9

Climate change and hydrological modes of the wet tropics

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9.1 INTRODUCTION

The Amazon Basin is the world's largest drainage system. In fact, 1,100 rivers make up the Amazon system. The Amazon River carries one-fifth of all the river water in the world. The source of the Amazon can be traced to the Apurimac River, located at 5,200 m above sea level. The 1,100 tributaries flow through nine South American countries: Brazil, Bolivia, Peru, Ecuador, Colombia, Venezuela, Guyana, Surinam, and French Guiana. The Amazon River and its tributaries drain most of the area of heavy rainfall. The area is called "Amazonia" and most of it is a sparsely populated rainforest. Most of the Amazon Basin is in Brazil (Amazonia Legal). The Amazon River represents 16% of annual global river runoff (Shiklomanov, 2001).

The Amazon River system is the single largest source of freshwater on Earth and its flow regime is subject to interannual and long-term variability represented as large variations in downstream hydrographs (Richey *et al*., 1989; Vörösmarty *et al*., 1996; Marengo *et al*., 1998a; Marengo and Nobre, 2001; Marengo 2004a, b). A better understanding of rainfall and river variability will depend on the physical mechanisms related to regional and large-scale atmospheric–oceanic–biospheric forcings that impact the temporal and spatial variability of the hydrometeorology of the Amazon Basin. The impacts could be felt on various timescales.

The implementation of field experiments in the region during the last 20 years—such as the ABRACOS (Anglo Brazilian Amazon Climate Observational Study) during the 1980s, the LBA (Large Scale Biosphere Atmosphere experiment in Amazonia), and the SALLJEX (South American Low Level Jet field experiment) during the late 1990s and early 2000s—has allowed for the development of new knowledge on climate and hydrology in the Amazon Basin, including the interaction between land surface processes in rainfall, and the development of regional and global climate models tuned with more realistic representations of physical processes for the region (Gash and Nobre, 1997; Silva Dias *et al.*, 2002; Vera *et al.*, 2006). Moisture transport into and out of the Amazon Basin has also been studied, and regional circulation features responsible for this transport and its variability in time and space have been detected and studied using observations from these field experiments and other global data sets (Marengo *et al.*, 2002, 2004a, b; Vera *et al.*, 2006).

On the basis of what is now known on climate variability in Amazonia and the moisture transport in and out of the basin based on observational studies and model simulations, the question that arises is: What are the possible impacts on the Amazon ecosystem of regional-scale deforestation or the increase of greenhouse gas (GHG) concentrations in the atmosphere and subsequent global warming. The issue of deforestation has been explored in various numerical experiments since the 1980s using atmospheric global climate models—general circulation models (GCMs)—all of which show that the Amazon will become drier and warmer (see reviews in Marengo and Nobre, 2001; Voldoire and Royer, 2004). Even though there are no clear signs of trends for reduction of rainfall in the basin due to deforestation—as suggested by climate models on deforestation scenarios—one study (Costa *et al.*, 2003) has detected changes in the Tocantins River discharges as a result of land-use changes in its upper basin following the construction of the city of Brasilia in the 1960s.

Furthermore, since the early 2000s new developments in atmosphere–ocean–biosphere coupled models—by the Hadley Centre for Climate Research and Prediction in the U.K., the Institute Pierre et Simon Laplace at the University of Paris in France, and the Frontier Research Center for Global Change in Japan—have allowed for better simulation of future climate change scenarios. The new models include interactive vegetation schemes that more realistically represent the water vapor, carbon, and other gas exchange between the vegetation and the atmosphere. Projections for future climate change from the Hadley Centre model have shown that an increase in the concentration of greenhouse gases in the atmosphere will produce changes in vegetation such that Amazonia will become a savanna by the 2050s, and the region will become drier and warmer with most of the moisture coming from the tropical Atlantic such that normally produced rainfall in the region will not find the environment to condense above the savanna vegetation by 2050, and the moist air stream will move to southeastern South America producing more rainfall in those regions. Therefore, after 2050, the Amazon Basin may well behave as a "source of moisture" rather than a sink (like its present day climate) (Cox *et al.*, 2000, 2004; Betts *et al.*, 2004; Huntingford *et al.*, 2004).

Therefore, this chapter is focused on the role of the Amazon Basin in the functioning and modulation of the regional climate and hydrology, in both present and future climates, by means of (a) description of hydrological regimes and maintenance of humidity and import/export of moisture, (b) variability of climate and hydrology in various timescales; and (c) sensitivity of the Amazon system to changes in land use or climate change due to an increase in the concentration of GHG in the atmosphere.

9.2 RAINFALL AND HYDROLOGICAL REGIMES IN THE AMAZON BASIN

9.2.1 Rainfall distribution and seasonal rainfall and river seasonal variability

The Amazon River drains an area of 6.2×10^6 km^2 and discharges an average of 6,300 km^3 of water to the Atlantic Ocean annually. The annual cycle of rainfall in the region has been extensively described in Rao and Hada (1990), Marengo (1992), Figueroa and Nobre (1999), Marengo and Nobre (2001), Liebmann and Marengo (2001), Marengo (2004b). The spatial distribution of rainfall shows three centers of abundant precipitation in the Amazon Basin. One is located in northwest Amazonia, with more than 3,600 mm per year. Another region with abundant rainfall is the central part of Amazonia around 5°S with 2,400 mm per year. A third center is found close to the mouth of the Amazon River near Belém, with more than 2,800 mm per year. In the Rio Negro basin area, in northwestern Amazonia, the rainfall is abundant throughout the year reaching its maximum in April–June, while southern Amazonia's rain peaks earlier (January–March). The extreme high and localized values of precipitation in narrow strips along the eastern side of the Andean slopes are thought to be due to upglide condensation and a rain shadow effect on the lee side, so the localized maximum is due to the easterly winds being lifted when they flow over the Andes. The coastal maximum is caused by nocturnal convergence between the trade winds and the land breeze. In the central-north and south-southeast sections, rainfall is lower—in the region of 1,500 mm. The peak of the rainy season occurs earlier (December–February) in southern Amazonia, while northern and central Amazonia experience maximum rain in March–May (Figure 9.1, see color section).

The seasonal variability in rainfall in Figure 9.1 can be better understood by viewing it with Figure 9.2, which shows the mean seasonal distribution of rainfall. The austral summer—that is, December to February—season is characterized by the peak of the rainy season in southern Amazonia and the dry season in the Amazon region north of the equator, with less than 360 mm in the entire season. March–May represents the peak of the rainy season in central Amazonia, all the way from western Amazonia to the mouth of the Amazon River, and June–August represents the dry season over most of the region, with less than 180 mm in the entire season in southern and eastern Amazonia and less than 360 mm in the entire season in central Amazonia, while the extreme north of Amazonia experiences its wet season.

Following the annual cycle of rainfall, river discharge peaks first in southern and eastern Amazonia (January–March)—as in the Tocantins-Araguaia and Madeira Rivers in Figure 9.3—while the Negro and Amazon River peak in March–May. Measurements taken at Óbidos integrate the contributions of the Solimêes River (southern and western Amazonia) and the Rio Negro (northern Amazonia). Chu (1982) shows that almost 70% of the contribution to Óbidos measurements comes from the waters of the Solimêes River. The records of the Amazon, Negro, Xingú and Tocantins Rivers (Marengo *et al.*, 1998b) are displayed for two El Niño years (1982–83, 1986–87) and La Niña years (1975–76, 1988–89). Extreme years were chosen since they may better show the associations between El Niño and rain in their basins. The

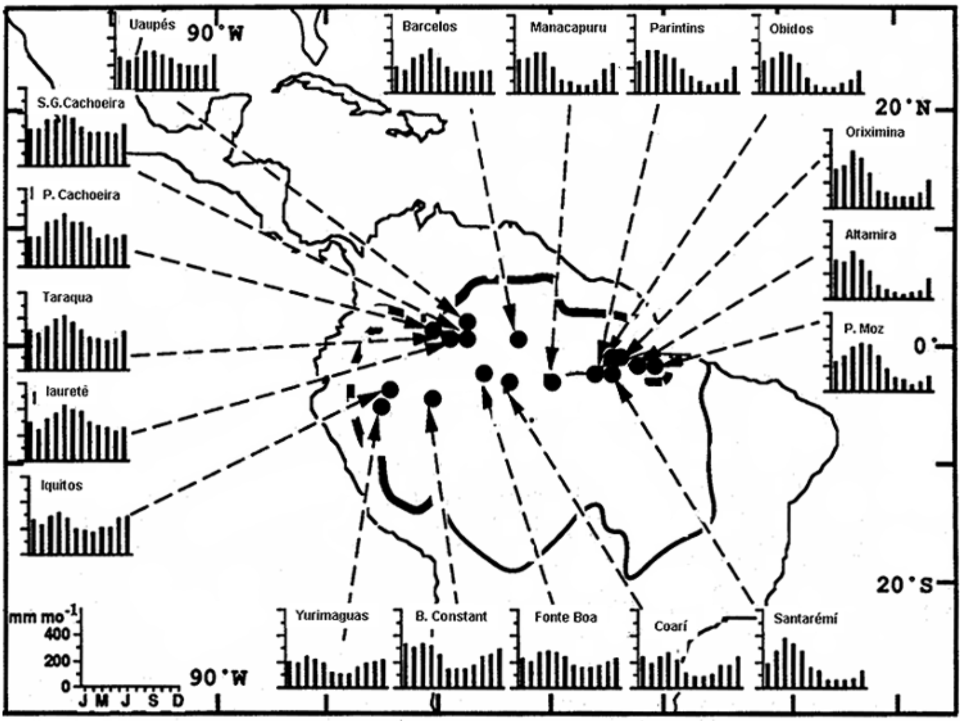

Figure 9.2. Seasonal cycle of rainfall in various stations across the Amazon Basin. Location of stations is indicated by a small square, and units are in mm month^{-1}. Scale is shown at the lower left side of the panel. Modified from Marengo (1992).

year before an El Niño peak, Amazon River discharges at Óbidos are anomalously high, while in actual El Niño years the discharges are lower than average. During La Niña years, the discharges at Óbidos are more than 7,000 m^3 s^{-1} above the normal.

River discharge peaks earlier in southern and eastern Amazonia (Tocantins and Madeira Rivers) than in northern Amazonia (Rio Negro) (Figure 9.2). Discharges of the Amazon River measured at Óbidos (200 km inland from the mouth of the Amazon) do not represent the true amount of water that reaches the mouth of the Amazon, since they do not include the waters of the Xingú and Tocantins Rivers (Marengo *et al.*, 1994). Mean discharge at the Óbidos gauging station is 175,000 m^3 s^{-1} (or 2.5 mm day^{-1}), while the correct value at the mouth of the Amazon (Roads *et al.*, 2002; Marengo, 2004b) is 210,000 m^3 s^{-1} (or 2.9 mm day^{-1}).

9.2.2 Atmospheric and hydrological water balance

The hydrological cycle of the Amazon region is of great importance since the region plays an important role in the functioning of regional and global climate. Variations in its regional water and energy balances at year-to-year and longer timescales are of

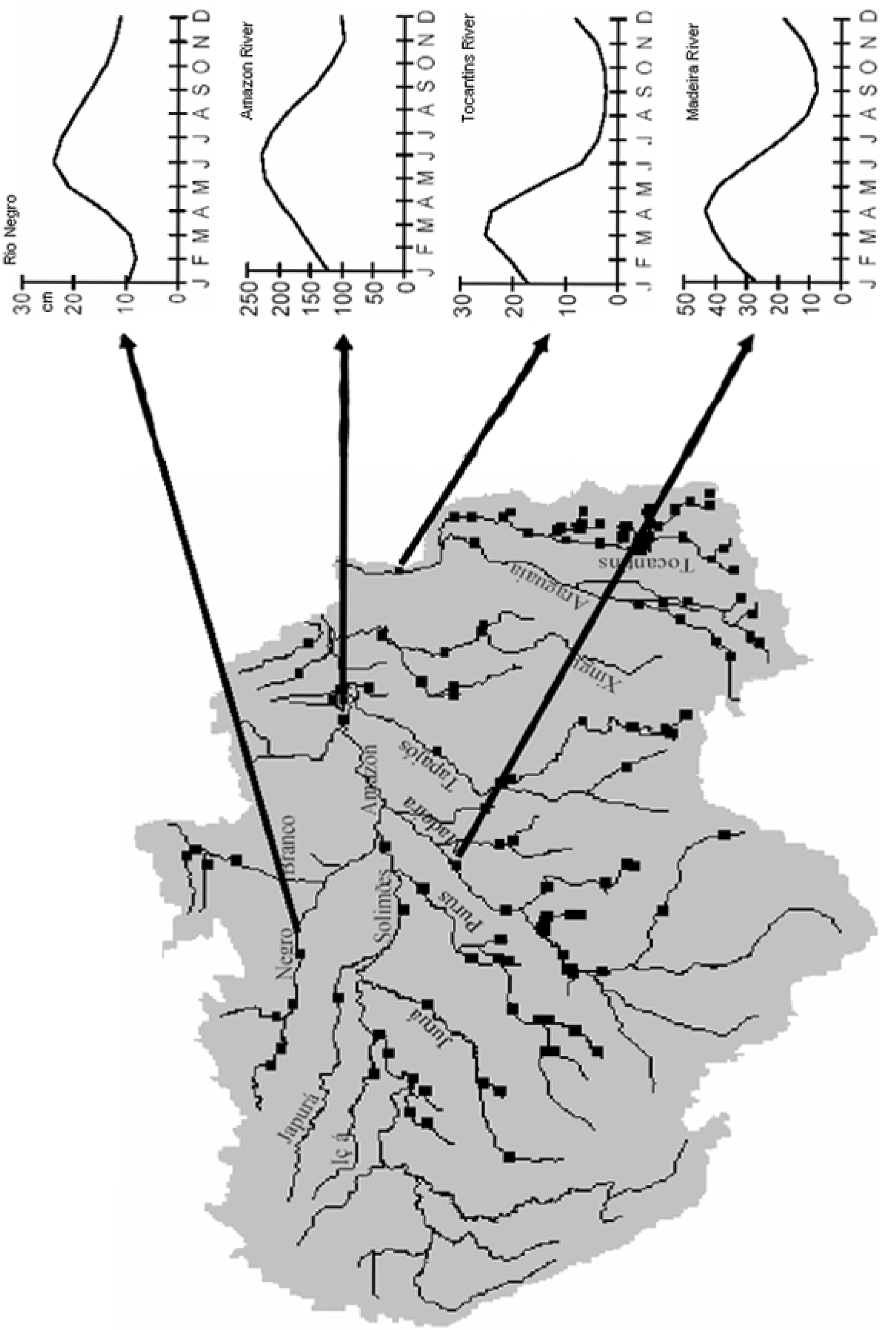

Figure 9.3. Seasonal cycle of river discharges and levels in northern and southern Amazonia. Levels are in cm (Rio Negro at Manaus). Discharges from the Amazon, Madeiras, and Tocantins Rivers are in $10^3\ m^3\ s^{-1}$. *Source*: M. Costa and M. Coe.

special interest, since alterations in circulation and precipitation can translate ultimately to changes in the streamflow of the Amazon River. In addition, these changes can also affect atmospheric moisture transport from the Amazon region to adjacent regions. Since the late 1970s, large-scale water budget studies have been conducted for this region using a variety of observational data sets varying from radiosondes to global reanalyses (Salati, 1987; Matsuyama, 1992; Eltahir and Bras, 1994; Marengo *et al.*, 1994; Rao *et al.*, 1996; Vörösmarty *et al.*, 1996; Costa and Foley, 1999; Curtis and Hastenrath, 1999; Zeng, 1999; Labraga *et al.*, 2000; Roads *et al.*, 2002; Marengo, 2004b and references quoted therein). Most of these reanalyses discuss the impacts of remote forcing on variability of the components of water balance, as well as the role of evapotranspiration in water balance.

The lack of continuous precipitation and evaporation measurements across the entire basin and of measurements of river discharge along the Amazon River and its main tributaries has forced many scientists to use indirect methods for determining the water balance for the region. Early studies by Salati and Marques (1984) attempted to quantify the components of the water balance by combining observations from the few radiosonde stations in Amazonia and models to estimate evapotranspiration. The Amazon Rivers drains an area of approximately 5.8×10^6 km^2, with an average discharge of 5.5×10^{12} m^3 yr^{-1}. However, different estimates of the area of the Amazon Basin by different authors have led to a wide range of computed discharges of the Amazon River during the last 25 years. Most of these estimates are based on the records of the Amazon at Óbidos, and are shown in Table 9.1. The differences are due to the different areas considered, and more recently (Marengo, 2004b) the discharges at Óbidos (available since the mid-1970s) were corrected by the Brazilian National Water Agency ANEEL so that they would be more representative of the discharge at the delta of the Amazon River.

Results of previous studies of the annual water budget in Amazonia are listed in Table 9.2 (compiled from Matsuyama, 1992; Marengo *et al.*, 1994; Costa and Foley, 2000; Marengo, 2004b). Main differences in results are due to the different areas considered for the basin that translate to different discharge and derived run-off, different precipitation networks and methods of assessment (mostly based on gridded rainfall data or from rain gauges distributed irregularly in the basin), and the methods used to determine annual water balance, where evapotranspiration ET is estimated as residual precipitation P and discharge R. However, the equation $P = ET + R$ does not guarantee accurate estimates of the possible role of the tropical forest in recycling moisture for rainfall.

The annual cycle of water balance terms shows some differences between northern and southern sections of the basin (Figure 9.3). There is seasonality in R and P: with R peaking between 3 and 4 months after P. The E/P ratio of the dry season is larger than that of the rainy season, indicating that the role of evaporation (and evapotranspiration) on the water cycle is relatively more important in the dry season than in the rainy season. The largest E/P is found during the dry season in the southern region, reaching values greater than 1, which is larger than respective values in the northern region (Marengo, 2004b). In general, in the present climate the Amazon Basin can be considered as a moisture sink ($P > E$).

Table 9.1. Observed river discharge for the Amazon River at Óbidos (Matsuyama, 1992; Marengo *et al.*, 1994; Marengo and Nobre, 2001; Marengo, 2004b).

Study	Amazon River discharge (10^3 m^3 s^{-1})
Leopold (1962)	113.2
UNESCO (1971)	150.9
Nace (1972)	175.0
UNESCO (1974)	173.0
Baumgartner and Reichel (1975)	157.0
Villa Nova *et al.* (1976)	157.0
Milliman and Meade (1983)	199.7
Nishizawa and Tanaka (1983)	160.0
Oki *et al.* (1995)	155.1
Matsuyama (1992)	155.1
Russell and Miller (1990)	200.0
Vörösmarty *et al.* (1989)	170.0
Sausen *et al.* (1994)	200.0
Marengo *et al.* (1994)	202.0
Costa and Foley (1998a)	162.0
Zeng (1999)	205.0
Leopoldo (2000)*	160.0
Leopoldo (2000)**	200.0
Roads *et al.* (2002)	224.0
Marengo (2004b)*	175.0
Marengo (2004b)**	210.0

* Measured at Óbidos.
** Measured (corrected) at the mouth.

Estimates of the water balance in the Amazon region exhibit some uncertainities, derived mainly from the use of different rainfall data sets (either gridded or station data), use of streamflow data at the Óbidos gauge site (corrected or uncorrected), and use of global reanalyses produced by some meteorological centers in the U.S. and Europe. The National Centers for Environmental Prediction (NCEP) and the European Center for Medium Range Weather Forecast (ECMWF) have carried out retrospective analyses (reanalyses) over the last decade using a single model and data assimilation to represent climate evolution from as early as World War II. These reanalyses can highlight characteristic features of circulation and water balance. However, while data assimilation should in principle provide for a description of the water flux field, there are no guarantees that this description will be superior to that obtained from objective analysis and radiosonde observations alone, especially over continental regions. There is a need for the level of uncertainty to be identified in the measurement or estimation of the components of the water budget.

Table 9.3 shows the annual values of the water budget components for Amazonia in its entirety, giving the mean and two extremes of interannual variability: El Niño

Table 9.2. The annual water budget of the Amazon Basin. P = Precipitation; ET = Evapotranspiration; R = Streamflow—all in mm y^{-1}. In this table the water balance equation $ET = P - R$ is used, P and R are measured, and ET is obtained as a residual. Marengo (2004b) used the water balance equation that considers the non-closure of the Amazon Basin (Marengo and Nobre, 2001; Marengo, 2004b).

Study	P	ET	R
Baumgartner and Reichel (1975)	2,170	1,185	985
Villa Nova *et al.* (1976)	2,000	1,080	920
Jordan and Heuveldop (1981)	3,664	1,905	1,759
Leopoldo *et al.* (1982)	2,076	1,676	400
Franken and Leopoldo (1984)	2,510	1,641	869
Vörösmarty *et al.* (1989)	2,260	1,250	1,010
Russell and Miller (1990)	2,010	1,620	380
Nishizawa and Koike (1992)	2,300	1,451	849
Matsuyama (1992)	2,153	1,139	849
Marengo *et al.* (1994)	2,888	1,616	1,272
Costa and Foley (2000)	2,166	1,366	1,800
Marengo (2004b)	2,117	1,570	1,050

1982–83, and El Niño 1997–98 and La Niña 1988–89. Reduced precipitation (P), run-off (R), and moisture convergence (C) are found during these two strong El Niño events while values larger than normal are found during La Niña 1988/89, and in all cases $P > E$, suggesting that the Amazon region is an atmospheric moisture sink. The difference between these two El Niño events in Amazonia is that during 1997/98 large-scale circulation anomalies over the Atlantic sector did not allow for much convergence of moisture. In the long term $P > E$, during La Niña events it is shown that $P > E$, and during El Niño 1982–83 and 1997/98 $P > E$, even though the difference is smaller than the mean and La Niña years.

Table 9.3. Climatological water budget 1970–99 for the Amazon Basin. Comparisons are made for the 1982–83 El Niño and the 1988–89 La Niña. P is derived from observations (Marengo, 2004b), E and C are derived from NCEP/NCAR reanalyses, and R is run-off from the historical discharge records of the Amazon River at Óbidos. Units are in mm day^{-1}. $+C$ = Moisture convergence (Marengo, 2004b).

Component	Mean	El Niño 1982–83	El Niño 1997–98	La Niña 1988–89
P	5.8	4.6	5.2	6.7
E	4.3	4.5	4.1	4.4
R	2.9	2.1	2.5	2.9
C	1.4	1.3	1.3	3.1
$P - E$	+1.5	+0.4	+0.9	+2.3
$P - E - C$	+0.1	−0.9	−0.1	−0.8
Imbalance = $[((C/R) - 1)]$	51%	38%	52%	6%

Table 9.4. Water budget 1970–99 of the entire Amazon Basin. P is derived from several data sources: Global Historical Climatology Network (GHCN), Xie and Arkin (CMAP), GPCP, NCEP, Legates–Wilmott (LW), Climate Research Unit (CRU) and from observations by Marengo (2004a). E and C are derived from NCEP/NCAR reanalyses, and R = Corrected run-off from the historical discharge records of the Amazon River at Óbidos. Units are in mm day^{-1}. $+C$ = Moisture convergence (Marengo, 2004b).

Component	GHCN	CMAP	GPCP	NCEP	LW	CRU	Marengo (2004b)
P	8.6	5.6	5.2	6.4	5.9	6.0	5.8
E	4.3	4.3	4.3	4.3	4.3	4.3	4.3
R	2.9	2.9	2.9	2.9	2.9	2.9	2.9
C	1.4	1.4	1.4	1.4	1.4	1.4	1.4
$P - E$	4.3	1.3	0.9	2.1	1.6	1.6	1.5
$P - E - C$	+2.9	−0.1	−0.5	+0.7	+0.2	+0.3	+0.1

The rain gauge based rainfall estimates used by Marengo (2004b) produced a mean of 5.8 mm day^{-1}, which is close to the values obtained in similar studies using different rainfall gridded data sets (CMAP, CRU, GHCN, and GPCP). The observed R at the mouth of the Amazon has been estimated as 2.9 mm day^{-1} (or 210,000 m^3 s^{-1} for a basin area of 6.1 million square kilometers), and this represents the combination of Amazon discharges at Óbidos and those of the Xingú and Tocantins Rivers. Table 9.4 shows that, depending on the rainfall observational data set used, the results for water balance in the region can vary and the $P - E$ difference can get as high as 4.3 mm day^{-1} (GHCN) or as low as 0.9 mm day^{-1} (GPCP).

9.2.3 Maintenance of humidity and import/export of moisture in the Amazon Basin

Under the present climate the Amazon Basin behaves as a moisture sink ($P > E$), and therefore the basin receives moisture from sources such as the tropical rainforest by means of intense recycling and by transport from the tropical Atlantic by near-surface easterly flows or trade winds. The former has generated plenty of concern due to the possible impact of deforestation on the hydrological cycle of the basin (see Section 9.2.5). In this context, water that evaporates from the land surface is lost to the system if it is advected out of the prescribed region by atmospheric motion, but recycled in the system if it falls again as precipitation (Brubaker *et al.*, 1993). Studies on recycling of water in the hydrological cycle of the Amazon Basin have been performed since the mid-1970s by Molion (1975), Lettau *et al.* (1979), Salati *et al.* (1979), Salati and Voce (1984), Salati (1987), and Eltahir and Bras (1993) among others. All indicate the active role of evapotranspiration from the tropical forest in the regional hydrological cycle. During highly active precipitation episodes, moisture convergence can account for 70 to 80% of precipitation. However, on the monthly or longer term, mean evapotranspiration is responsible for approximately 50% of the precipitation, especially in the southeastern parts of the basin.

Bosilovich *et al.* (2002) adapted the passive tracer methodology developed by Koster *et al.* (1986) to the NASA GEOS climate model in order to identify the sources of precipitation in various continental-scale basins, among them the Amazon Basin. They found that the largest contribution to rainfall in the Amazon Basin comes from the South American continent (45.5%) and the tropical Atlantic contributes 37%. The continental sources for Amazon precipitation are large throughout the year, but the oceanic sources vary with the seasonal change of the easterly flow over the tropical Atlantic. In the Amazon, significant amounts of water are not transported from very long distances, contrasting with the Mackenzie River in Canada that receives a significant contribution from Asian sources, or the North American sources for rainfall in the Baltic Sea region in Europe.

In the context of regional circulation in South America, the role of moisture transport from the tropical North Atlantic to the Amazon region has been documented in previous studies (see reviews in Hastenrath, 2001), and the interannual variability of rainfall anomalies in the region has been linked to variability in moisture transport and the intensity of the trade winds in the tropical Atlantic sector. Furthermore, the Amazon Basin is a region that provides moisture to regions in subtropical South America—such as southern Brazil and the La Plata River Basin—as has been shown in various studies.

A relevant feature of South American low-level circulation during the wet warm season is a poleward warm and moist air stream immediately to the east of the Andes often referred to as a low-level jet, because of its resemblance to the U.S. Great Plains Low-Level Jet east of the Rocky Mountains. This moist air current is referred as the "South American Low-Level Jet east of the Andes" or SALLJ—a component of the seasonal low-level circulation in the region that is detected all year long but mostly during the warm season (Berbery and Barros, 2002; Marengo *et al.*, 2004a). Figure 9.4 shows a conceptual model of the SALLJ. It illustrates moisture transport reaching the Amazon Basin by means of tropical North Atlantic easterly trade winds. The moisture transport typical of austral summer time is enriched by evapotranspiration from the Amazon Basin. Once the trade winds reach the Andes they are deflected by the mountains, changing the near-surface flow from northeast to southeast. During winter, subtropical Atlantic highs move towards southern Brazil and northern Argentina, and the winds from the northwest—at the western flank of this anticyclone—seem to replace the northwestern flow from Amazonia typical of summer. This winter flow carries less moisture than the summer flow even though it can sometimes be stronger. In both seasons, this northwest stream at the exit region of the jet converges with air masses from the south, which can favor the development of convective activity and rain at the exit region of the jet in southeastern South America over the La Plata River Basin. The conceptual model also shows the effects of topography in the SALLJ through dry and moist processes, the impact of the energy balance terms (sensible and latent heat) released from the Bolivian Plateau, while the near-surface heat low is important in terms of the impacts of transients (cold fronts, cyclogenesis, etc.) on the SALLJ (see reviews in Nogues-Paegle *et al.*, 2002; Marengo *et al.*, 2004a).

SALLJ variability in time and space is relatively poorly understood because of the limited upper-air observational network in South America east of the Andes, which

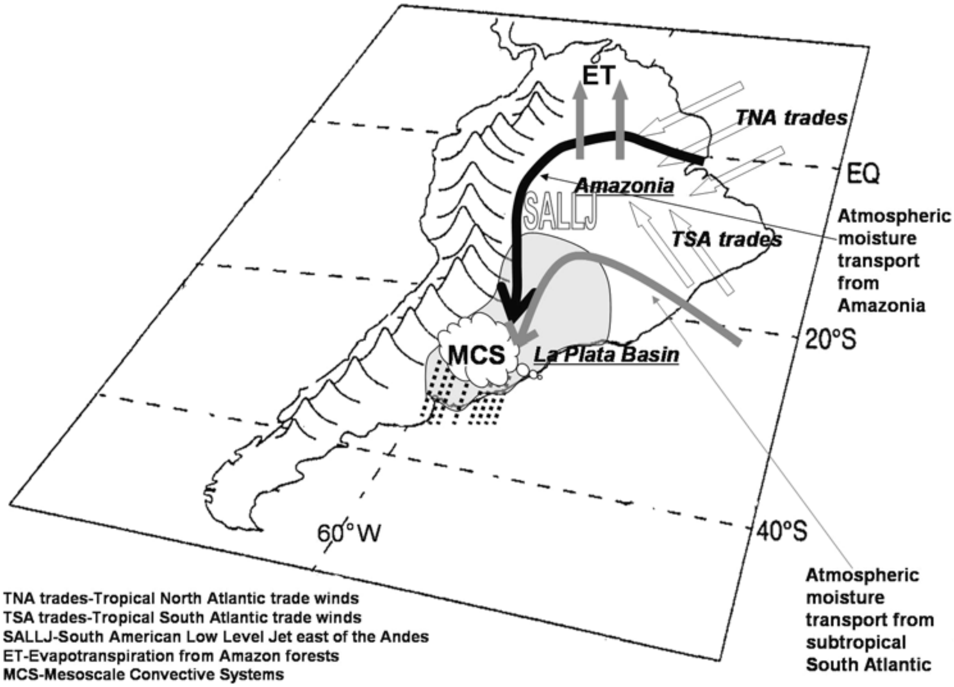

Figure 9.4. Conceptual model of the South American Low Level Jet (SALLJ) east of the Andes. *Source*: Marengo *et al.* (2004a).

seems to be unsuitable to capture the occurrence of the low-level jet, its horizontal extension and intensity, or temporal variability. Regarding time variability, SALLJ events seems to occur all year long, being more intense in terms of wind speed and moisture transport during the austral summer. As shown in Figure 9.4, more frequent wintertime SALLJs are related to the intensity and position of the subtropical South Atlantic anticyclone, and the source of moisture is the tropical–subtropical South Atlantic. During the summertime, the most important source of moisture is the tropical Atlantic–Amazon Basin system, when northeast winds coming from the tropical North Atlantic are deflected to the southeast by the Andes and in doing so get enriched by moisture from the Amazon Basin. This SALLJ, which brings tropical moisture from the Amazon to southern Brazil and northern Argentina, is more frequent in the warm season.

From the moisture budget calculations of Saulo *et al.* (2000), using regional models, a net convergence of moisture flux is found over an area that includes the La Plata Basin, with a maximum southward flux through the northern boundary at low levels that represents the moisture coming from Amazonia via the SALLJ. While there is evidence to suggest that this model provides a realistic description of the local circulation, it is emphasized that observational data are needed to gain further understanding of the behavior of the South American Low-Level Jet and its role in the regional climate.

Another feature of the low-level circulation in South America is the semi-permanent South Atlantic Convergence Zone (SACZ). The SACZ is influenced by SST anomalies over the southwestern tropical Atlantic, has a strong impact on the rainfall regime over southern northeast Brazil, southeast and southern Brazil, and contributes to modulate underlying SSTs over the southwest tropical Atlantic (Chaves and Nobre, 2004). There is evidence that the phases and location of the SACZ respond to Rossby wave activity (Liebmann *et al.*, 1999) and to Madden Julian Oscillation (MJO) (Carvalho *et al.*, 2004). On the other hand, its intensity depends on moisture coming from the Amazon region during summertime. Analyses performed by Nogues-Paegle (2002), Herdies *et al.* (2002), and Marengo *et al.* (2004a) suggest that there is an out-of-phase relationship between the SALLJ and SACZ. Moisture transport and possibly rainfall downstream of the jet or in the SACZ show a contrasting pattern, with enhanced convection and rainfall due to enhanced SALLJ consistent with periods of weak SACZ and *vice versa*. The SALLJ and SACZ are components of the South American Monsoon System (SAMS).

9.2.4 Interannual variability: El Niño and tropical Atlantic impacts

Rainfall variability in various timescales in Amazonia has been the subject of several studies regarding physical causes that could include remote and local forcings. At seasonal and interannual timescales, remotely forced seasonal variations are usually linked to SST anomalies in the tropical Pacific and Atlantic Oceans. The Southern Oscillation (SO) and its extremes—linked to anomalies in the tropical Pacific (El Niño or La Niña at interannual scales), and to sea surface temperature (SST) anomalies and meridional contrasts in the tropical Atlantic—have been associated with rainfall anomalies in the Amazon Basin. Various papers have been devoted to studies on the impact of regional and global SST anomalies in the tropical Pacific and Atlantic Oceans on rainfall anomalies in the region (Ropelewski and Halpert, 1987, 1989; Aceituno, 1988; Richey *et al.*, 1989; Rao and Hada, 1990; Marengo, 1992, 2004a; Meggers, 1994; Nobre and Shukla, 1996; Rao *et al.*, 1996; Guyot *et al.*, 1997; Marengo *et al.*, 1998a, b; Uvo *et al.*, 1998; Fu *et al.*, 1999, 2001; Botta *et al.*, 2002; Foley *et al.*, 2002; Ronchail *et al.*, 2002).

The low SO phase, which is associated with the El Niño phenomenon, is related to negative rainfall anomalies in northern and central Amazonia and anomalously low river levels in the Amazon River, while the high SO phase (related to the La Niña phenomenon) features anomalously wet seasons in northern and central Amazonia. Below-average southern summer rainfall throughout the Amazon Basin during seasons with weaker northeast trades, due to reduced moisture flux from Amazonia, has been identified during extreme El Niño years. In fact, a tendency towards drier rainy seasons and lower Rio Negro levels was detected during El Niño events in 1925–26, 1982–83, and more recently during 1997–98, while wetter conditions were observed during La Niña years in 1988–89 and 1995–96 (Figure 9.5).

The drought of 1998 in north and central Amazonia is generally considered as the most intense of the last 118 years. Kirchoff and Escada (1998) described the ‘‘wildfire of the century’’ in 1998 as one of the most tragic that have ever occurred in Brazil.

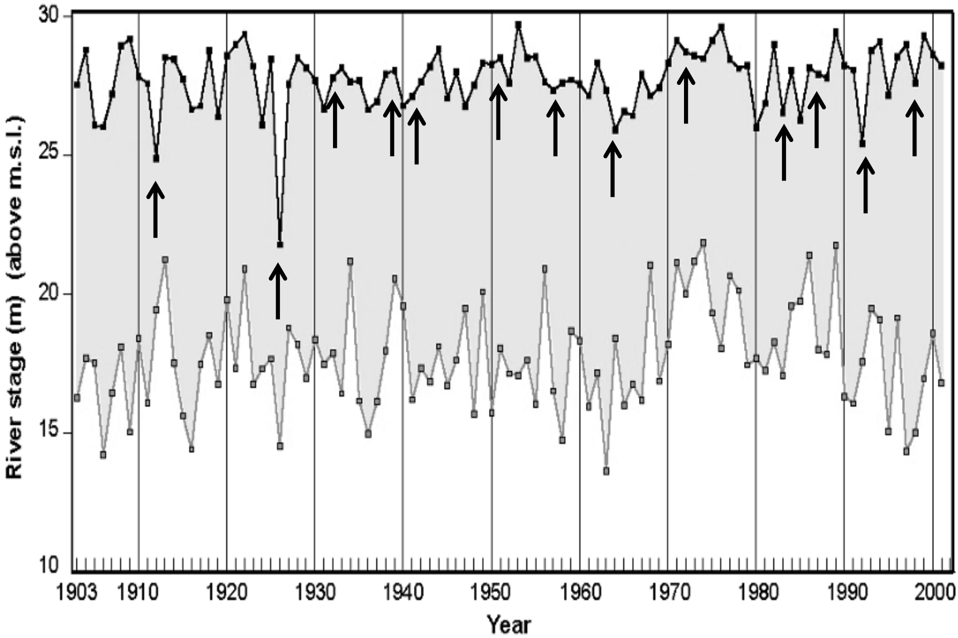

Figure 9.5. Time series of annual maximum (*top*) and annual minimum (*bottom*) river stage (in cm) of the Rio Negro at Manaus, Amazonas. *Source*: Williams *et al.* (2005). Arrows indicate occurrence of the El Niño phenomenon. The 1903–2004 long-term mean stage is 23.22 cm.

However, Williams *et al.* (2005) have suggested that the most severe drought in tropical South America during the 20th century occurred in 1926 during the El Niño of 1925–26. They established that dryness in the northern portion of the Rio Negro basin in 1925 also contributed to the major drought in 1926, through both depletion of soil moisture and possibly a negative feedback on rainfall from the abundant smoke aerosol (see below for elaboration). Annual rainfall deficits are broadly consistent with the reduction in annual discharge for 1926—estimated as 30–40%. The reduction in peak discharge during 1926 is closer to 50%. Sternberg (1987) describes an unparalleled drop in the high-water levels of the Rio Negro at Manaus during the El Niño event in 1926, during a severe dry season in which a great fire blazed for over a month, scorching the vegetation along the main channel. The drought also affected the Orinoco Basin with widespread and drought-related fires in the savannas. Evidence for a dry year in 1912 is also apparent in the Amazon discharge record in Figure 9.5, as is the minimum discharge prior to 1926, 1983, and 1998.

Regional forcing is associated with the systematic buildup of planetary boundary layer moisture and can affect the onset of the rainy season (Fu *et al.*, 1999; Marengo *et al.*, 2001). The atmosphere in southern Amazonia is quite stable; therefore, a great amount of surface heating is required to force the transition between dry and wet seasons, with soil moisture seeming to play a role in the predictability of the rainy season in southern Amazonia (Koster *et al.*, 2000; Goddard *et al.*, 2003; Marengo *et* *al.*, 2003). Near the equator, stability during the dry season is weaker and the transition from the dry to the wet season is more dependent on adjacent SST anomalies. Recent studies have demonstrated that the transition between the dry and wet seasons in southern Amazonia is also dependent on the presence of biogenic aerosol and aerosol produced by biomass burning in the region, which play a direct role in surface and tropospheric energy budget due to their capacity to scatter and absorb solar radiation (Artaxo *et al.*, 1990). This aerosol can also influence atmospheric thermodynamic stability in as much as it tends to cool the surface (by scattering radiation that would otherwise be absorbed at the surface) and by warming atmospheric layers above by absorption. Recent modeling and observational results have indicated that the aerosol plume produced by biomass burning at the end of the dry season is transported to the south and may interact with frontal systems, thus indicating a possible feedback to the precipitation regime by affecting the physics of rainfall formation (Freitas *et al.*, 2004) through the radiative forcing of cloud microphysical processes (Silva Dias *et al.*, 2002).

9.2.5 Decadal- and longer timescale climate and hydrology variability

On longer timescales, studies have documented long-term variations in rainfall in the basin, associated with possible trends or cycles in both rainfall and river levels (Rocha *et al.*, 1989; Chu *et al.*, 1994; Dias de Paiva and Clarke, 1995; Marengo *et al.*, 1998a; Chen *et al.*, 2001; Zhou and Lau, 2001), as well as climatic tendencies in water balance and moisture transport in the basin (Costa and Foley, 1998; Curtis and Hastenrath, 1999; Marengo, 2004b). Botta *et al.* (2002) and Foley *et al.*, (2002) found a 3–4-year peak in Amazonian rainfall which was related to the typical mode of variability of El Niño events (2–7 years). They also found a 24–28-year oscillation which has been discussed previously by Marengo (2004b) and Zhou and Lau (2001), and is indicative of decadal-mode variability. Coe *et al.* (2002) also identify an $\sim$28-year mode variability in the Amazon Basin that drives spatial and temporal variability in river discharge and flooded areas throughout the Amazon/Tocantins River. Previous studies by Marengo (1995) and Callede *et al.* (2004) have also identified positive trends in water levels of the Rio Negro in Manaus and the reconstructed series of discharge for the Amazon River at Óbidos, respectively. The latter author measured increases of 9% in mean annual discharges and 10% in floods between 1903 and 1999.

Recently, Marengo (2004a) performed an analysis of decadal and long-term patterns of rainfall, and—using a combination of rain gauge and gridded rainfall data sets for 1929–98—identified slightly negative rainfall trends for Amazonia in its entirety. However, the most important findings were the presence of decadal timescale variations in rainfall in Amazonia, with periods of relatively drier and wetter conditions, contrasting between northern and southern Amazonia. Northern Amazonia exhibits a weak non-significant negative trend, while southern Amazonia exhibits a positive trend statistically significant at the 5% level (Figure 9.6).

Shifts in the rainfall regime in both sections of the Amazon Basin were identified in the mid-1940s and 1970s (Figure 9.6). After 1975–76, northern Amazonia exhibited less rainfall than before 1975. Changes in the circulation and oceanic fields after 1975

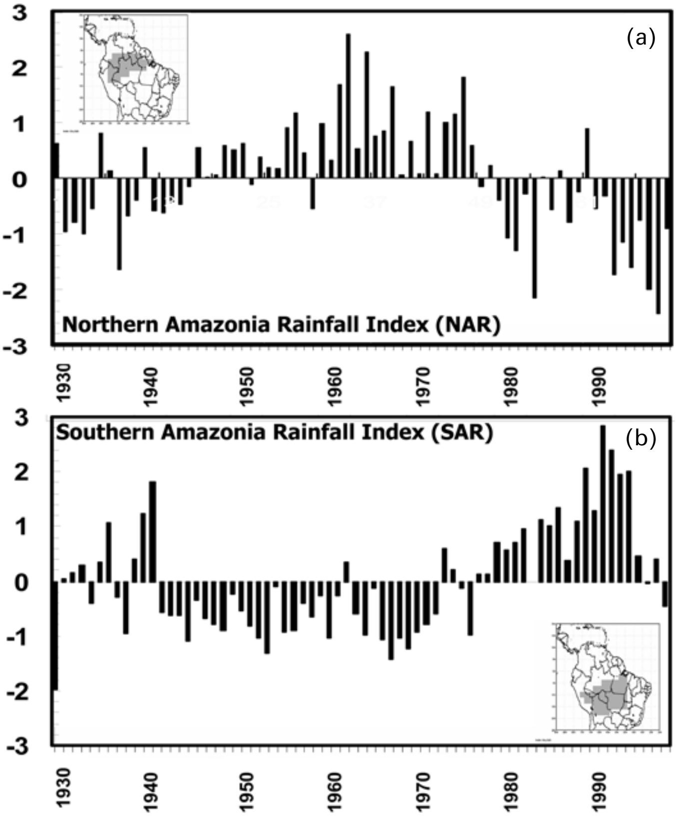

Figure 9.6. Time series of normalized departures of rainfall in northern Amazonia (NAR index) and southern Amazonia (SAR index). The domains of the regions appear in the map inside each panel. *Source*: Marengo (2004a).

suggest the important role played by warming of the tropical central and eastern Pacific on decreasing rainfall in northern Amazonia, due to more frequent/intense strong El Niño events during the relatively dry period 1975–98. This decadal-scale variability has also been detected in other regions in South America, and it seems that other regions—such as the La Plata Basin—also experience these shifts. They have been linked to phases of the Pacific Decadal Oscillation (PDO) (Zhang *et al.*, 1997). The positive PDO phase started in the mid-1970s and apparently ended in the early 2000s. A negative PDO phase started in the mid-1940s and extended until 1975–76, a period with more frequent and intense El Niño events.

Callede *et al.* (2004) suggested a reduction in the interannual variability of mean discharge during 1927–70 that is consistent with the positive phase of the PDO, and positive/negative rainfall departures in northern/southern Amazonia as shown by Marengo (2004a). Their analysis of mean annual discharge and average annual rainfall during 1945–98 demonstrates an increase in flow in relation to rainfall, which could be, according to the authors, the consequence of Amazonian deforestation.

9.2.6 Changes in land use and impacts on Amazon climate

A variety of human activities can act to modify various aspects of climate and surface hydrologic systems. Historically, land-surface changes in Amazonia got intensified in the mid- and early 1970s, when strategic governmental plans—such as Brazil's *Programa de Integração Nacional*—first attempted to promote the economic development of the region. Those plans included the construction of extensive roads throughout the basin and the implementation of fiscal incentives for new settlers, triggering a massive migration of landless people into the region. Changes in land cover can significantly affect surface water and energy balance through changes in net radiation, evapotranspiration, and run-off. However, because of the intricate relationships between the atmosphere, terrestrial ecosystems, and surface hydrological systems, it is still difficult to gauge the importance of human activities in the Amazonian hydrologic cycle. As indicated in Section 9.2.2, aerosol and smoke from biomass burning during the dry season in Amazonia seems to have an impact on the onset of the rainy season in southern Amazonia, and ultimately increase in the concentration of greenhouse gases and aerosol could affect the energy balance and thus climate of the region. Recent data from remote sensing show that large areas of Amazonia (mostly Brazilian Amazonia) have been changed from forest to pasture and agricultural land and that observed deforestation rates in the Brazilian Amazon increased in 2004 relative to 2003. Deforestation rates stabilized somewhat in the early 1990s, mainly in Brazilian Amazonia, but the underlying pressures to continue land-use change are still present: a growing population in the developing nations of Amazonia and plans for a road network criss-crossing the region.

Costa *et al.* (2003) have identified increases in the annual mean and high-flow season discharge of the Tocantins River in southeastern Amazonia since the late 1970s, even though rainfall has not increased. They suggest that changes in the land cover in the basin for agricultural purposes and urban development have altered the hydrological cycle of the basin. Callede *et al.* (2004) suggest that increases in the mean annual discharge of the reconstructed series of the Amazon River at Óbidos during 1945–98 could be the consequence of Amazon deforestation.

The construction of reservoirs for hydroelectric generation in Amazonia has an impact on the hydrological regime as well as on biodiversity and water quality (Tundisi *et al.*, 2002), depending on the size of the inundated area in the tropical rainforest. Brazil has five reservoirs dedicated to hydroelectricity generation (Coaracy Nunes, Curua-Una, Tucurui, Balbina, and Samuel) and a further six are planned to be built (Manso, Cachoeira, Ji-Parana, Karanaô, Barra do Peixe, and Couto Magalhães). Studies on Samuel, Balbina, and Tucurui show that there are different degrees of biomass degradation, and that it is more advanced in Tucurui because it is older. Land-use changes have also been reported near the site of the reservoir due to human settlements in the region.

In an attempt to investigate the possible impact of Amazon deforestation on regional climate and hydrology, global climate model simulations of land-use changes—where forest is replaced by grassland throughout the whole basin—have suggested a possible change in regional and global climate as a result of tropical

Table 9.5. Comparison of climate simulation experiments of Amazon deforestation from global climate models. Results show the differences between deforested and control runs. E = Change in evapotranspiration (mm day^{-1}); T = Change in surface air temperature (°K); P = Change in precipitation (mm day^{-1}); R = Run-off, calculated as the difference between P and E ($R = P - E$). Modified from Marengo and Nobre (2001).

Experiment	E	T	P	R
Dickinson and Henderson-Sellers (1988)	−0.5	+3.0	0.0	+0.5
Dickinson and Kennedy (1992)	−0.7	+0.6	−1.4	−0.7
Henderson-Sellers *et al.* (1993)	−0.6	+0.5	−1.6	−1.0
Hahmann and Dickinson (1995)	−0.4	+0.8	−0.8	−0.4
Zeng *et al.* (1996)	−2.0	—	−3.1	−1.1
Hahmann and Dickinson (1997)	−0.4	+1.0	−1.0	−0.6
Costa and Foley (2000)	−0.6	+1.4	−0.7	−0.1
Lean and Warrilow (1989)	−0.9	+2.4	−1.4	−0.5
Lean and Warrilow (1991)	−0.6	+2.0	−1.3	−0.7
Lean and Rowntree (1993)	−0.6	+1.9	−0.8	−0.3
Lean and Rowntree (1997)	−0.8	+2.3	−0.3	+0.5
Lean *et al.* (1996)	−0.8	+2.3	−0.4	+0.4
Manzi and Planton (1996)	−0.3	−0.5	−0.4	−0.1
Nobre *et al.* (1991)	−1.4	+2.5	−1.8	−0.4
Shukla *et al.* (1990), Nobre *et al.* (1991)	−1.4	+2.5	−1.8	−0.4
Dirmeyer and Shukla (1994)	−0.4	—	−0.7	−0.3
Sud *et al.* (1990)	−1.2	+2.0	−1.5	−0.3
Sud *et al.* (1996b)	−1.0	+3.0	−0.7	+0.3
Walker *et al.* (1995)	−1.2	—	−1.5	−0.3
Polcher and Laval (1994a)	−2.7	+3.8	+1.0	+3.7
Folcher and Laval (1994b)	−0.4	+0.1	−0.5	−0.1
Zhang *et al.* (2001)	−0.4	+0.3	−1.1	−0.01
Voldoire and Royer (2004)	−0.6	−0.1	−0.4	—

deforestation (see reviews in Salati and Nobre, 1991; Marengo and Nobre, 2001). Under a hypothesized Amazon Basin deforestation scenario, almost all models show a significant reduction in precipitation and evapotranspiration (Table 9.5), and most found a decrease in streamflow and precipitation and increases in air temperature. Deforestation results in increased surface temperature, largely because of decreases in evapotranspiration. The combined effect of deforestation and a doubling of CO_2, including interactions between processes, is projected to increase temperature in the order of +1.4°C.

However, such predictions disagree with results found by mesoscale models, which have been consistently predicting the establishment of enhanced convection—and potential rainfall—above sites of fragmented deforestation. Modeling studies have tried to reproduce that effect, and it was noted that mesoscale circulations between forested and deforested patches may significantly affect the timing and formation of clouds, potentially altering both intensity and distribution of precipitation (Chen and Avissar, 1994). It was estimated that, at the mesoscale, a landscape with a relatively large discontinuity tends to produce more precipitation than a homogenous domain, inducing a negative feedback that ultimately tends to eliminate the discontinuity (Avissar and Liu, 1996). In some cases, the thermal circulation induced may get as intense as that of a sea breeze—such as that over domains with extended areas of unstressed, dense vegetation bordering areas of bare soil. The horizontal scale of such landscape heterogeneities is another factor that may affect the establishment of precipitation (Pielke, 2001), while the optimum scale for triggering convection seems to depend on the air humidity level (Avissar and Schmidt, 1998). A strong enough synoptic (or background) wind-field may also interact with the induced circulation, possibly masking its existence at times (Segal *et al.*, 1988). It was noted that a mild background wind of $5\,m\,s^{-1}$ may be sufficient to virtually remove all thermal impacts generated by land-surface discontinuities (Avissar and Schmidt, 1998), although more recent studies have revealed that a strong background wind may only advect instabilities elsewhere rather than disperse them (Baidya Roy and Avissar, 2003). The results by Baidya Roy and Avissar (2003) and Weaver and Baidya Roy (2002) were derived from high-resolution mesoscale models simulating the effect of land-surface and land-use changes in Rondonia, southern Amazonia, during the LBA-WET AMC field campaign in 1999 (Silva Dias *et al.*, 2002). They found that coherent mesoscale circulations were triggered by surface heterogeneity and that synoptic flow did not eliminate these circulations but advected them away. These circulations affect the transport of moisture, heat the synoptic scale, and can affect climate.

All these projected changes in Amazonia (Table 9.5) may have climatic, ecological, and environmental implications for the region, the continent, and the globe. A sound knowledge of how the natural system functions is thus a prerequisite to defining optimal development strategies. The complex interactions between the soil, vegetation, and climate must be measured and analyzed so that the limiting factors to vegetation growth and soil conservation can be established. New knowledge and improved understanding of the functioning of the Amazonian system as an integrated entity and of its interaction with the Earth system will support development of national and regional policies to prevent exploitation trends from bringing about irreversible changes in the Amazonian ecosystem. Such knowledge, in combination with enhancement of the research capacities and networks between Amazonian countries, will stimulate land managers and decision-makers to devise sustainable, alternative land-use strategies along with forest preservation strategies.

9.3 PROJECTIONS OF FUTURE CHANGES IN CLIMATE AND HYDROLOGY OF THE AMAZON BASIN

9.3.1 Detected changes in air temperature

According to the IPCC *Third Assessment Report* (IPCC, 2001), observed air temperature trends have shown a warming of $0.6 \pm 0.2^{\circ}C$ during the 20th century. Climate reconstructions have shown the 20th century as being the warmest of the last 1,000 years. Specifically, the 1990s was the warmest decade of the millennium, with 1998 as the warmest (an El Niño year), followed by 2002, 2003, and 2004. Warming since 1976 was of the order of 0.19°C/decade—higher than the warming of the period 1910–1945 (0.14°C/decade). The warmest years since 1976 (in decreasing order) were: 1998, 2002, 2003, 2004, 2001, 1995, 1997, 1990, 1999, 1991, and 2000.

For the Amazon region, few studies have identified long-term trends in air temperature. Sansigolo *et al.* (1992) and Marengo (2004b) detected warming in many of the major cities in Brazil, including Manaus and Belém in the Amazon Basin, but this warming was linked to the urbanization and heat island effects in big cities. At a regional level, Victória *et al.* (1998) detected an observed warming of +0.56°C/100 years until 1997, while Marengo (2003) updated this warming to +0.85°C/100 years until 2002. Recently, Vincent *et al.* (2005) have identified some positive trends in extreme air temperatures and negative trends in the diurnal temperature range at quite a few stations in South America. Similarly, positive trends were identified in nighttime and daytime air temperatures at some stations in Amazonia.

9.3.2 Future projections of changes in air temperature

Regarding future climate change projections in the region, different IPCC models for A2 and B2 scenarios were assessed by Marengo and Soares (2003), who detected a warming of different magnitudes as a result of their use. The largest warming was approximately 4–6°C in central Amazonia during austral winter in the A2 scenario and a 2–3°C increase in the B2 scenario of the Hadley Centre model for the year 2100. At the level of time slices, Marengo and Soares (2003) show that the HadCM3 model shows a larger warming in Amazonia after 2080 than those expected in 2020 and 2050—especially during austral spring (September–November)—reaching up to 11°C in the A2 scenario and 8°C for the B2. The other models show similar tendencies with warming a bit lower, with the exception of the CSIRO model that actually shows cooling in Amazonia.

Figure 9.7 shows a time series of mean annual air temperature up to 2100 from six IPCC models for the A2 scenario: HadCM3 from the U.K., the CCCma from Canada, the CSIRO from Australia, R30 from the U.S., and the CCSR/NIES from Japan. All IPCC models exhibit warming up to 2100, with the warmest projections being from the HadCM3 (almost 10°C warmer than the 1961–90 long-term mean) and the CCSR/NIES (7°C warmer than the 1961–90 long-term mean). The other models exhibit warming of the order of 3°C or less. The figure shows consistent warming in all models but with a large spread among models after 2070.

9.3.3 Detected and projected changes in precipitation

The IPCC *Third Assessment Report* (IPPC, 2001) has shown no clear signs of negative trends in Amazonia due to increased deforestation as one would expect (Section 9.2.3). Section 9.2.5 describes trends and long-term variability in observed mean seasonal and annual rainfall and discharges. The magnitude and size of the trends depend on rainfall data sets, length of records, etc. and the uncertainty is high since

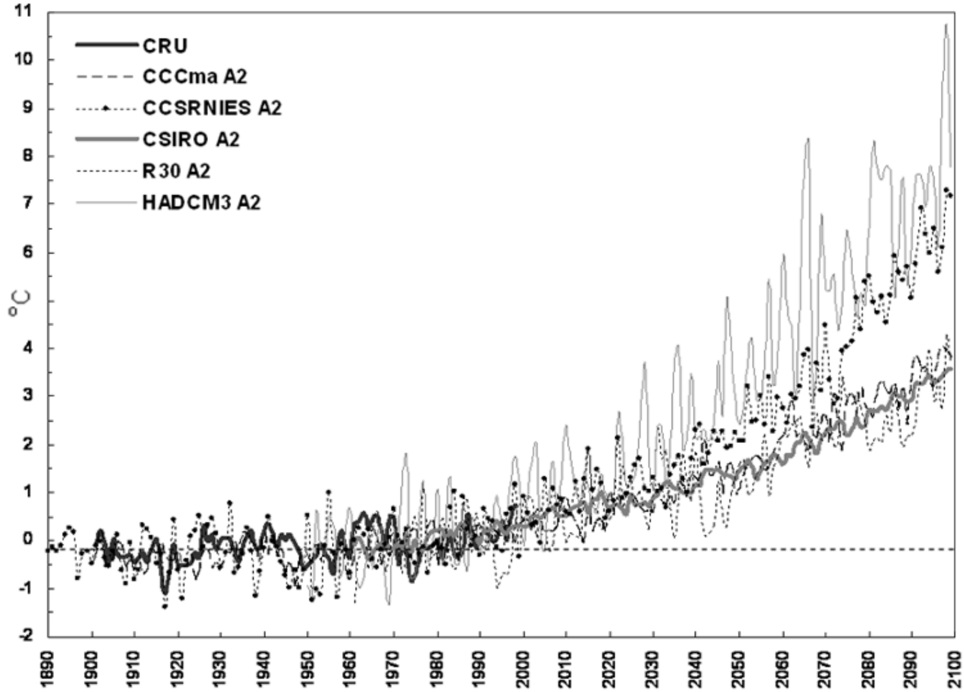

Figure 9.7. Air temperatures in the Amazon region as simulated by five IPCC models: R30 (GFDL, U.S.A.), CCCMa (Canada), HadCM3 (U.K.), CSIRO (Australia), CCSR/NIES (Japan). Anomalies are in °C from the 1961–1990 long-term mean. Observations are from the CRU rainfall data set for the 1903–1998 period. Air temperatures are for the A2 SRES.

various studies (Section 9.2.4) found trends that vary in direction when different length periods are used. What is clear is the presence of decadal-scale rainfall variability, with wetter periods during 1945–76 and relatively drier periods during 1977–2000 in northern Amazonia. Analysis of extreme rainfall events in Amazonia (Haylock *et al*., 2006) has also shown this, with some degree of spatial coherence and uncertainty due to the reduced number of stations used for the region. Using the records at Manaus and Belém, since 1961 an increase in rainfall exceeding the 99th and 95th percentiles has been detected, suggesting a tendency for more intense and extreme events during the last 40 years, even though the annual totals may have not exhibited a significant positive trend.

9.3.4 Future projections of changes in precipitation

For the future, the different IPCC models for the A2 scenario (Hulme and Sheard, 1999; Marengo and Soares, 2003) show negative rainfall anomalies in most of the Amazon Basin (up to 20% reduction), while for the lower emission scenario (B2) this reduction was 5%. In addition to these emission scenarios, there are climate change scenarios for the 21st century: new developments in dynamic vegetation schemes and coupled climate–carbon models (Cox *et al.*, 2000, 2004; Betts *et al.*, 2004) have shown an effect—named ''die-back of the Amazon forest''—by means of which rising atmospheric CO_2 is found to contribute to a 20% rainfall reduction and to more than a 30% increase in surface temperatures in the Amazon Basin, through physiological forcing of stomatal closure. They also show an increase in rainfall in southern Brazil and northern Argentina. These projections for drought in Amazonia after 2040 also show systematic warming in the tropical Pacific indicative of an El Niño-like mode of variability becoming persistent after 2040. However, the likelihood of this extended El Niño or more frequent/intense El Niño mode scenarios in a global warming world is still an open issue. Furthermore, these studies do not give any information on the likelihood of change in extreme rainfall events.

If we consider the climate change scenarios discussed by Marengo and Soares (2003) and the model results from Cox *et al.* (2004) and Betts *et al.* (2004), increase in the concentration of greenhouse gases in the atmosphere may result in increased warming and rainfall reduction rates in Amazonia, implying more intense respiration and the closing of stomata, leading to the decline of Amazon tropical forests. These projections also suggest that vegetation will be more susceptible to fires due to the drying of Amazonia. Figure 9.6 shows the observed increase in rainfall in southern Amazonia contrasting with the slight decrease of rainfall in northern Amazonia. Systematic increases in rainfall in southern Amazonia and southeastern South America, as detected by Obregon and Nobre (2003), are consistent with a tendency for more intense/frequent transport of moisture from Amazonia to the La Plata Basin by the South American Low-Level Jet (SALLJ) east of the Andes (Marengo *et al.*, 2004c). Thus, it could be hypothesized that—following Betts' results from the HadCM3 model after the mid-2050s—the drying of the Amazon Basin and humidification of the southern Brazil and northern Argentina region in an extended El Niño mode predicted by this model could be explained by changes in regional circulation, with an increase in SALLJ frequency and/or intensity in a global warming world. However, not much can be said regarding changes in extreme rainfall events.

A reduction in rainfall will also have impacts on river and water levels, as well as on rainfall distribution in them, with longer dry periods and intense rainfall events concentrated on a few days. However, the uncertainty level is still high since these experiences were obtained from a model—in this case, the HadCM3 model—and other models are being tested now to see if they also simulate Amazon die-back. It is clear that models must improve or include the representation of natural processes, such as clouds and the impacts of aerosol.

9.3.5 Changes in the hydrology of the Amazon River Basin

Macroscale hydrological models, which model the land-surface hydrological dynamics of continental-scale river basins, have rapidly developed during the last decade (Russell and Miller, 1990; Marengo *et al.*, 1994; Miller *et al.*, 1994; Nijssen *et al.*, 1997, 2001). These models can act as links between global climate models and water resource systems on large spatial scales and long-term timescales. Predictions of river discharge in the Amazon Basin for present climates and $2 \times CO_2$ future scenarios have been calculated by Russell and Miller (1990) and Nijssen *et al.* (2001) using global models. Some problems in the parameters of the model—or perhaps the availability of suitable run-off data for validation—indicate that in most models rainfall and run-off in Amazonia are underestimated (Marengo *et al.*, 2003). This also generates uncertainty in the projected values of run-off in the future, forced either by increase in GHGs or in changes in land use and land cover. More recent simulations by Coe *et al.* (2002), using a terrestrial ecosystem model, have been successful in simulating interannual and seasonal run-off variability in Amazonia, and—even though the discharge is consistently underestimated—the model captures climate variability and the impacts of El Niño events since the early 1950s.

Land-surface changes are accompanied by alterations in climate, and consequently in the hydrological cycle. Water flux anomalies associated with these changes have already occurred in many parts of the globe and have been detected, for example, over tropical and subtropical basins—such as the Yangtze (Yin and Li, 2001; Yang *et al.*, 2002), Mekong (Goteti and Lettenmaier, 2001), Amazon River Basin (see reviews in Marengo and Nobre, 2001)—as well as on several catchment areas within the African continent (Calder *et al.*, 1995; Hetzel and Gerold, 1998). Recently, major land-surface changes have been observed in various parts of the tropics (Aldhous, 1993), and Amazonia—which holds more than 40% of all remaining tropical rainforests in the world—has been the focus of many studies about the impact of these changes on hydrological dynamics.

Increasing trends in discharge and precipitation were observed at all but the eastern parts of the Amazon Basin between the late 1950s and the early 1980s, and—despite contentions that these trends were associated with upstream deforestation (Gentry and Lopez-Parodi, 1980)—most time series retreated to their long-time means by the end of the 2000 period. These are more indicative of decadal tendencies—rather than of any unidirectional trends—as a response to fluctuations over the Tropical Pacific associated with ENSO events, and not deforestation. These findings thus support the idea that the atmospheric fluctuations induced by remote forcings (Richey *et al.*, 1989; Fu *et al.*, 2001) can potentially offset or overshadow the effects of deforestation (Chen *et al.*, 2001). The existence of trends in additional terms of the hydrological cycle in Amazonia has also been tested, and the swings and tendencies of significant changes on spatial averages for the input and output fluxes of water vapor (decreasing) vary according to the type and length of time series used. The use of spatially aggregated point data may not be appropriate for the detection of trends, due to the inevitable "dilution" of the signal during the upscaling process, while the use of gridded data sets may also create artificial trends. What has also been observed is decadal timescale variability, more so than any unidirectional trend towards systematic drying or moistening of the Amazon region in the long term. However, these tendencies do not provide information on whether significant changes in precipitation extremes would occur in the future.

The deforestation experiments listed in Table 9.6 show that most models simulate a decrease in rainfall and run-off due to the large-scale removal of forests in the Amazon Basin. However, these experiments did not alter at all the concentration of

Table 9.6. Natural and anthropogenic forcing, climatic tendencies, and human and ecosystem dynamics in Amazon countries.

Forcing (natural or anthropogenic)	Impacts on water resources	Consequences
El Niño and tropical Atlantic sea surface temperature anomalies	Changes in rainfall distribution in the Amazon region	Drought in northern Amazon region; problems in transportation due to low river water levels; high risk of forest fires at seasonal level; impacts on natural river ecosystems; impacts on agriculture; impacts on water storage for hydroelectric generation
Climate change due to increase in concentration of GHG	Possible changes in the hydrological cycle; changes in the energy balance and warming; changes in biodiversity and natural ecosystems	Dynamics of vegetation affected; Amazon forests die and become savanna; drying of the Amazon region; floods or extremely low water levels likely to occur; more frequent forest fires; impacts on water storage for hydroelectric generation
Deforestation and land-use change	Possible changes in hydrological and energy cycles; changes in water quality and chemistry due to deforestation in the east flank of the Andes (upper-Amazon countries)	Regional rainfall reduction; regional warming; erosion; sedimentation along the main channel and accumulation of sediments in reservoirs; water quality and biodiversity may be affected
Biomass burning (natural and man-made)	Changes in water and energy cycles; changes in air quality	Impacts on the onset of the rainy season and physics of rainfall; impacts on air quality and sensitivity to warming due to release of large amounts of GHG and aerosol

GHGs or aerosol in the atmosphere. Later experiments on $2 \times CO_2$ scenarios have produced reductions in streamflows of the Amazon River and major rivers in the tropics as compared to present day streamflows, even though present day streamflows were somewhat over- or under-estimated in the past.

Table 9.6 lists natural and anthropogenic forcing that affects climate and hydrology in the Amazon region, the expected impacts on water resources of the region, and the possible consequences on the components of the climate system. The major forcings can be natural or human-induced and their impact could be on various timescales: intraseasonal (onset of the rainy season in the region), interannual (drought or flood associated with El Niño), and long-term (changes in vegetation, soil hydrology, etc.). The table can serve as a starting point in the assessment of possible impacts of future climate change in Amazonia.

9.4 UNCERTAINTIES IN CLIMATE AND HYDROLOGY VARIABILITY, LONG-TERM TRENDS, AND CLIMATE CHANGE IN THE AMAZON REGION

As can be seen from the analysis of observed hydrometeorological fields in previous sections, no significant trends have been detected in the rainfall regime of the Amazon region as a whole, or on the discharge of the Amazon River and its tributaries, that could be attributed to deforestation and human-induced land-use changes, or to increase in the concentration of greenhouse gases. The detected contrasting rainfall trends in northern and southern Amazonia on decadal timescales are somewhat surprising, since it would suggest the dominance of natural climate variability on rainfall variability on both sides of the basin, linked to El Niño or El Niño like modes. Some previous studies have detected upward discharge trends in Peruvian Amazon rivers in the mid-1980s, but these trends were more a consequence of changes in the main channel or in the measurement techniques, and not of rainfall changes. However, rainfall records do not go back more than 50 years, and there are large sections of the basin with no data, so there is considerable uncertainty about climate trends in Amazonia. River data exhibit longer time records, but there is uncertainty in climate trends detected using river series or on short-time rainfall series, since some of the river data could be affected by non-climatic signals, such as the back-water effect in the Rio Negro series in Manaus, Brazil. Gridded rainfall data sets have helped in solving problems on regional coverage, but they may have added even more uncertainties since there are differences among data sets. National hydrometeorological networks are deteriorating, and there are fewer stations in 2000 than there were in the 1970s. Automatic hydrological stations are being installed in some countries, but the operational costs are high and not all Amazon countries have adopted the same operational system, and problems with calibration and data intercomparison have been reported.

The choice of rainfall data sets has an impact on results regarding the water budget. Uncertainties exist in these results and they are sensitive to rainfall network density—when using station data—and to numerical interpolation techniques and use of satellite data to fills gaps in time and space across the basin—when using gridded data. The lack of continuous upper-air observations in the basin makes it difficult to estimate moisture transport into and out of the basin, and we have to rely on model data for this—such as reanalyses. Furthermore, the use of uncorrected discharge under-estimates freshwater discharge, and thus adds to the uncertainty in closing the water budget. Therefore, it can be stated that estimation of water vapor fluxes in the Amazon Basin is another source of uncertainty in water balance calculations. In addition, there are very few observations of evaporation in some isolated parts of the basin. Perhaps there is need for a unique rainfall data set for the entire region, with rainfall information coming from rainfall stations in the Amazonian areas of each country on a continuous basis. Some Amazon countries are implementing networks of automatic weather and hydrological stations in remote areas of the basin to solve the problem of areas devoid of data, but they need calibration and are expensive to implement and maintain. In addition, data have only recently started to be collected.

Finally, some of the studies reviewed for the writing of this chapter suggest that increases in the concentrations of GHG and aerosol in the atmosphere, as well as changes in land cover for agriculture, have already affected the hydrology of the Amazon Basin. Some uncertainties can be attached to these results due to model limitations, and to the lack of continuous and long-term observational series of climate and hydrological variables. Large uncertainties were also identified in anthropogenic aerosol forcing and response, and changes in land use leading to biomass burning and their impacts on rainfall in the basin. Analyses of the few available long-term rainfall and river series suggest an absence of significant, unidirectional trends towards drier or wetter conditions in the basin. However, more evidence of variability at interannual and decadal timescales is apparent, and these observed trends may be due to natural climate variability and observed climate shifts, while no signs of land-use changes have yet appeared in the long-term climate and hydrologic variability of the region.

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