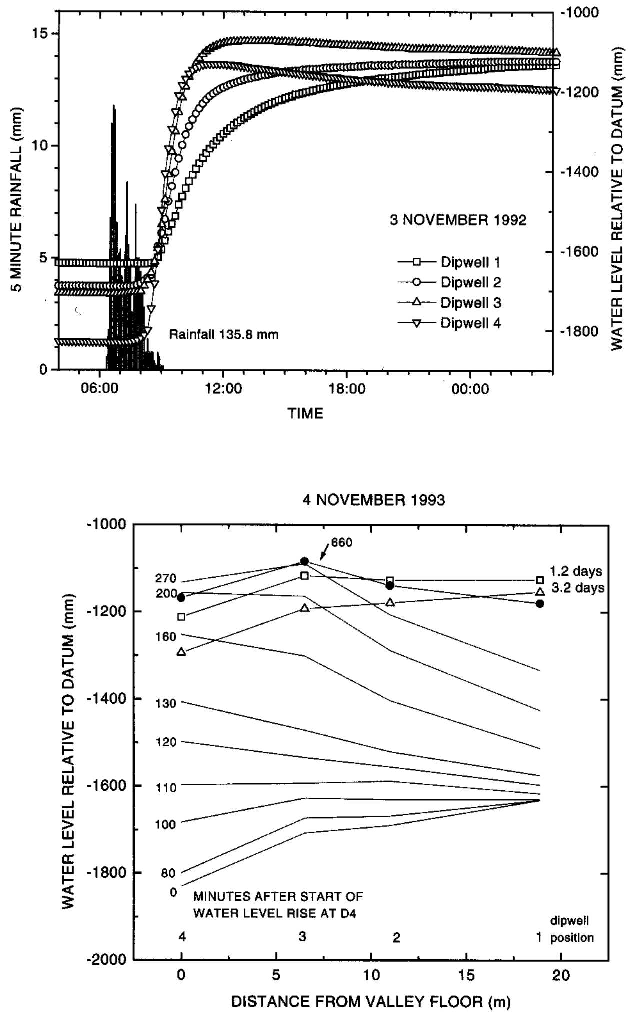

Fig. 2(a) *10 minute water level data for the storm on 3 November 1992 (135 mm total rainfall)* (b) *Cross-section showing the water table response during and after the storm on 3 November 1992.*

rainfall intensity of 138 mm h^{-1} was recorded. The water table then began to rise in D3, D2 and D1 (the furthest up the slope) in sequence, with a lag of 70 minutes between the start of the rise in D4 and D1. The data suggest that the start of the rise at D1 was the result of the reversal of the water table gradient, causing flow from the floodplain to beneath the hillslope. The water table at D4 rose slowly initially and only began to rise rapidly 75 minutes after the start of the storm, when 92 mm of rain had fallen.

After about 4 hours, the gradient between D3 and D4 reverted to being towards the valley but, between D1 and D3 it remained 'away from' the valley for about 30 hours before reversing. The rate of water level rise was fastest at D4 and progressively slower up the slope. Peak water levels occurred after 4.7, 6.0, 19 and 31 hours in D4, D3, D2 and D1, respectively and remained below the level of the base of the stream channel. In this event, the rise in water level beneath the slope (and by extrapolation, beneath the plateau also) was caused by recharge on the floodplain and subsequent groundwater movement to below the plateau.

In the next storm 9 days later, 68 mm of rain fell in the first hour, with a total event rainfall of 71 mm. The time from the onset of rain to the peak water level was reduced to 9 and 19 hours in D2 and D1 but was unchanged in D3 and D4. In this, and all subsequent events, the water table started to rise simultaneously (within a 10 minute period) in all 4 dipwells and the rate of rise was always fastest at D4. Figure 3 shows the simultaneous rise in response to an event of 19 mm on Day 319. By-pass flow through macropores may be responsible, although a lag in response might be expected because of the large difference in the depth of unsaturated zone at the 4 dipwells. There was no reversal of gradient as occurred in the larger events. The simultaneous rise may occur because recharge raises levels first on the floodplain, reducing the gradient upslope and causing water moving down-gradient from beneath the hillslope to 'back-up' as a result.

Individual events, mid and late wet season Figures 4a and 4b show the last event when gradient reversal occurred, on 31 January 1993 (Day 397). The reversal was very brief (80 minutes) and mainly limited to between the floodplain and the foot of the slope (D3 and D4), not between all of the dipwells, as on Day 308. In the study period, gradient reversal was observed on 5 occasions, but only during large events during the first part of the recharge period when the water table was deeper than 0.5 m on the floodplain.

Figure 5a shows the time-series data and cross sections for the storm on 6 June 1993 (Day 523) in which 57 mm fell in the first 50 minutes, with 36.9 mm in 15 minutes. Gradient reversal could not occur because the water level at D3 was already above the level of the floodplain when the event began and water ponding on its surface could discharge readily to the stream. Following the larger events, D3 usually showed the largest rise compared to the

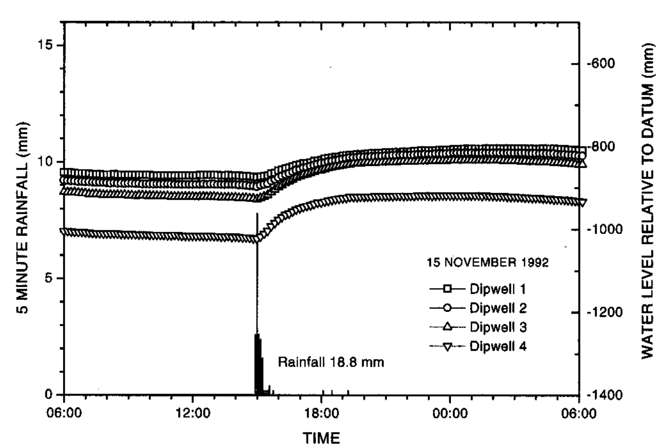

Fig. 3 *10 minute water level data for the storm on 15 November 1992 (rainfall 19 mm)*

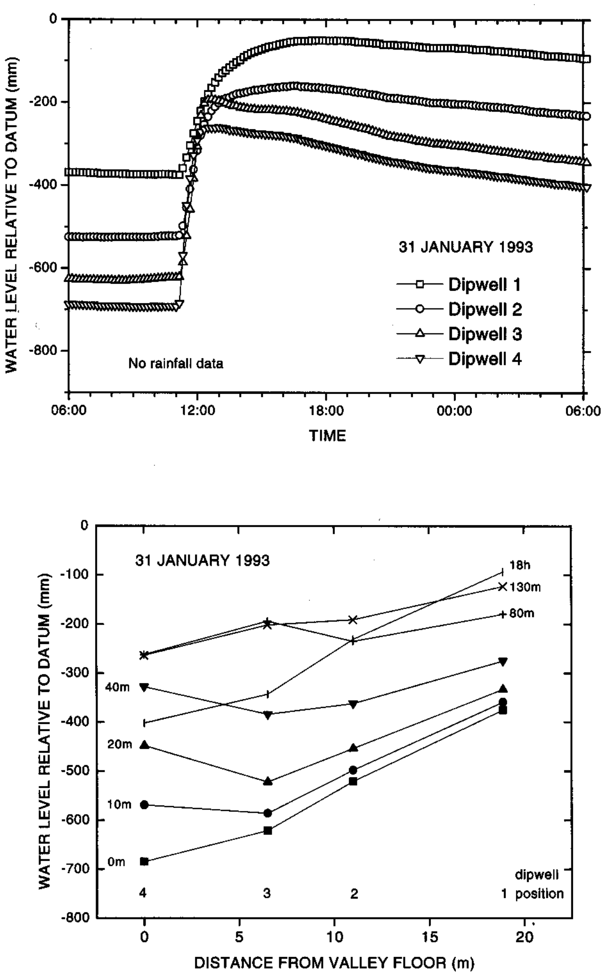

Fig. 4(a) *10 minute water level data for the storm on 31 January 1993 (rainfall not known)* (b) *Cross-section showing the water table response during and after the event on 31 January 1993.*

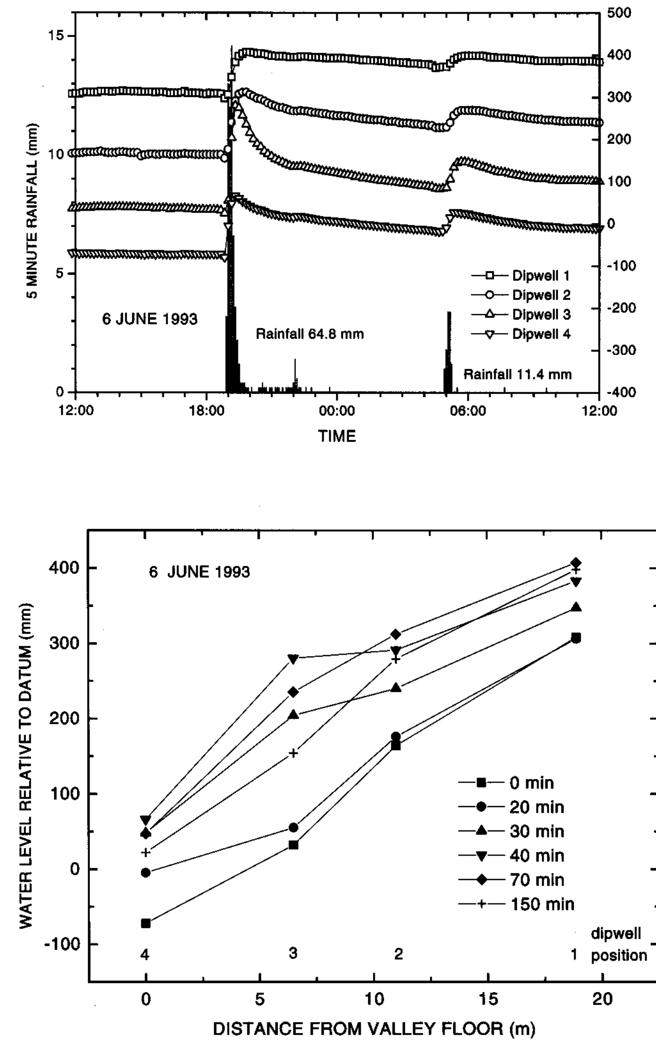

Fig. 5(a) *10 minute water level data for the storm on 6 June 1993 (rainfall 57 mm in 50 minutes, total 64.8 mm).* (b) *Cross-section showing the water table response during and after the event on 6 June 1993.*

other dipwells, which may reflect the arrival of interflow at the foot of the slope. Table 2 compares the maximum water table rise, and time to peak level, at each of the 4 dipwells, for 2 storms of similar magnitude early and late in the recharge period, on 12 November 1992 (Day 317) and 6 June 1993 (Day 523) respectively. As the general water table level rose, the response time became more rapid and the peaks of lower magnitude.

Floodplain soil/ground water storage

Figure 6 shows the water table depth plotted against the mean profile storage to 1 m depth for the access tubes adjacent to D4, D6 and D7. When the water table was below the manually read dipwell, data from the transducer equipped dipwell were used. Two sets of points are shown; one for the period before the water table fell below the dipwells (September 1990–November 1991), and the other for the period after the water table had risen again, following the 1992 drought. The sets show linear, but different, relationships between storage and water table depth. Before the drought, about 30 mm of rain would have raised the water table from a depth of 0.5 m to the surface, compared to 40 mm after the drought. On the pre-drought curve, there was approximately 15 mm of storage between the

Table 2 *Water table response to rainfall at Dipwell 1 (on the slope) and Dipwell 4 (on the floodplain) for storm of similar size and duration, early and late in the wet season.*

Date/Tube	Start WL (m bgl)	End WL (m bgl)	Rise (m)	Time to peak (hours)
317/ D1	1.65	1.07	0.58	4.7
317/ D4	4.08	3.66	0.42	18.7
524/ D1	0.18	0.03	0.15	0.5
524/ D4	2.55	2.44	0.11	1.0

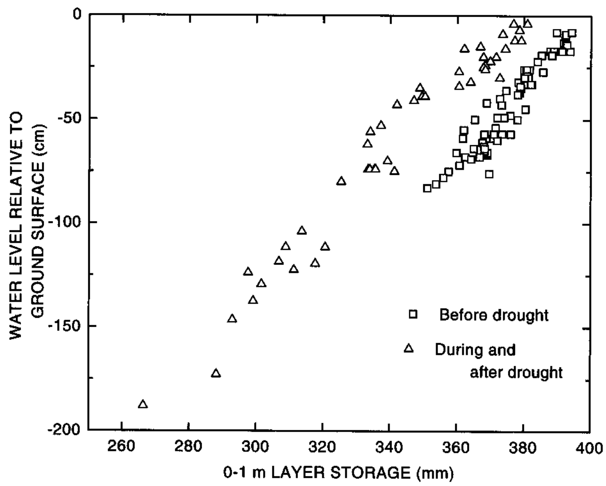

Fig. 6 *Relationship between floodplain water level and soil water storage for the period before and after the 1992 drought.*

surface and the water table at 0.2 m, compared to 65 mm between the surface and 1m depth.

Although the water table reached the same level in 1993 as in 1991, the water content of the layers below 0.3 m remained consistently lower than in 1990–91. In most years, the profile below 0.5 m remains saturated, or very close to saturation, throughout the year but, in 1992, it remained unsaturated for a year. The additional root growth which would have been required to take up water from greater depth, and the drying caused, could have led to a change in soil properties. Air entrapment as the water table rose after the drought may also have contributed. However, the plateau and slope water content data showed no evidence of a similar change in properties; wet season water contents were similar in all years.

DEEP DRAINAGE ESTIMATES

Figure 7 shows the seasonal deep drainage/recharge estimated from the Reserva Ducke daily rainfall record for the period from 1966 to 1989. The mean seasonal deep drainage was 1087 mm (SD 306 mm), with a range from 290 mm in 1982–83 to 1601 mm in 1987-88. The SDs of both the deep drainage and the mean annual rainfall are the same, even though the rainfall is much higher (2442 mm). If, on an annual basis, baseflow is consistently 91% of total runoff, these data indicate that the mean annual total runoff must be about 1190 mm. The monthly streamflow record for the Barro Branco from January 1981 to December 1983 (Leopoldo *et al.* 1995) is examined below. The estimated deep drainage totals for the 3 wet seasons in that period were all below average (823 mm, 1073 mm and 290 mm) and included the lowest seasonal deep drainage in the record (1982-83). The next lowest seasonal deep drainage was 657 mm.

ANALYSIS OF PUBLISHED DATA

Barro Branco

Figure 8 shows a rainfall-runoff plot of the monthly total runoff and direct surface runoff data from the Barro Branco (Leopoldo *et al.* 1995) with the points labelled with month numbers (1–36). There is no unique relationship

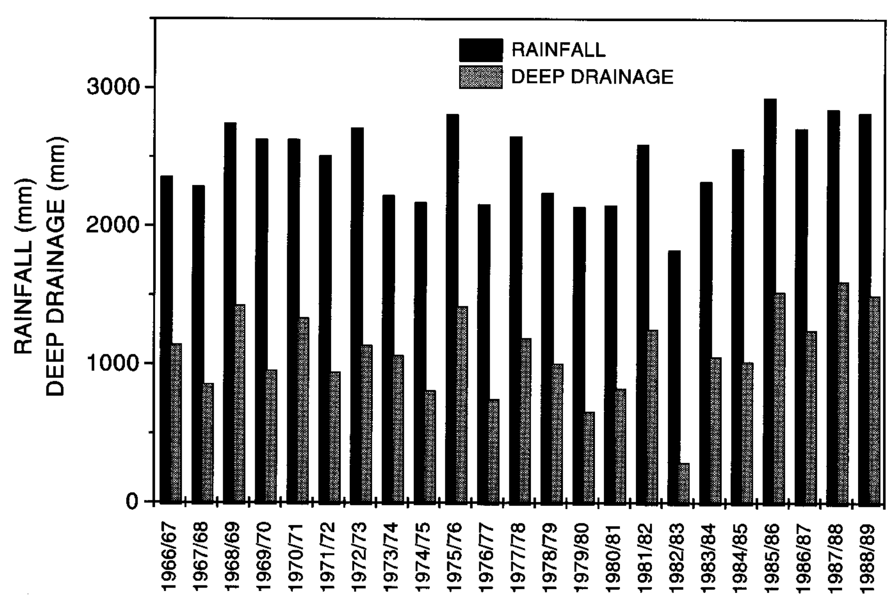

Fig. 7 *Seasonal rainfall and deep drainage for the period from 1966 to 1992 estimated from the daily rainfall record from Reserva Ducke, Manaus (the Barro Branco stream is in this reserve).*

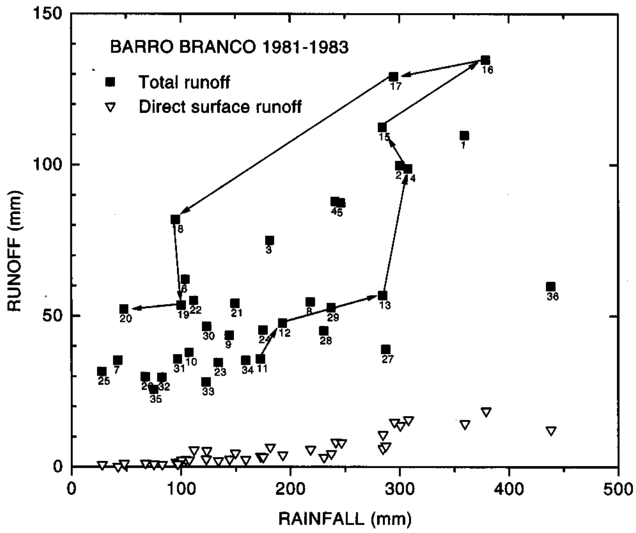

Fig. 8 *Monthly rainfall-runoff plot for the Barro Branco stream, 1981–1983 (data from Leopoldo et al. 1995).*

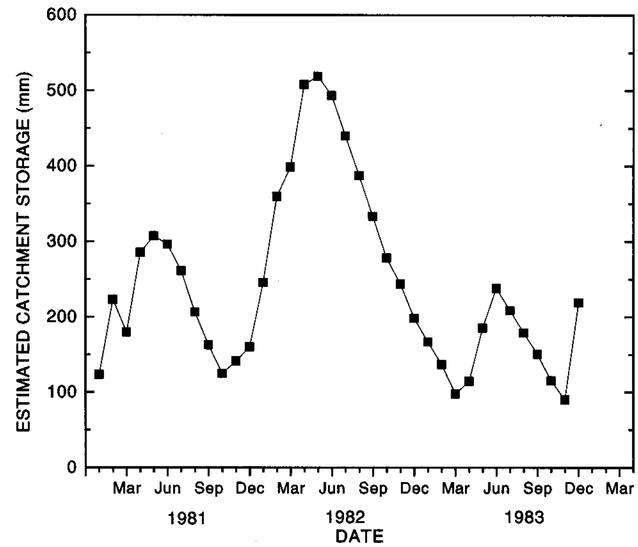

Fig. 9 *Estimated monthly catchment storage for the Barro Branco stream, 1981–1983.*

between rainfall and total runoff, largely because it consists of 79%–100% baseflow and is more dependent on groundwater storage than on rainfall. Despite the scatter of points, there are patterns within the sequence which reflect the important influence of the soil water storage.

The early wet season months (eg 11, 12, 13 and 36) always show a relatively low runoff for a given rainfall. Month 36 had the highest rainfall (438 mm) but a runoff of only 60 mm, in contrast to Month 18, immediately after the wet season, which had a rainfall of 95 mm and a runoff of 82 mm. As the wet season progresses, the ratio of runoff to rainfall increases. This is shown by Months 13–15, in which runoff increased from 57 mm to 112 mm although each had about the same rainfall (300 mm). This reflects the increasing baseflow, as the soil water storage was replenished and recharge to groundwater began. The wet season months in 1983 (27–29) had among the lowest runoff-rainfall ratios because of the exceptionally low recharge. The direct runoff data show a stronger relationship with rainfall; a linear regression gives a slope of 0.046 (r^2 = 0.793). Expressed as a percentage of rainfall, the direct runoff in the early wet season months of January 1982, March 1983 and December 1983, was 2.1%, 2.4% and 2.8% respectively, compared to 4.7%–5.1% for wet months later in the wet season.

Figure 9 shows the large differences in estimated catchment storage (deep unsaturated zone and saturated zone) through the 3 year period from 1981 to 1983. The zero of the storage scale is arbitrary. Between May 1982 and March 1983 there was a decrease of 421 mm over 10 consecutive months which illustrates clearly the role of catchment storage in maintaining streamflow in this area, particularly in drier years. The 10 month period of decreasing storage was the result of the exceptionally low recharge in the 1982–83 wet season. It is not known how the Barro Branco would have responded in a wet season such as that of 1987–88, when the deep drainage was estimated to be 1601 mm.

The Barro Branco direct surface runoff and baseflow data are plotted against the estimated catchment storage (groundwater and deep unsaturated zone) in Fig. 10. The baseflow data show considerable scatter, but the driest

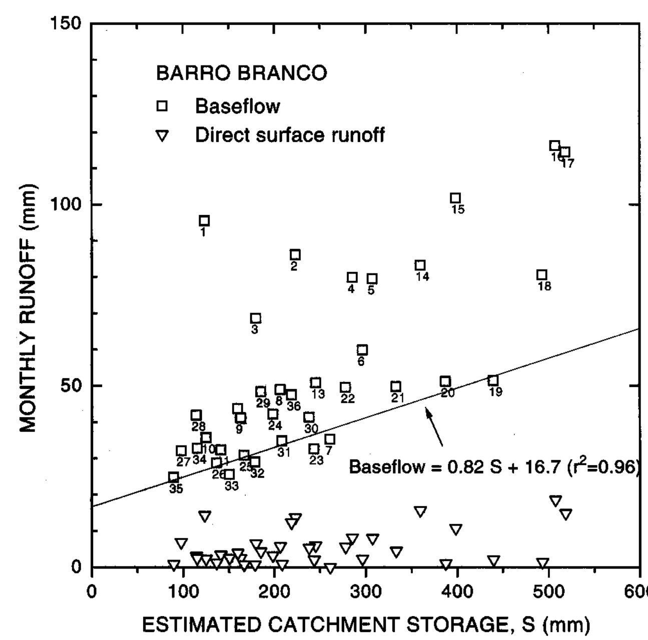

Fig. 10 *Monthly runoff plotted against estimated catchment storage for the Barro Branco stream, 1981–1983 (rainfall and runoff data from Leopoldo et al. 1995).*

months, which also have the lowest direct surface runoff (7, 18, 19, 20, 25, 26, 31, 32 and 35), all lie close to the lower edge of the group. These points probably reflect the relationship between actual baseflow and catchment storage. The straight line fitted to these points (omitting month 18) has an r^2 of 0.965, and if extrapolated, indicates that baseflow would cease only after catchment storage had decreased 204 mm below the arbitrary zero used. The data points for the wetter months lie above this 'base' line, which appears to suggest that another process of flow generation may be operating, which is not related to groundwater storage. Interflow may be responsible.

Bacia Modelo

Figure 11 shows the monthly runoff data for Bacia Modelo plotted against the estimated catchment storage (arbitrary zero) which, in this case, includes the soil water. The estimated maximum catchment storage change was 307 mm. The data show similar patterns to those of the Barro Branco, with the lowest rainfall months, which are likely to have the lowest direct surface runoff, along the lower edge of the group of points. If it is assumed that there was no surface runoff in these months, the regression line through these points can be used as a crude means of separating the baseflow in each month; this gave a direct runoff of 15% and a baseflow of 85% for this 23.5 km^2 catchment.

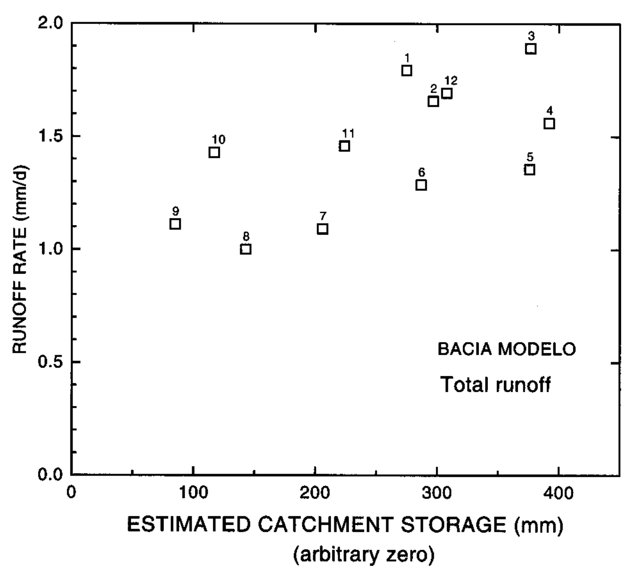

Fig. 11 *Monthly runoff plotted against estimated catchment storage for the Tarumã Açu stream (Bacia Modelo) for 1980–81 (rainfall and runoff data from Leopoldo et al., 1982).*

Discussion

WATER TABLE BEHAVIOUR

The water level data show a very different behaviour in the early wet season compared to the late wet season. Observations began when the water table was unusually deep (1.8 m) below the valley floor and between October 1992 and January 1993 the water level was below the stream bed. During this time, the groundwater could not drain to the stream, except by moving through the floodplain deposits to reach it further downstream. It may also have been draining to the adjacent larger valley. Before the first major rainfall event, the water levels were falling, probably as a result of root uptake. After the event, the early part of the recession was caused by the movement of water away from the groundwater ridge on the valley floor to beneath the hillslope. Later, the recession will have been due to root uptake, and down-valley movement in the floodplain deposits.

After mid-January 1993 (1 month after the soil profiles on the plateau had rewetted to 3.6 m), the groundwater gradients from beneath the slope increased, indicating the onset of recharge from the plateau and slope areas. There were strong indications that a soil water deficit developed in the profile below 3.6 m in the 1992 dry season; this would have had to be replenished before recharge recommenced (Hodnett *et al.*, 1997). After January, the rate of groundwater recession after storms increased, probably because the connection to the stream channel had been re-established as the water table rose. The main influence on the rate of recession would then have been the transmissivity of the floodplain deposits, determined by the water table depth, through the variation in saturated thickness, and as a result of the higher permeability of the upper layers (Nortcliff and Thornes, 1984). The water table recession must be a good indication of the quickflow response in the stream.

PROCESSES OF STREAMFLOW GENERATION

Nortcliff and Thornes (1984, 1988) found that, for the Barro Branco, there was little difference between the percentage of storm rainfall leaving the basin as quickflow in the 1977 wet season (5.04%) and the 1978 dry season (5.6%). Surprisingly, the wet season water levels (0.2 m–0.44 m, in April/May 1977) were slightly lower than the dry season levels (0.1 m–0.3 m, in August/September 1978) but Fig. 7 shows that the deep drainage/recharge for the 1976–77 season was well below average (749 mm) compared to the 1977–78 season, which was above average (1189 mm). This may explain the similarity in the wet and dry season water levels and quickflow percentages in the seasons studied. The Barro Branco flow data (Leopoldo *et al.*, 1995) for the late wet season also showed quickflow percentages of about 5%, but values were lower (2%–3%) in the early wet season when the water table was deeper. Lesack (1993) found quickflow percentages from 0.5%–4% in the mainly V-cross-section 23.4 ha catchment studied, with a volume weighted mean of 2.8%.

The late wet season water level data presented here confirm the hypothesis of Nortcliff and Thornes (1984) that 'the quickflow hydrographs appear to be almost completely produced by saturated overland flow derived from the floodplain areas adjacent to the channel with little direct coupling between the slope and the channel during the quickflow period'. The main process of quickflow generation appears to be saturation overland flow (SOF), where 'the soil becomes saturated by the perennial groundwater table rising to the surface' (Dunne, 1978). The conclusion of Nortcliff and Thornes (1984) that the water table is maintained near the surface by groundwater discharge from beneath the plateau and slope areas has been confirmed by Hodnett *et al.* (1997). In volume terms, on an annual basis, the most important process of flow generation is sub-surface flow, entering the stream through the perennial groundwater body (Dunne, 1983). This flow is not generated on the floodplain, but has to pass through, or beneath it to reach the stream channel. Nortcliff and Thornes (1988) noted that, in this environment, the floodplain hydrology appears to play a far more substantial role in the overall stream hydrology than is normally the case, and that 'a variety of flow generating mechanisms will occur in different environments and in the same environment at different times'.

The quickflow percentages for the late wet season months would indicate that the extent of floodplain in the overall landscape is about 5%. The link between the quickflow percentage and the contributing area will only be unique when the water table is at, or very close to the surface. However, when the water table is deeper, in the late dry season and early wet season (conditions not observed by Nortcliff and Thornes), the groundwater level recessions presented here suggest that the quickflow response will be slower when the water table does not reach the surface.

The estimated deep drainage for the Barro Branco for the 1982–83 wet season was the lowest in the 27 year record, but streamflow was maintained through the following dry season, with a minimum discharge rate of 0.83 mm d^{-1}. The stream at the Fazenda Dimona site dried up in 1992, suggesting that the recharge in 1991–92 was even lower than in the Barro Branco in 1982–83. However, the floodplain studied is within a smaller catchment (whose exact area is not known); as the regional water table falls, baseflow will cease in first order basins (those highest in the landscape) and then in progressively larger basins.

The creation of a groundwater ridge (and the short-term discharge of groundwater from the floodplain to beneath the hillslope) observed on 5 occasions after major rainfall events early in the 1992–93 wet season occurred because the floodplain water table was unusually deep. However, it probably occurs to a lesser extent in the early wet season in most years. Although unusual, these data give a valuable indication of the response if drier years become more frequent as a result of climate change, or if forest clearance leads to reduced infiltration and lower recharge in plateau and slope areas.

INTERFLOW

The role of interflow in these catchments is not clear. The analysis of monthly flow and catchment storage data suggests that there may be a process of streamflow generation (other than SOF from the valley floor) which is independent of groundwater storage. Hodnett *et al.* (1997) have indicated that, at Fazenda Dimona, there is a potential interflow route on the slopes but suggested that very large storms, and/or very wet antecedent conditions would be necessary to allow significant quantities of interflow to reach the valley floor because vertical drainage would limit its duration. Interflow would probably contribute to the recession stage of the quickflow peak as travel distances would be longer than for SOF generated on the valley floor.

The two example wet and dry season storm hydrographs published by Nortcliff and Thornes (1984) show a well defined flow peak. Discharge rose and decayed sharply over a period of about 6 hours, but in both cases, there was a further 2 stage decay lasting about 20 hours before the discharge returned to its pre-storm rate. This suggests another, slightly slower process of flow generation, which is responsible for about the same volume of flow as the quickflow peak. This may be evidence of interflow, but may be the result of discharge of water stored temporarily on the floodplain above the pre-storm water level. The conventional method of baseflow separation is arbitrary (Linsley *et al.* 1958) and it is not known whether the Bacia Modelo baseflow data (Leopoldo *et al.*, 1995) would have included the part of the curve after the main peak. Caution must be exercised in ascribing processes to the different portions of the hydrograph, particularly when different processes may be occurring through the year. The contribution of interflow can probably be assessed only through detailed process studies on hillslopes.

DEFORESTATION AND STREAMFLOW

If the permeability of the soil surface is not reduced when the forest is cleared, a change to pasture will result in more deep drainage because of lower dry season transpiration and reduced interception losses. Following forest clearance Edwards (1979) in East Africa observed increases in streamflow and Peck and Williamson (1987) in Western Australia observed increases in groundwater levels. It is estimated that, in the study area, the increase in deep drainage, and therefore baseflow, would be of the order of 200 mm $year^{-1}$. The higher recharge will increase baseflow, and, by maintaining valley floor water levels near to the surface for longer periods, will lead to an increase in the overall percentages of quickflow.

However, if the soil permeability is reduced, large changes in catchment response could occur because the soil surface might then become the 'throttle' layer (Bonell, 1993). If, on an annual basis, 5% of the rainfall input ran off the whole area directly (as Hortonian overland flow) because of reduced permeability, this would effectively double the annual volume of storm runoff. However, since the surface runoff would occur in the high intensity storms, the proportional increase in peak flows during such events could be many times this amount and would lead to serious flooding and damage. Erosion could become a serious problem, particularly in the channels carrying the flow. Methods of clearance which minimise compression of the surface soil are very important if large increases in flood flows are to be avoided. Although flood flows would increase, recharge, and baseflow, would be reduced proportionally.

Summary and conclusions

Early in the wet season, the water table in the valley floor responds to the local vertical recharge through the unsaturated zone of the floodplain alone. Once the soil water storage in the plateau and slope areas (depleted by plant water uptake in the dry season) has been replenished, deep drainage commences and begins to recharge the groundwater beneath the plateau and slope areas, increasing the groundwater gradient towards the valley and raising the valley floor water levels. At the end of the wet season, valley floor water levels may rise, or remain unchanged close to the surface, for periods of 10 days or more without rain, indicating a lag in the transmission of the recharge through the deep unsaturated zone.

Streamflow measurements from a 1.3 km^2 catchment (Leopoldo *et al.*, 1995) show that baseflow accounted for 91% of the total annual runoff and was maintained through a long dry season with a minimum rate of 0.83 mm day^{-1}. Flow data from a larger catchment showed a slightly higher dry season rate. Streamflow is thus largely determined by the groundwater recharge. For the period 1966–1992, this was estimated, on a seasonal basis, to vary from 290 mm to 1601 mm, with a mean of 1087 mm. The CV of recharge was 28%, compared to 13% for rainfall. The baseflow recession is controlled mainly by the transmissivity of the aquifer and, probably, when levels are high, by the floodplain itself.

Storm runoff is generated on the valley floor and is typically 5% of the rainfall when the water table is close to the surface of the floodplain, but 2%–3% when the water levels are deeper. This study has shown that the floodplain water level typically ranges from 0.8 m below the surface between October and as late as February, to close to the surface in May and June. When the floodplain water table is near the surface, there is little storage to be filled during an event. Resistance to flow over the surface is less than through the floodplain deposits, and runoff response will be very rapid, with the flow peak consisting mainly of overland flow. After the peak flow has occurred, the water temporarily stored in the floodplain deposits which lie above the pre-storm water level will then drain to the stream. In very wet years, the water levels beneath the hillslope may rise higher than observed in this study and extend the contributing areas up the slopes as seeps, leading to enhanced storm runoff and larger quickflow percentages. However, because the slopes are relatively steep, the increase in contributing area would probably be fairly small. If the pre-storm water level is deeper, a slower rise to peak flow is likely as the storage fills, particularly if the water table does not reach the surface.

In this study and those of Nortcliff and Thornes (1984, 1988), the term 'floodplain' is used to describe the largely flat valley floor which may be flooded during in the wet season. In most cases, a floodplain becomes flooded as a result of overbank flow, where the water has originated from the catchment upstream. However, in the environment studied here, the "floodplain" is the contributing area and becomes flooded as a result of the water table reaching the surface during storms, with a possible contribution from interflow.

Acknowledgements

The ABRACOS project was funded by the UK Overseas Development Administration (ODA) in collaboration with the Agência Brasileira de Cooperação (ABC) for whose support the authors are grateful. The study could not have been carried out without the support of the Instituto Nacional de Pesquisas da Amazônia and particular thanks are due to the conscientious group of technicians and observers who made the observations regularly.

References

Bonell, M., 1993. Progress in the understanding of runoff generation dynamics in forests. *J. Hydrol.* **150**, 217–275.

Chauvel, A., 1982. Os latossolos amarelos, alicos, argilosos dentro dos ecossistemas das bacias experimentais do INPA e da regiao vizinha. *Acta Amazonica* 12(3), 47–60.

Chauvel, A., Lucas, Y., and Boulet, R. 1987. On the genesis of the soil mantle of the region of Manaus, central Amazonia, Brazil. *Experientia* **43**, 234–241.

Dunne, T., 1978. Field studies of hillslope flow processes. Ch 7. In: *Hillslope Hydrology*, Ed. Kirkby, M.J. John Wiley, Chichester, UK. 389 pp.

Dunne, T., 1983. Relation of field studies and modeling in the prediction of storm runoff. *J. Hydrol.* **65**, 25–48.

Edwards, K.A., 1979. The water balance of the Mbeya experimental catchments. *E. Afr. agric For. J.* 43, 231–247.

Elsenbeer, H., Cassel, D.K., and Castro, J., 1990. Spatial analysis of soil hydraulic conductivity in a tropical rainforest catchment. *Wat. Resour. Res.*, **28**, 3201–3214.

Elsenbeer, H. And Lack, A., 1996. Hydrological pathways and water chemistry in Amazonian rainforests. Ch 43, in: *Advances in Hillslope Processes* Vol 2. Eds. Anderson, M.G. and Brooks, S.M., John Wiley, Chichester, UK. pp. 939–959.

Franken, W., 1979. Untersuchungen im Einzugsgebiet des zentralamazonischen Urwaldbaches 'Barro Branco' auf der 'terra firme'. I. Abflußverhalten des Baches. *Amazoniana* **6**, 459–466.

Franken, W. and Leopoldo, P.R., 1984. Hydrology of catchment areas of Central-Amazonian forest streams. Ch 19, in: *The Amazon, limnology and landscape ecology of a mighty tropical river and its basin.* Ed. Sioli, H., Junk, Dordrecht, The Netherlands. pp. 501–519.

Franken, W. and Leopoldo, P.R., 1987. Relações entre fluxos de água subterranea e superficial em bacia hidrográphica caracterizada por cobertura florestal Amazônica. *Acta Amazonica* 16/17, 253–262.

Gash, J.H.C., Nobre, C.A., Roberts, J.M. and Victoria, R.L.,1996. An overview of ABRACOS. In: *Amazonian Deforestation and Climate*, Eds Gash, J., Nobre, C.A., Roberts, J.M., and Victoria, R.L. John Wiley, Chichester, U.K. pp. 1–14.

Grayson, R.B., Moore, I.D., and McMahon, T.A., 1992. Physically based hydrologic modeling 2. Is the concept realistic? *Wat. Resour. Res.* **26**,10 2659–2666.

Hodnett, M.G., Pimentel da Silva, L., da Rocha, H.R. and Cruz Senna, R., 1995. Seasonal soil water storage changes beneath Central Amazonian rainforest and pasture. *J. Hydrol.* **170**, 233–254.

Hodnett, M.G., Oyama, M.D., Tomasella, J., and Marques Filho, A. de O., 1996a. Comparisons of long-term soil water storage behaviour under pasture and forest in three areas of Amazonia. In: *Amazonian Deforestation and Climate*, Eds Gash, J., Nobre, C.A., Roberts, J.M., and Victoria, R.L. John Wiley, Chichester, U.K. pp. 57–77.

Hodnett, M.G., Tomasella, J., Marques Filho, A. de O., and Oyama M.D., 1996b. Deep soil water uptake by forest and pasture in central Amazonia: predictions from long-term daily rainfall using a simple water balance model. In: *Amazonian Deforestation and Climate*, Eds Gash, J., Nobre, C.A., Roberts, J.M., and Victoria, R.L. John Wiley, Chichester, U.K. pp. 79–99.

Hodnett, M.G., Vendrame, I., Oyama, M.D., Marques Filho, A. De O., and Tomasella, J. Soil water storage and groundwater behaviour in a catenary sequence beneath forest in Central Amazonia. I. Comparisons between plateau, slope and valley floor. *Hydrol. Earth Sys. Sci.* 1(2), 265–277.

Klemes, V., 1983. Conceptualisation and scale in hydrology. *J. Hydrol.* **65**, 1–23.

Leopoldo, P.R., Franken, W., Matsui, E., and Salati, J., 1982. Estimativa de evapotranspiração de floresta amazônica de terra firme. *Acta Amazonica* (Suppl) 12, 23–28.

Leopoldo, P.R., Franken, W., and Matsui, E., 1984. Hydrological aspects of the tropical rainforest in Central Amazon. *Interciencia* **9**, 125–131.

Leopoldo, P.R., Franken, W., and Matsui, E., 1985. Hydrological aspects of the tropical rainforest in the central Amazon. Ch. 7, in *Change in the Amazon Basin, Vol 1. Man's impact on forests and rivers.* Ed J.Hemming. Manchester University Press, Manchester, U.K. 222 pp.

Leopoldo, P.R., Franken, W.K., and Villa Nova, N.A., 1995. Real evapo-transpiration and transpiration through a tropical rainforest in central Amazonia as estimated by the water balance method. *For. Ecol. Manag.* **73**, 185–195.

Lesack, L.F.W., 1993. Water balance and hydrologic characteristics of a rainforest catchment in the central Amazon basin. *Wat. Resour. Res.*, **29**, 759–773.

Linsley, R.K., Kohler, M.A. and Paulhus, J.L.H., 1958. Hydrology for engineers, McGraw-Hill, New York. 689pp.

Nortcliff, S., and Thornes, J.B., 1978. Water and cation movement in a tropical rainforest environment: 1. Objectives, experimental design and preliminary results. *Acta Amazonica*, **8**, 245–258.

Nortcliff, S., and Thornes, J.B., 1981. Seasonal variations in the hydrology of a small forested catchment near Manaus, Amazonas, and its implications for management. Ch. 2.2, in: *Tropical Agricultural Hydrology*. Eds R. Lal and E.W. Russell. John Wiley, New York. pp. 37–57.

Nortcliff, S., and Thornes, J.B., 1984. Floodplain response of a small tropical stream. Ch. 5 in: *Catchment experiments in fluvial geomorphology*, Eds T.P. Burt and D.E. Walling. Geo Books, Norwich, UK. pp. 73–85.

Nortcliff, S., and Thornes, J.B., 1988. The dynamics of a tropical floodplain environment with reference to forest ecology. *J. Biogeogr.* **15**, 49–69.

Nortcliff, S., Ross, S.M., and Thornes, J.B., 1990. Soil moisture, runoff and sediment yield from differentially cleared tropical rainforest plots. Ch.25, in: *Vegetation and Erosion*, ed J.B. Thornes. John Wiley, Chichester, UK. pp. 419–436.

Peck, A.J., and Williamson, D.R., 1987. Effects of forest clearing on groundwater. *J. Hydrol.*, **94**, 47–65.

Ross, S.M., Thornes, J.B. and Nortcliff, S., 1990. The Maracá rainforest Project, II. Soil hydrology, nutrient and erosional response to the clearance of terra firme forest, Maracá Island, Roraima, northern Brazil. *Geogr. J.* **156**, 267–282.