9. THE SOUTH AMERICAN MONSOON SYSTEM

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The main characteristics of the South American monsoon system are reviewed. According to some diagnostics, the wet season in tropical South America begins in early October over the Brazilian highlands and spreads northward. Wet season rain rates are somewhat smaller than those in other continental monsoons. The annual cycle of precipitation is most pronounced in the southern Amazon, where some of the largest seasonal rainfall occurs. Amazonian rainfall extends to the southeast in the South Atlantic convergence zone. Precipitation in this area is out of phase with that farther to the south and is driven synoptically, although intraseasonal variations are evident, including associations with the Madden-Julian oscillation. The El Niño /Southern Oscillation, the largest known forcing of interannual variability, results in decreased precipitation near the Equator and increased precipitation in east-central South America. Within an ENSO cycle, large subseasonal changes also occur as a result of surface feedback. In the Amazon Basin, long-term trends of precipitation are small. Deforestation of the Amazon is likely to reduce rainfall, but its influence on the large-scale circulation is unclear. The effect of increasing CO_2 on precipitation is also unclear. Although improving, the present-day climate in coupled models is still not well-simulated and remains an impediment to improved climate forecasts.

1. Introduction

Although the trade winds from the Atlantic Ocean are strong year-round, we refer to the warm season circulation over South America as a monsoon system since the large seasonal changes observed have all the characteristics of a "canonical" monsoon regime (Zhou and Lau 1998). These changes include a large increase of precipitation over the Amazon Basin, the establishment of an upper-level anticyclone known as the Bolivian High, and the strengthening of the thermally driven "Chaco" low in northwest Argentina and Paraguay (e.g., Zhou and Lau 1998; Nogués-Paegle *et al.* 2002).

The South American monsoon system (SAMS) is part of the monsoon system of the Americas (Nogués-Paegle *et al.* 2002; Mechoso *et al.* 2005; Vera *et al.* 2006a). The full geographical extent of the "system" includes the core monsoon region and its areas of influence.

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2. Mechanistic Studies

There is still some debate on the source of diabatic heating that is associated with the upper troposheric high component of the SAMS (Zhou and Lau 1998). Rao and Erdogan (1989) showed that heating over the northeastern Altiplano in January is as strong as over mid- and eastern Tibet in July. Silva Dias *et al.* (1983), however, using a simple one-mode baroclinic model, suggested the Bolivian high results from the stationary equatorial Rossby response to transient heating in the Amazon, rather than from direct heating on the Altiplano. Other modeling studies, such as those of DeMaria (1985), Silva Dias *et al.* (1987), Kleeman (1989), Gandu and Geisler (1991), Figueroa *et al.* (1995), Lenters and Cook (1997), and Chen *et al.* (1999) drew similar conclusions. Garreaud (2000b), however, argued that no cause-effect relationship can be determined between Altiplano and Amazon convective activity.

Lenters and Cook (1995) simulated a realistic SACZ in a general circulation model (GCM). They also pointed out the importance of transient moisture flux from the Amazon, as related to extratropical cyclones and fronts, in maintaining the model SACZ. Kalnay *et al.* (1986) and Grimm and Silva Dias (1995) showed that atmospheric heating in the South Pacific convergence zone could force stationary waves that affect the SACZ. Kodama (1999) reproduced the conditions favorable to a convergence zone in an aqua-planet general circulation model with a localized off-equatorial heat source. Figueroa *et al.* (1995), on the other hand, using an eta-coordinate model and a heat source intended to mimic Amazon convection, were able to reproduce low-level convergence in the vicinity of the SACZ provided that orography and background wind field were realistic, but found the simulation success was dependent on a diurnally varying heat source. Lenters and Cook (1999) found that position of the SACZ has a strong influence on the position and intensity of the Bolivian high.

Fu *et al.* (1999) argued that a moistening of the planetary boundary layer and lowering of the temperature at its top, thereby reducing convective inhibition energy (CINE), control the conditioning of the large-scale thermodynamics prior to onset. Li and Fu (2004) found that the main increase in convective available potential energy (CAPE) and reduction in CINE occur prior to rainy season onset, although in the tropical atmosphere, CAPE often exists in the absence of deep convection (Williams and Renno 1993; Collini *et al.* 2008). The intensity of convective storms over the Amazon is greater during the pre-monsoon than during the wet season itself (e.g., Peterson and Rutledge 2001).

3. Precipitation over South America

While annual total precipitation in the Amazon Basin is comparable to that in the other monsoon regions of the world (e.g., Xie and Arkin 1998), rain rates during wet periods are lower than in other land areas (Peterson and Rutledge 2001), and are similar to the rates observed over the ocean (Nesbitt and Zipser 2003). To a first approximation, warm season precipitation over South America follows a northwest to southeast path from the boundary with Central America to the southeastern Amazon, technically the Tocantins, Basin (Horel *et* *al.* 1989). Figure 1 shows the average annual total rainfall. In most of SAMS region precipitation peaks in summer (seasons in this paper refer to the southern hemisphere), although near the Equator the peak is a few months later, and north of the Equator the wet season is in winter (e.g., Rao and Hada 1990; Marengo 1992; Grimm 2003). Near the equator in the western Amazon Basin rainfall is plentiful year-round, with a near doubling from the driest (November) to the wettest (May) month (Fig. 2). The largest contrasts between summer and winter rainfall are in the central Amazon basin at about 10°S, with almost all rainfall occurring in summer. Over Southern Brazil, rainfall is nearly evenly distributed throughout the year.

A southeastward extension of the wet season maximum at 10°S is known as the South Atlantic convergence zone (SACZ). Satellite-derived estimates of rainfall in the SACZ show the time mean to be at least as strong over the ocean as it is over land (e.g., Nogués-Paegle and Mo 1997), although either the oceanic or continental component may be stronger at a particular time (Carvalho *et al.* 2002, 2004). Downstream convergence of moisture advected from the Amazon Basin by a strong low-level jet flowing southward along the eastern flank of the Andes (e.g., Marengo *et al.* 2004; Salio *et al.* 2007) results in some of the most frequent and largest mesoscale systems on earth over the northern part of La Plata Basin (Laing and Fritsch 2000). Those powerful systems contribute a larger proportion of total rainfall than any other region on earth (E. Zipser, personal communication), and some of them became the most intense convective storms observed by TRMM (Zipser *et al.* 2006).

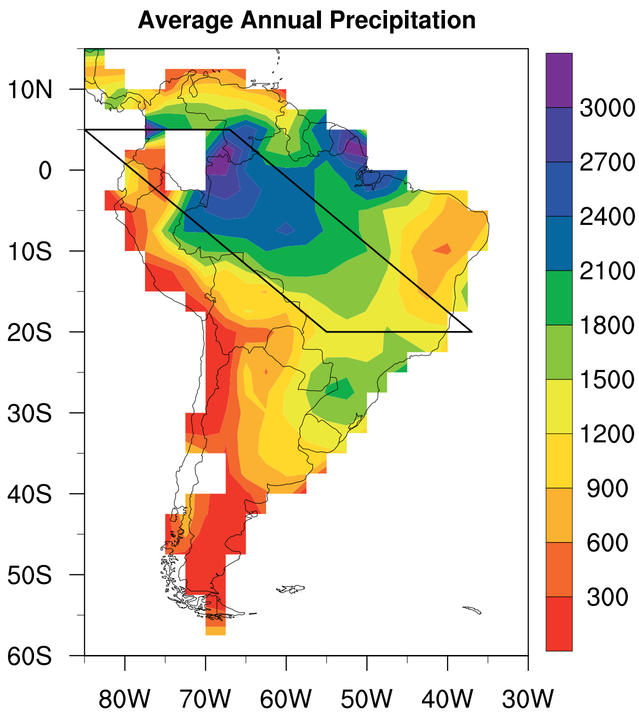

Figure 1. Annual total precipitation in mm averaged from 1976-2004. Parallelogram represents domain used to produce Fig. 4 and is described in text.

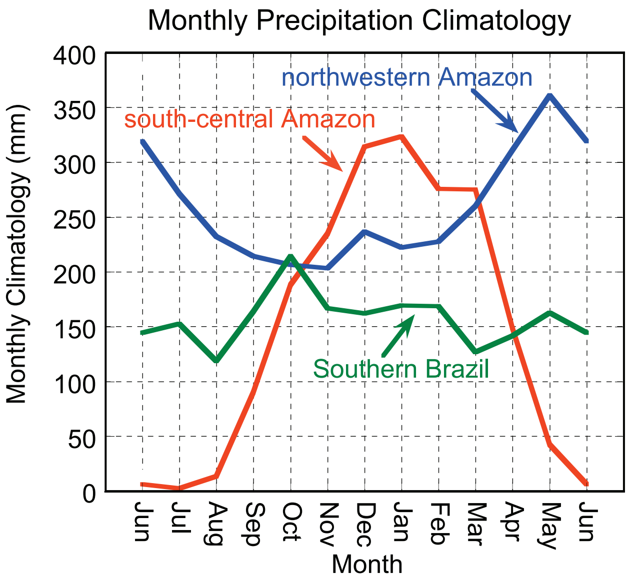

Figure 2. Monthly total precipitation climatology at 2.5° resolution for grid points centered at (0°, 67.5°W) (northwestern Amazon), (10°S, 52.5°W) (south-central Amazon), and (27.5°S, 52.5°W) (Southern Brazil).

4. Monsoon Onset

Kousky (1988) defined monsoon onset at a particular location as occurring when outgoing longwave radiation (OLR) is less than 240 W m^{-2} provided that OLR was above the threshold in 10 of 12 preceding pentads and remained below the threshold in 10 of 12 following pentads. This definition yields a northwest to southeast progression of the onset, and a southeast to northwest progression of the withdrawal. In the perpetually wet northwest Amazon the onset date was not determined because OLR almost always remains low. Marengo *et al.* (2001) used a rainfall-based definition of onset analogous to that used by Kousky (1988). Composites of wind about onset did not reveal a well-defined precursor, suggesting the importance of local thermodynamical processes in providing conditions ripe for onset (Fu *et al.* 1999). Raia and Calvacanti (2008) noted the importance of vapor flux from the Atlantic (Rao *et al.* 1996; Doyle and Barros 2002) and proposed a humidity-based definition of onset.

Marengo *et al.* (2001) noted that increasing the threshold to characterize onset, this appeared to reverse direction. In view of such sensitivity to an arbitrary threshold, Liebmann and Marengo (2001) defined onset as occurring when the accumulation of precipitation exceeds that expected from the annual mean daily average. This definition is local and robust, as the threshold depends on the local climatology. Figure 3 shows the average date of wet season onset using this definition. The wet season progresses northward from an area in southern Brazil, just north of the Paraguay border. Figure 4 illustrates this behavior in a Hovmoeller diagram with time in the abscissa and the zonal average within the west and east limits of the parallelogram in Fig. 1 in the ordinate. Figure 4 shows that along the Equator (at this latitude the zonal average is from 79°W to 61°W) rainfall is at a near-minimum in

December-February, while the wettest season occurs in April and May. Similarly, at about 5°S it begins to rain earlier than at 10°S, but rainfall stays below the climatological average until after the heavy rains begin farther south. Janowiak and Xie (2003) noted that only the South American monsoon onset progresses toward the Equator. See also Fig. 1b of Paegle and Mo (2002).

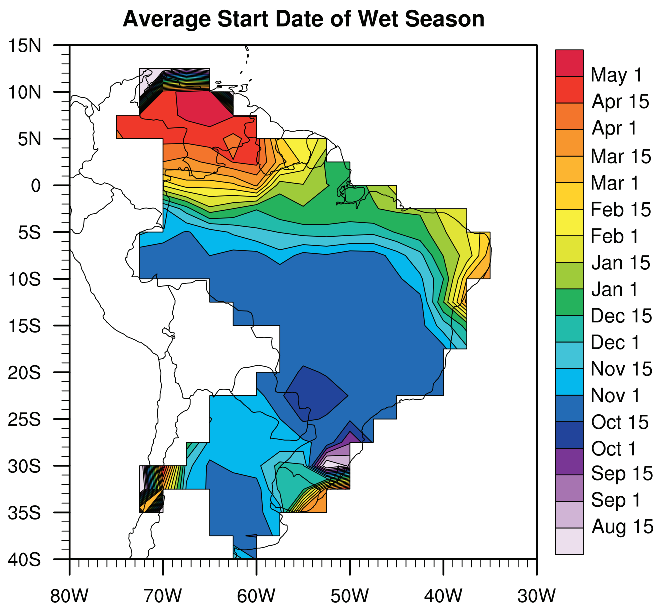

Figure 3. Average start date of wet season. The methodology used to determine starting date is discussed in text.

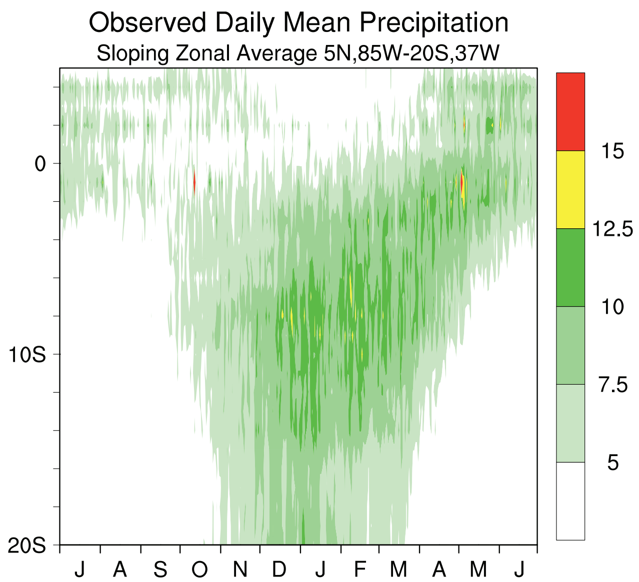

Figure 4. Time-latitude cross-section of climatological (1976-2004) daily rainfall. Zonal average range varies by latitude, and is shown outlined in Fig. 1.

Marengo *et al.* (2001) and Liebmann and Marengo (2001) showed that a substantial fraction of the interannual variability of seasonal precipitation in parts of the Amazon is related to variations in either the onset or ending date of the wet season, and that these variations are often related to sea surface temperatures (SST) anomalies in the tropical Atlantic or Pacific oceans. Their findings are consistent with the hypothesis of Fu *et al.* (1999), who argued that the influence of SSTs on onset should be larger near the Equator (where land-sea temperature contrasts are small), than off the Equator (where onset is controlled mainly by surface heating). In the early stages of monsoon development, soil moisture changes may modulate the development of precipitation (Collini *et al.* 2008).

5. Intraseasonal Variability and Monsoon Breaks

The Amazon Basin itself exhibits little intraseasonal variability in OLR (Jones and Weare 1996; Liebmann *et al.* 1999). This is true for low-level winds as well, except in the southwest (Jones and Carvalho 2002). Jones and Carvalho found local westerly anomalies at low levels in association with an active SACZ, while local easterly anomalies are associated with negative OLR anomalies confined to northern South America.

In the SACZ, there are many more short time scale extreme precipitation events than over the most convective regions of the Amazon (Carvalho *et al.* 2002). Although some mid-latitude fronts propagate into Northeast Brazil (Kousky 1979), many seem to stall as they approach the SACZ motivating its classification as a stationary frontal zone. Casarin and Kousky (1986) and Nogués-Paegle and Mo (1997) identified SACZ anomalies that are part of a dipole pattern, which is present on scales from intraseasonal to interdecadal (Casarin and Kousky 1986; Nogués-Paegle and Mo 1997; Liebmann *et al.* 1999; Paegle *et al.* 2000; Díaz and Aceituno 2003). One phase of the dipole is characterized by an enhanced SACZ and suppressed precipitation to the south, while the other phase is characterized by a suppressed SACZ and increased precipitation over Uruguay, Southern Brazil, and northeastern Argentina (Fig. 5). A strengthening of the low level jet east of the Andes and associated transports of massive amounts of moisture from the Amazon Basin into the subtropics accompanies the latter phase (Berri and Inzunza 1993; Douglas *et al.* 1998; Paegle 1998; Marengo *et al.* 2004; Salio *et al.* 2007). Strong jet events are linked to short-term extreme precipitation events in the plains of central Argentina (Nicolini *et al.* 2002; Liebmann *et al.* 2004a; Salio *et al.* 2007).

Somewhat consistent with this picture is the existence of different convection "regimes" in Amazonia as identified in recent field campaigns. An intense mode consisting of vertically developed convection is associated with an easterly wind regime, while a weaker, monsoon-type mode, is associated with a westerly wind regime (Herdies *et al.* 2002; Ferreira *et al.* 2003). Ma and Mechoso (2007) demonstrated that lag composites of geopotential height anomalies at 700 mb prior to the establishment of the wind regimes in Amazonia show the development of structures in the South Pacific that resemble the principal modes of variability in that region. This indicates the existence of links between the wind regimes and the modes of intraseasonal variability over South America. There are also suggestions of links between the wind regimes and a continental-scale eddy centered off southeast South

America (Fig. 6; see Mechoso *et al.* 2005). In its cyclonic phase, the SACZ intensifies with anomalous descent to the southwest and weakened low-level flow east of the Andes; the anticyclonic phase shows a weak SACZ and anomalous anticyclone.

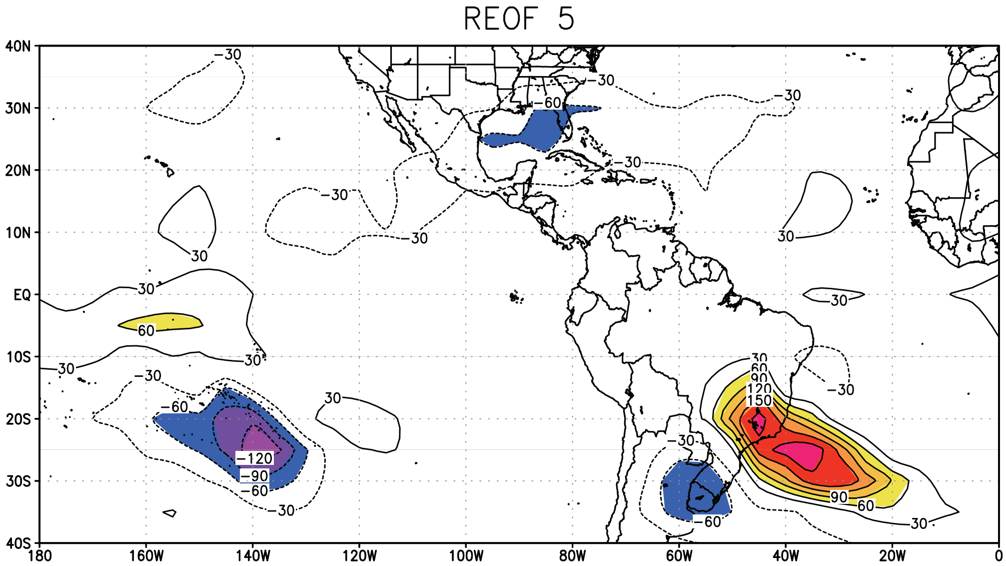

Figure 5. Fifth rotated empirical orthogonal function pattern of filtered OLR (filter retains 10-90 day periods). (from Nogués-Paegle and Mo 1997)

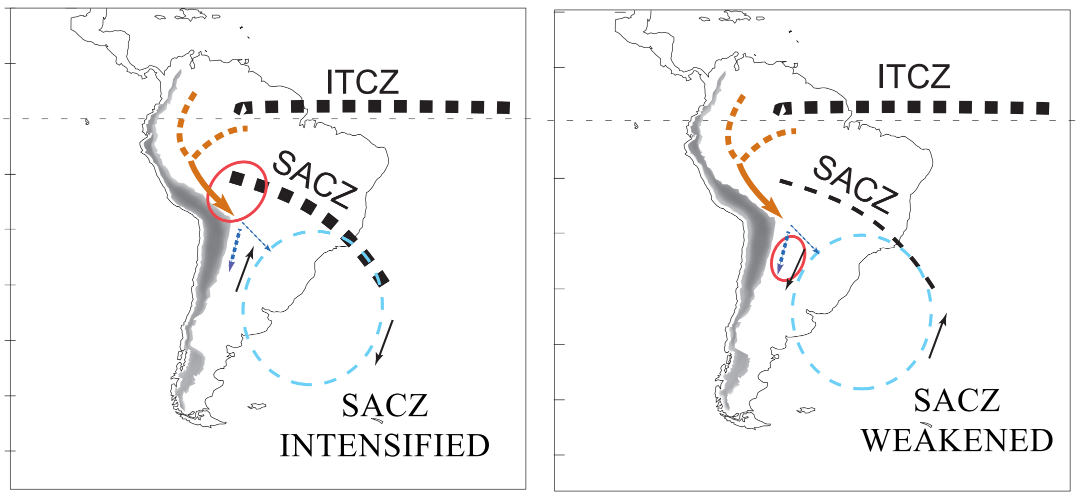

Figure 6. Opposite phases of the dominant mode of variability over South America during the warm season. Thick arrows indicate low level jets. The areas bounded by red circles are those in which enhancement of mesoscale convective systems is expected. (from Mechoso *et al.* 2005)

The Madden-Julian oscillation (MJO; e.g., Madden and Julian 1994) is presently the best-understood source of intraseasonal variability to affect South America. Further, it is the only source of intraseasonal variability for which potential predictability has been demonstrated (e.g., Ferranti *et al.* 1990; Jones *et al.* 2004). Casarin and Kousky (1986) noted anomalies in the Southern Hemisphere subtropics that affect the SACZ move eastward at about the same speed as the 30-60 day oscillation. Kiladis and Weickmann (1992) regressed 30-70 day filtered OLR in the western Pacific onto 200 mb wind and found (primarily zonal) anomalies over near-equatorial eastern South America. Berbery and Nogués-Paegle (1993) posited that during summer, Rossby waves propagate from a source in the vicinity of Australia along a ray path to South America. Nogués-Paegle and Mo (1997) found that the SACZ tends to be enhanced when convection is enhanced near the dateline and suppressed over Indonesia, which they speculated to be related to the 30-60 day oscillation. Paegle *et al.* (2000) showed that the SACZ is strong when the MJO-related mode is reinforced by a higher-frequency mode propagating into the SACZ from midlatitudes. Westerly jets act as strong wave guides in the vicinity of South America (Ambrizzi *et al.* 1995; Ambrizzi and Hoskins 1997). Equatorward propagation is evident in the exit region of the subpolar jet stream over South America (Berbery *et al.* 1992). Carvalho *et al.* (2004) quantified that an intense SACZ most often persists during the end of the MJO life cycle, which is consistent with the anomalies described by Nogués-Paegle and Mo (1997). Liebmann *et al.* (2004a) obtained a consistent result and argued that the MJO influenced the phase of the midlatitude synoptic disturbances that affect the SACZ.

6. Interannual Variability

Interannual variability of precipitation in the Amazon Basin is small compared to the annual cycle amplitude (Horel *et al.* 1989). Nonetheless, there can be large anomalies in individual seasons (Marengo and Hastenrath 1993; Williams *et al.* 2005; Marengo *et al.* 2008). The most studied aspect of interannual variability is its association with the phase of the El Niño/ Southern Oscillation (ENSO) cycle, although ENSO itself does not account for a large fraction of the total interannual variability (e.g., Rao and Hada 1990; Marengo 1992; Nobre *et al.* 2006). Many studies have concluded that during the warm phase of ENSO, precipitation tends to be below average in northern South America during the summer wet season and above average in southeastern South America (e.g., Ropelewski and Halpert 1987; Aceituno 1988; Kiladis and Diaz 1989; Kousky and Ropelewski 1989; Rao and Hada 1990; Pisciottano *et al.* 1994; Grimm *et al.* 1998, 2000; Grimm 2003; Berri *et al.* 2002; Mechoso and Perez Iribarren 1992; Paegle and Mo 2002; Coelho *et al.* 2002; Ronchail *et al.* 2002). The suppression of precipitation over northern South America during El Niño probably results from anomalous subsidence caused by an anomalous Walker circulation (e.g., Kousky and Ropelewski 1989; Buchmann *et al.* 1989; De Souza and Ambrizzi 2002). The subtropical enhancement of rainfall may be related to a change in the Rossby wave pattern emanating from the tropics (e.g., Kidson 1999; Grimm 2003) or to a change in the regime of persistent anomalies (Cazes *et al.* 2003; Robertson and Mechoso 2003). Ropelewski and Bell (2008) showed in this region a robust shift in histograms of daily gridded precipitation toward an increase in wetter amounts, including an increase in heavy precipitation events, during the wet season of an El Niño year. During the cold phase of ENSO (La Niña), anomalies are approximately opposite to those during El Niño (Grimm 2004).

The impacts of ENSO vary throughout the wet season. Grimm (2003) found large anomalies associated with El Niño using monthly data that may be washed out or not emerge at all using seasonal anomalies. This finding led to the suggestion that regional processes compete with remote influences during at least part of the season. In spring and early summer of an El Niño year, remote influences dominate (Grimm 2003). Anomalous rising motion over the eastern Pacific results in anomalous subsidence over the Amazon and anomalous Rossby wave activity propagating into South America via midlatitudes, resulting in anticyclonic low-level anomalies over tropical South America and southeastern Brazil. These anomalies divert Atlantic moisture into northern South America and Southern Brazil, resulting in negative precipitation anomalies in north and central-east Brazil and positive anomalies in south Brazil (Grimm 2003). The negative precipitation anomalies in central-east Brazil (including the highlands of Southeast Brazil) during spring cause warm surface temperatures and an anomalous low-level cyclonic circulation in southeastern Brazil during the peak monsoon season. Thus, moisture flux towards central-east Brazil is enhanced (Grimm 2003; Grimm *et al.* 2007), and precipitation increases there. Furthermore, the dry anomalies in north Brazil are displaced northward, and the positive anomalies in south Brazil are reduced.

In February precipitation decreases as surface temperature anomalies in central-east Brazil become negative after the above-average precipitation during January. Negative precipitation anomalies prevail in north Brazil and in the SACZ, and positive anomalies return to south Brazil (Grimm 2003). A schematic of the sequence proposed is shown in Fig. 7. A nearly opposite sequence occurs during La Niña years (Grimm 2004). The reversal of anomalies from spring to summer in central-east Brazil also occurs in other years, provided that spring anomalies are large (Grimm *et al.* 2007; Grimm and Zilli 2009).

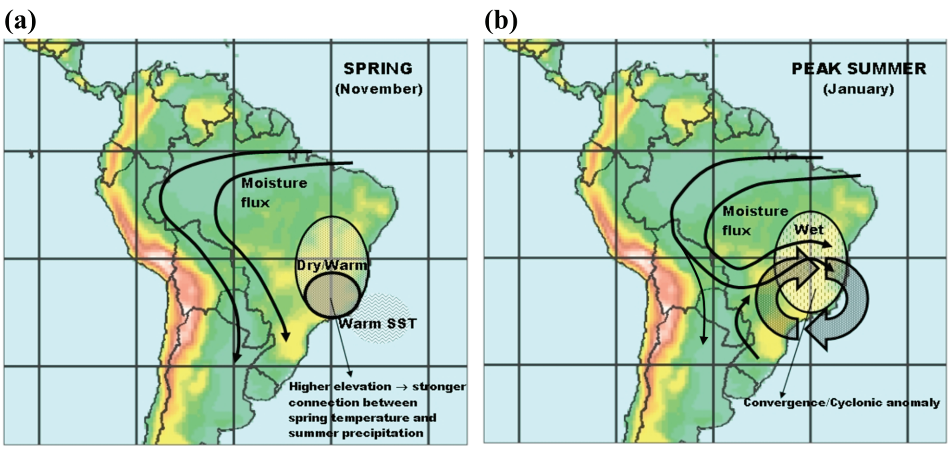

Figure 7. Schematic evolution from (a) spring dry conditions to (b) peak summer wet conditions in central-east Brazil, through decreasing low-level pressure, convergence, and a cyclonic anomaly over southeast Brazil. (from Grimm *et al.* 2007).

Studies of the Atlantic SST influence on monsoon precipitation are less numerous than studies of the Pacific SST/Southern Oscillation relationship with the monsoon. Most have shown a clear relationship between Atlantic SST and precipitation in Northeast Brazil (e.g., Mechoso *et al.* 1990; Hastenrath and Greischar 1993a; Uvo *et al.* 1998; Paegle and Mo 2002; Ronchail *et al.* 2002). The association is most pronounced from March to May, when precipitation amounts in the northern Amazon are largest (Marengo and Hastenrath 1993). Positive precipitation anomalies occur when Atlantic SST anomalies are positive south of the Equator, resulting in a southward displacement of the ITCZ (e.g., Hastenrath and Greischar 1993b). To the extent that these anomalies associated with a displaced ITCZ occur early and extend westward into the Amazon, this may be a mechanism by which Atlantic SST influences monsoon rainfall.

Ferreira *et al.* (2003) found that the South American low-level jet in January-March was much stronger and transported more moisture poleward during El Niño in 1998 than during La Niña in 1999. This is consistent with Amazon convection being suppressed by the enhanced moisture outflow, and with twice as much rainfall in the subtropics during El Niño than La Niña.

The low level jet (and SACZ) are also affected by cold surges that propagate from southern South America into the Amazon (Saulo *et al.* 2004; Wang and Fu 2004). These surges (e.g., Marengo *et al.* 1997; Garreaud and Wallace 1998; Garreaud 2000a) can reach the Equator and enhance precipitation there (Liebmann *et al.* 2009).

7. Long-term Trends

Trend analysis of precipitation in the monsoon region is made nearly impossible by a lack of observing stations (e.g., Haylock *et al.* 2006) and short, incomplete records of stations that do exist (e.g., Liebmann *et al.* 2004b). Studies with limited data have indicated statistically insignificant positive trends in the Amazon Basin. The Intergovernmental Panel on Climate Change report (Trenberth *et al.* 2007) confirms that insufficient data records are available over much of monsoonal South America to support any strong conclusions about the existence of precipitation trends there.

River discharge records, on the other hand, can be useful, as they integrate over large areas and are often longer and relatively complete compared to those of precipitation (Hastenrath 1990), although one must account for artificial and natural river-bed changes (e.g., Marengo *et al.* 1998). Still, results may vary substantially depending on period of analysis. For example, García and Mechoso (2005) found positive trends in all the major South American rivers since the early 1970s. On the other hand, Richey *et al.* (1989) and Marengo *et al.* (1998) did not detect long-term trends in Amazon Basin discharge. As illustration, river flow data provided by Dai *et al.* (2009) are examined. See their paper for details on how missing data were filled.

Figure 8a shows the annual discharge rate of the Amazon at Obidos, Brazil for averages of varying lengths and periods (C. Jones, personal communication). The flow at Obidos captures waters from the basin west of ~55°W. The ordinate displays the averaging length and the abscissa shows the ending year for each of those averages. For example, the value at y=30, x=1980 is the average annual flow from 1951 to 1980. The anomalies with the longest-lasting influence are the low flows in the later 1960s, high flows in the mid-1970s (Dai *et al.* 2009) and low flows again in the early 1980s. While the periods of large anomalous flow can last several years, from a perspective of ~15 year averages or longer, they represent little more than noise, as the long-term average is more-or-less constant.

The pattern of flow in southeastern South America is quite different from the Amazon. Figure 8b shows the Paraná River flow at Timbúes, Argentina (61°W, 33°S). At this point the river has captured most of the La Plata Basin runoff, including the Paraguay. The available record is longer than that of the Amazon (Fig. 8a). From mid-1982 to near the end of 1983 the flow was remarkably large, with the most severe flooding on record starting in May 1983 (Camilloni and Barros 2003). It is widely thought that this extreme event was related to the 1982-83 El Niño (e.g., Grimm *et al.* 2000). While those flows have not been equaled since, they have remained high since that event compared to the pre-event flows (Fig. 8b). Camilloni and Barros (2003) found that flood incidence (defined as at least two standard deviations above the monthly climatology) was almost 6 times higher in the 20 years from 1980-1999 than in the 60 years from 1920-1969.

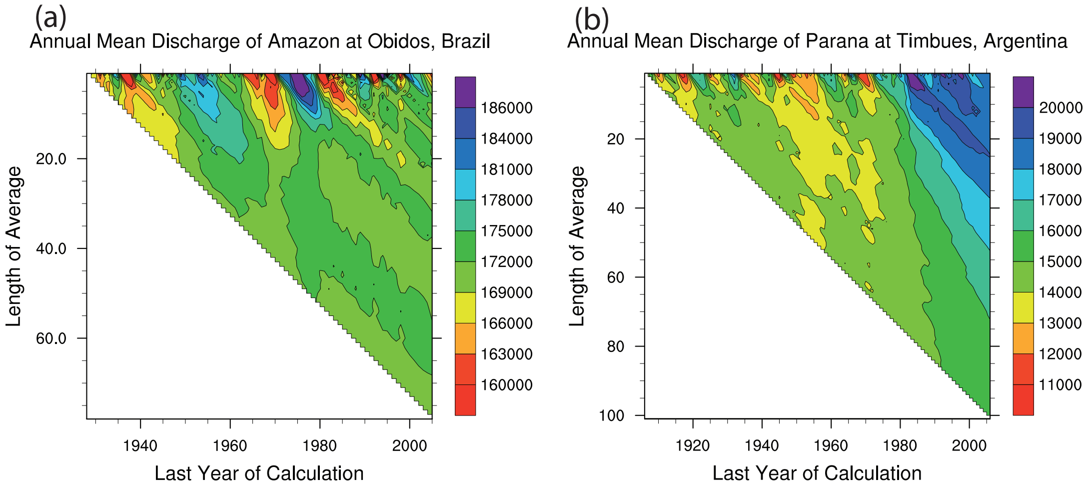

Figure 8. Annual mean discharge rate of (a) Amazon at Obidos, Brazil (55.5°W, 2.5°S) and (b) Paraná at Timbúes, Argentina (60.7°W, 32.7°S) in $m^3\ s^{-1}$, as a function of length of average and ending year of calculation.

Using 30-year averages of data from the early 20th century through 1995, Genta *et al.* (1998) found an upward trend in the Paraguay and Paraná rivers. Both these rivers extend into the edge of the region under direct monsoon influence. Flows increased from the mid-1960s and leveled off near the end of the record. Consistent with this finding, Dai *et al.* (2009) found a statistically significant upward trend in the Paraná downstream of its confluence of the Paraguay from 1948-2004 that leveled off after 1997. Robertson and Mechoso (1998) also detected a nonlinear upward trend in the annual Paraguay-Paraná average flows from 1911-1993, during which the flows decreased for a decade until 1960 and then increased rapidly until 1980 (see also García and Mechoso 2005). These papers documented decadal variations in Paraguay-Paraná flows.

The trend analyses of the Paraná River are conveniently summarized by Fig. 9, which shows every possible change (longer than 2 years) at Timbúes as a function of length of calculation and ending year of calculation. It is important to note that the quantity plotted is the annual linear trend multiplied by the length of the calculation. Generally, in spite of multiplying the annual trend by the length of the calculation, longer records exhibit smaller trends. For the available record, almost the entire increase occurred from 1968-1983. That 16-year sequence is exceptional because it has the largest change of any period longer than 6 years. The change from 1906-1968 is weakly negative, as it is from 1984-2006.

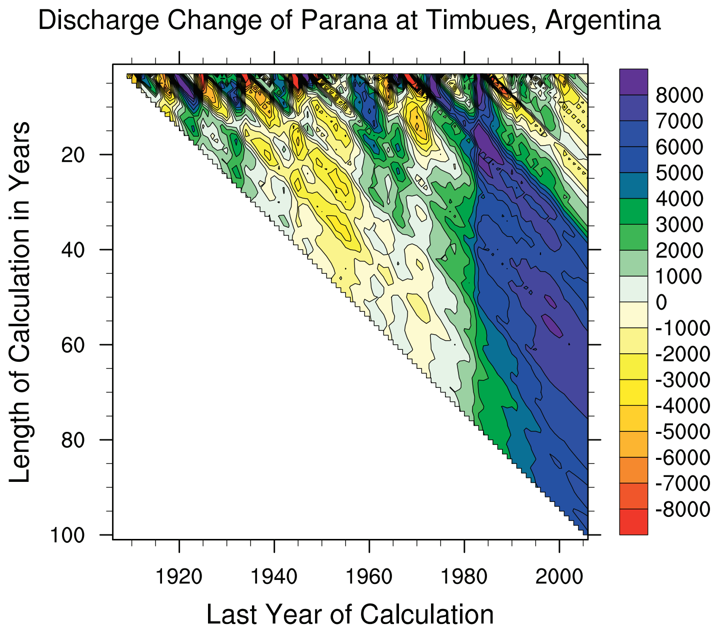

Figure 9. Change of mean annual discharge rate (in $m^3 s^{-1}$) of Paraná River at Timbúes, Argentina as a function of varying length of calculation and ending year of calculation. Value at each point is computed as the linear trend multiplied by the length of calculation.

8. Environmental Changes

Deforestation is the most direct anthropogenic forcing of SAMS. The Brazilian National Institute for Space Research (INPE) estimates that as of the early 21st century, about 15% of the Amazon has been deforested, and the rate of deforestation is increasing. If the trend continues, 40% of the Amazon will be deforested by 2050 (Soares-Filho *et al.* 2006), much of it replaced by pasture or soybean. Should this occur, 25% of the mammal species in the Amazon basin will lose more than 40% of the forest within their ranges (Soares-Filho *et al.* 2006).

Deforestation and loss of canopy results in a larger amplitude diurnal cycle of surface temperature and increased average surface temperature and longwave emission. There is also a shift in the diurnal cycle, owing to the reduced thermal capacity of pasture compared to forest (Gash and Nobre 1997). Surface albedo also increases (Culf *et al.* 1995). Thus there is a net radiation decrease (Gash and Nobre 1997). Higher surface temperatures exist in spite of a net radiation decrease because the shallow rooting depth of pasture means that it dries out quicker than forest, especially in the dry season (e.g., Wright *et al.* 1996), and so more radiation is converted to sensible heat (Gash and Nobre 1997). In contrast, the forest, with its deeper root structure, shows no decrease in evaporation during the dry season (Hodnett *et al.* 1996; Wright *et al.* 1996), although albedo does increase (Gash and Nobre 1997).

Modeling studies indicate that the end result of deforestation on the large scale is a decrease in precipitation (e.g., Shukla *et al.* 1990; Nobre *et al.* 1991; Dickinson and Kennedy 1992; Lean and Rowntree 1993; Sud *et al.* 1993; Hahmann and Dickinson 1997; Costa and Foley 2000; Sampaio *et al.* 2007). Replacement of forest by soybean rather than pasture, which is increasingly common, leads to a larger decrease in precipitation, owing to the higher albedo of soybean (Costa *et al.* 2007). Xue *et al.* (2006) showed that consideration of explicit vegetation processes in an atmospheric GCM does not alter the monthly mean precipitation at the planetary scale. On continental scales, however, explicit vegetation contributes to an improved simulation of SAMS.

Complete Amazon deforestation will likely change global precipitation through changed patterns of large-scale wave propagation (e.g., Gedney and Valdes 2000; Avissar and Werth 2005). Wang *et al.* (2000) noted Eltahir and Bras (1994) had confirmed that at smaller scales (~250 km) deforestation might reduce rainfall but not affect the large-scale circulation.

Wang *et al.* (2000) simulated precipitation changes associated with deforestation in Rondônia, Amazonia, using a mesoscale model with an innermost domain resolution of ~4 km. They noted the scale of deforestation to be ~10 km. In the wet season, synoptic forcing dominates and the effects of land-use changes are negligible. During the dry season, however, shallow clouds were enhanced, which has been observed over deforested areas (Cutrim *et al.* 1995).

A second major anthropogenic influence on SAMS is likely to be a continued increase of carbon dioxide (CO_2). According to Costa and Foley (2000) and Notaro *et al.* (2007), increasing CO_2 has radiative and physiological consequences. The radiative influence warms the surface, resulting over Amazonia in soil drying and reduced forest cover (Notaro *et al.* 2007). The physiological effect decreases evapotranspiration, as plants reduce their stomatal openings to maximize their intake of CO_2 and minimize water loss (Notaro *et al.* 2007), while the radiative effect increases evapotranspiration by providing more energy (Costa and Foley 2000). Costa and Foley (2000), using an atmospheric GCM, found an increase in Amazon precipitation from a doubling of CO_2, but cautioned that this result is highly uncertain and depends largely on the model used.

Costa and Foley (2000) carried out a GCM experiment to estimate the combined effects of deforestation and increased CO_2. They noted that while deforestation decreases evapotranspiration and precipitation, increased CO_2 does the opposite. In addition, both deforestation and CO_2 increase result in warmer surface temperature. Deforestation dominates, and the net effect is a decrease in rainfall. In this experiment, however, the entire rainforest was replaced by grass, which may be a too extreme scenario.

9. Challenges and Recommendations for Future Research

The challenge of predicting climate variability on scales ranging from intraseasonal to decadal and beyond is a laudable, yet daunting, challenge. Climate results from interactions on an extraordinarily broad range of scales, virtually all of which are important. Research must be conducted in context of the climate system, with results understood in the context of previous work.

Before reliable short and long-range forecasts can be developed, the present-day climate must be accurately simulated in coupled models. The SACZ is extremely important to the climate of South America, in part because it lies over some of the most populated areas of the continent. It is a matter of great concern, therefore, that this major feature of SAMS is not well reproduced in coupled ocean-atmosphere GCMs. While a few of the models are able to approximate the position and intensity of the SACZ, in most of them it is displaced to the north and is not distinct from the intertropical convergence zone that in observations is located near and parallel to the Equator (Vera *et al.* 2006b).

Land surface processes seem to play an important role in SAMS. Obtaining accurate observational estimates of surface quantities such as soil moisture and evapotranspiration, which could be used to validate model outputs, is a painstaking, yet crucial aspect to improving understanding, simulations, and predictions of the South American warm-season climate.

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