THE AMERICAN MONSOON SYSTEMS

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Abstract

This paper examines similarities and differences among major features of the North and South American monsoon systems. Over both North and South America the summertime circulation shows upper-level anticyclone/low-level heat low structures. These develop at different distances from the equator. It is argued that ascent to the east where subtropical convergence zones develop, and subsidence over the cool waters of the eastern Pacific where stratocumulus decks provide a radiative heat sink to the tropical atmosphere are integral and unifying aspects of both monsoon systems. The intraseasonal and interannual variability of the systems are contrasted. The reported links between anomalies in soil conditions and sea surface temperatures are marginal, and consistently long-range predictability is low. Ropelewski *et al.* (2004) and Grimm *et al.* (2004) focus more closely on each of the American monsoon systems in companion papers.

1. Introduction

Whilst there could be some debate as to whether the seasonal changes in the atmospheric circulation over the Americas satisfy the conditions set by Ramage (1971) for an “official monsoon label”, there is no doubt that in this particular case the issue reduces to reaching quantitative thresholds while qualitative criteria are met.

The warm season flow over the Americas shows the classical monsoon-type surface low pressure /upper-level anticyclone and intense low-level inflow of moisture from the ocean. The flow is affected by large-scale land-sea surface temperature contrasts, as well as by land-atmosphere interactions associated with elevated terrain and land surface conditions (e.g. soil moisture). Associated seasonal changes in regional precipitation show the shift from low or relatively low to very intense.

We will refer to the North American and South American warm season circulations in the tropics as the North American and South America Monsoon Systems (NAMS and SAMS, respectively). Both the NAMS and SAMS provide a useful framework for describing and diagnosing warm season climate. Climate anomaly patterns during the warm season can be characterized in terms of changes in the intensity and/or features of either the NAMS or SAMS. For example, the summertime precipitation regime over North America during the 1988 spring/summer drought or the 1993 summer flood closely mimic a weakening and amplification of the NAMS, respectively.

Our focus in this paper is on the similarities and differences among major features of the NAMS and SAMS. Ropelewski *et al.* (2004) and Grimm *et al.* (2004) give a closer examination of each system in companion works. Higgins *et al.* (2003), Paegle *et al.* (2002) and Vera et al (2004) are other relevant recent papers on the American monsoons, to which the reader is referred for an extensive bibliography on the subject. We start in section 2 with an overview of the monsoon systems over the Americas. Sections 3 and 4 outline the intraseasonal and interannual variability of the systems, respectively. Section 5 presents a discussion on the dynamics of the systems, and section 6 deals with predictability aspects.

2. Structure of the Monsoon Systems over North and South America

The Americas form a landmass of great meridional extent, reaching unbroken from over 50S to over 70N. The equator intersects South America, which forms a cone, narrowing down with increasing latitude. The west coast of Central and North America tilts in the northwest/southeast direction, while that of South America in the tropics does not have such a pronounced tilt. High mountain ranges extend along the Pacific coast of both continents. The Andes, in particular, effectively block the influence at low levels of the Pacific Ocean on the climate of South America.

Another key feature that affects, and is affected by, the monsoon systems is the sea surface temperature (SST; Fig. 1). During the warm season, tropical North American is flanked to the west by the eastern Pacific warm pool extending to about 20°N and by the cold Pacific waters off-California north of that latitude, and to the east by the warm waters of the Gulf of Mexico and Caribbean. Tropical South America is flanked by the cold Pacific waters off Peru and Ecuador and by the warm waters of the tropical Atlantic. Consistently with the continental spread across the equator, the seasonal evolution of precipitation shows a migration in latitude with maximum values around the equator during the solstices.

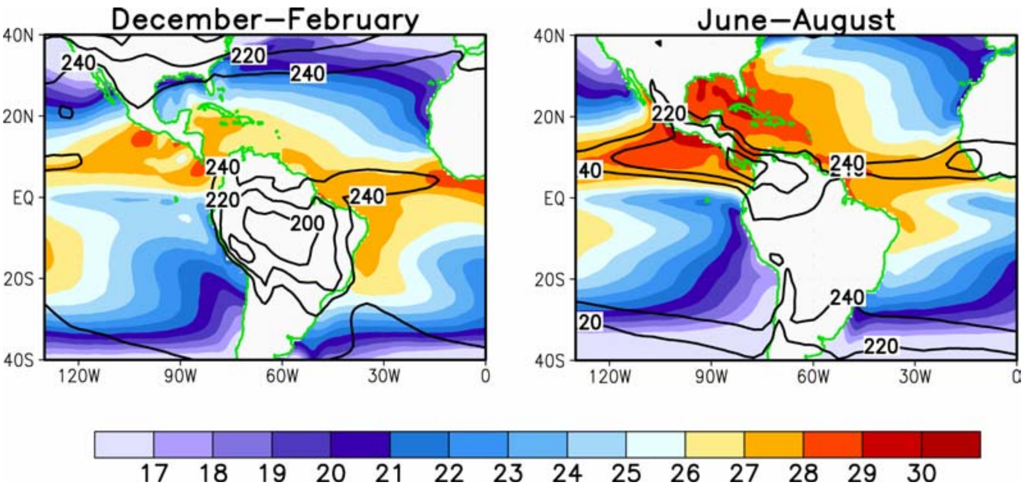

Figure 1. Distributions of sea surface temperature (ºC, shading) on outgoing longwave radiation (Wm^{-2}, contours) for December-February (left panel) and June-August (right panel). (Courtesy V. Kousky.)

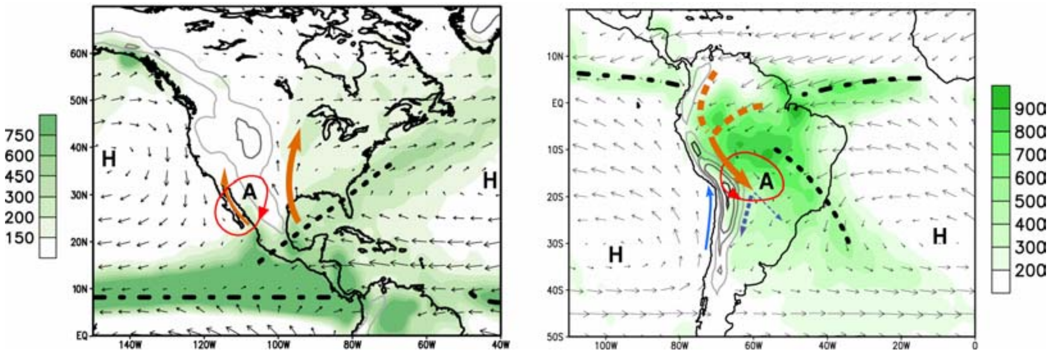

Figure 2. Schematic illustration of the North and South American monsoons systems (left and right panels, respectively). Shading indicates precipitation and dashed lines indicate convergence zones. Small arrows show low-level (900 hPa) winds, and thick arrows represent low-level jets. An "H" shows a subtropical surface high center, and an "A" indicates the monsoon anticyclone. (Adapted from V. Kousky.)

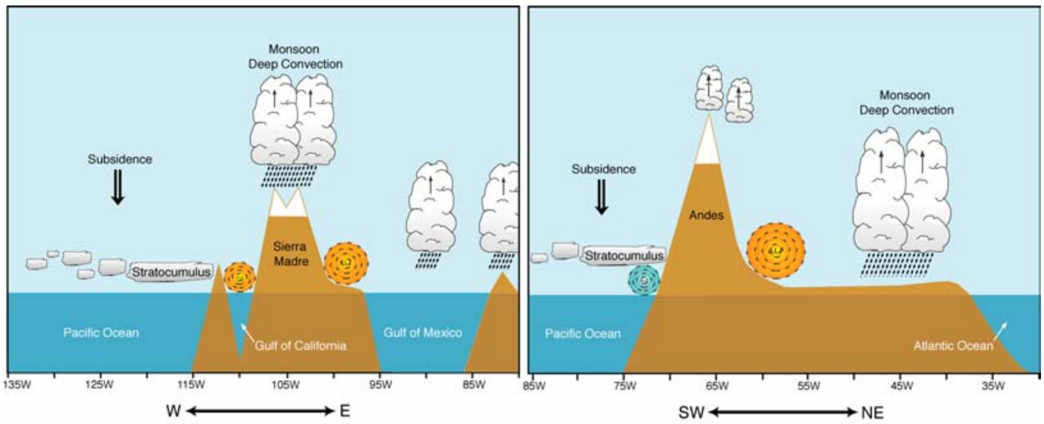

Figure 3. Schematic vertical section for the corresponding summer season at around 30°N (left panel) and southwest-northeast (right panel). Regions of deep convection and low-level jets are indicated. (Panel for NAMS adapted from W. Higgins.)

Figures 2 and 3 show schematically the major features of the warm season circulation over North and South America. The NAMS is characterized by a region of intense precipitation emanating from the eastern Pacific intertropical convergence zone (ITCZ), extending northward over Mexico to the southwest United States (U.S.), with largest values over the western slopes of the mountain range. High values of precipitation also extend northeastward over the Gulf of Mexico, reaching up along the eastern flank of North America, and merging into the North Atlantic storm track. There is also a relative maximum to the southwest of the Great Lakes. The continental east-west contrast between the arid west and the humid east is a key characteristic of a monsoonal circulation, which we elaborate upon in section 5. The upper-level monsoon anticyclone associated with the NAMS shifts northward with season from southwestern Mexico to northwestern Mexico and southwestern United States. Moisture transport onto the North American continent is associated with broad-scale advection from the Gulf of Mexico, and with important low-level jets (LLJs) over the Gulf of California and east of the Rockies. The latter LLJ is a warm season, primarily nocturnal, feature; less detail is known about the other one.

The SAMS is characterized by intense precipitation over central Brazil and Bolivia, in a region that is linked to the Atlantic ITCZ to the northeast. The extension of the SAMS precipitation into the South Atlantic Convergence Zone (SACZ) to the southeast mirrors the northeast extension of NAMS precipitation. In both cases, there is a substantial maritime component on the western flank of the subtropical Atlantic anticyclones. The upper-level anticyclone associated with the SAMS ("Bolivian High") establishes close to the Altiplano. The trade winds from the tropical Atlantic Ocean provide the moisture source for the SAMS. Moisture transport intensifies locally along the eastern scarp of the Andes, where the South American LLJ (SALLJ) develops with strongest winds over Bolivia. In contrast to NAMS, the SALLJ is present throughout the year and is not solely a warm season feature.

Figure 3 includes a sketch of the descending motion associated with the NAMS and SAMS. An integral and unifying aspect of both monsoon systems is the subsidence over the cool SSTs of the eastern Pacific. Here extensive stratocumulus decks provide a radiative heat sink to the tropical atmosphere that can balance the adiabatic warming due to the monsoonal descent. These stratocumulus decks arguably provide a direct coupling between both American monsoons and the Pacific Ocean. We return to this feature in section 5 of the paper. It has also been suggested that descent associated with SAMS occurs in the tropics across the equator, where it establishes links with the North Atlantic climate (e.g. Robertson *et al.* 2000).

3. Intraseasonal Variability

Precipitation during NAMS shows a relative minimum in the warm season along the Sierra Madre Oriental and the Caribbean. There does not seem to be such a feature in SAMS. The reasons for the NAMS feature are related to the seasonal displacements of the ITCZ in the eastern Pacific. SAMS interacts less directly with the tropical ITCZs.

During the warm season, tropical intraseasonal oscillations such as the Madden Julian Oscillation (MJO) modulate a number of different weather phenomena affecting both the NAMS and SAMS (e.g. tropical cyclones, tropical easterly waves, Gulf of California surges). MJO-related impacts are linked to more regional meridional adjustments of the precipitation pattern over the eastern tropical Pacific.

Figure 4 is a schematic of the 700-hPa circulation for wet and dry moisture surges in Arizona, U.S. (AZ). Whether a surge is "wet" or "dry" depends on the relative location of the upper-level monsoon anticyclone at the time of the gulf surge. If the ridge axis is an eastward position, the situation is wetter-than-average in AZ and to the east, while if the ridge is towards the west then the situation is drier-than-average in the same location. Roughly one-half of gulf surges are not associated with enhanced precipitation in AZ.

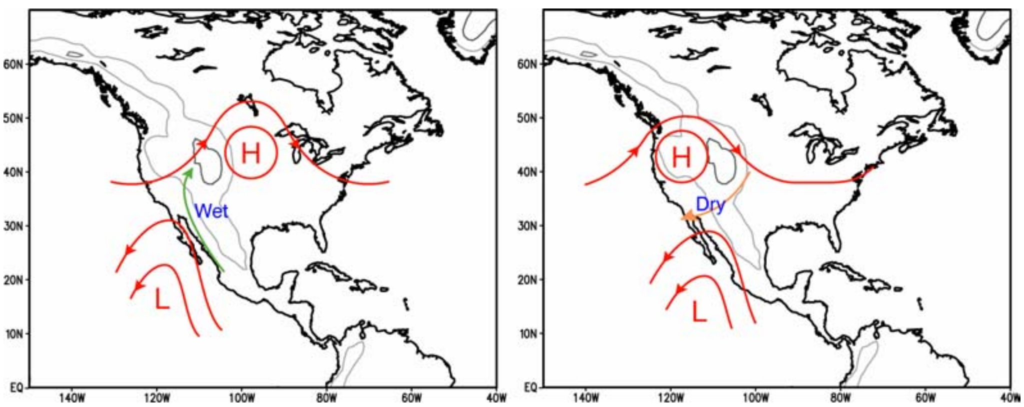

Figure 4. Schematic illustration of the circulation during a dry (left) and wet (dry) Gulf of California surge event for Arizona. (Courtesy Wayne Higgins)

The intraseasonal (interannual and even intedecadal) variations of SAMS appear to be associated with a continental-scale eddy (Fig. 5). In the cyclonic phase of the circulation, the SACZ intensifies with anomalous descent to the southwest and weakened low-level flow east of the Andes; the anticyclonic phase shows opposite characteristics. The vertical velocity distribution associated with the mode is consistent with the reported dipole defined by persistent wet and dry anomalies over tropical and subtropical eastern South America during the austral summer, with one center over southeastern Brazil in the vicinity of the SACZ and another center over southern Brazil, Uruguay and northeastern Argentina. 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).

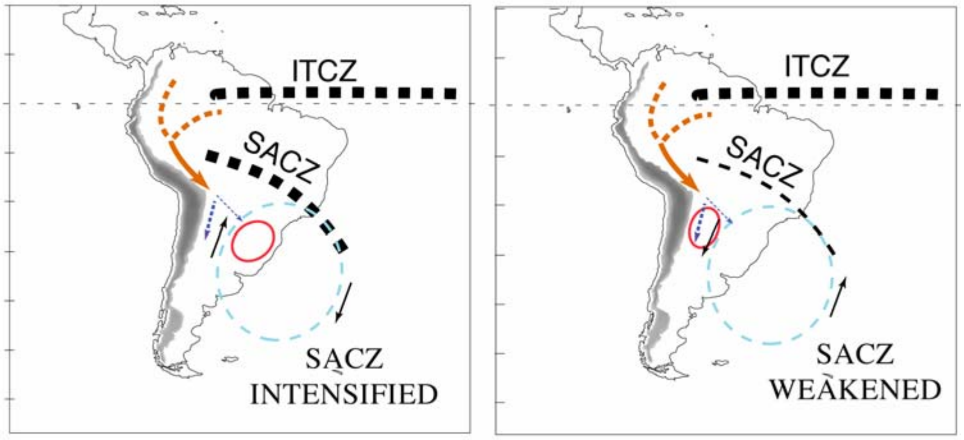

Figure 5. Opposite phases of the dominant mode of variability over South America during the warm season. Thick arrows indicate low level jets. The area bounded by a red circle is one in which enhancement of mesoscale convective systems are expected.

4. Interannual Variability

The continental-scale pattern of NAMS interannual variability shows that anomalously wet (dry) summers in the southwest U.S. tend to be accompanied by similar conditions over the southeast U.S and by dry (wet) summers in the Great Plains of North America. The SAMS exhibits a similar type of behavior, with a dipolar relationship between precipitation over the SACZ and over southeastern South America. Warm seasons with an active SACZ tend to be accompanied by dry conditions in southeastern South America, and vice-versa, although ENSO effects modulate this tendency.

Some studies have reported that the intensity and extent of NAMS is correlated with SST anomalies in the Gulf of California. According to one of those studies, anomalously wet monsoon years in AZ are associated with significantly higher SSTs (>29°C) in the northern Gulf than dry years. The association of SST anomalies in the Atlantic and SAMS is more subtle. Northeasterly and southeasterly trade winds advect moisture from the tropical Atlantic over the continent, but this moisture has to travel some distance before precipitating in the monsoon region. There is a suggestion that cold SST anomalies in the tropical North Atlantic are associated with stronger SAMS rains. On decadal timescales, there is evidence from river-flow records of a relationship between the North Atlantic and precipitation over the subtropical plains of South America.

ENSO can potentially exert an influence on the NAMS and SAMS through several pathways. Changes in the Walker and local Hadley circulations can modulate the monsoonal divergent circulations. The potential for this influence is greater for the SAMS due both to its being closer to the equator, as well as the seasonality of ENSO whose mature phase develops in the northern (southern) cold (warm) season. A second pathway of influence may be through ENSO's effect on eastern Pacific SSTs, both through changes in the region of warm SSTs off the west coast of Mexico (for NAMS) and stratus decks (especially for SAMS).

For both the NAMS and SAMS in the equatorial belt, El Niño and La Niña tend to be associated with anomalously dry and wet events, respectively. During El Niño the ITCZ shifts toward the equator, the Hadley circulation intensifies in the eastern Pacific, and there is a tendency for dry conditions over Mexico. The opposite conditions develop during La Niña. During the northern winter, there are significant and positive correlations between SST anomalies in the eastern equatorial Pacific and precipitation anomalies over the southwestern U.S. There is also evidence that northern winters characterized by wet (dry) conditions are often followed by dry (wet) conditions in the southwest U.S. (i.e., stronger and weaker NAMS, respectively). The out-of-phase relationship between precipitation in the southwestern/southeastern U.S. and the Great Plains of North America suggests that summer drought (flood) episodes in the latter region are (at least indirectly) related to anomalies the preceding winter. These associations between anomalies two seasons apart indicate the existence of mechanisms that provide a memory for the system. We return to this issue in section 6. For SAMS in the equatorial belt, a qualitatively similar and marginally significant El Niño-dry/La Niña-wet relationship has been suggested. The mechanism in this case seems to be associated with anomalies in the Walker circulation. During El Niño convection increases over the eastern tropical Pacific and subsidence increases over equatorial South America, which disfavors convection.

There is still another pathway for ENSO influence on the NAMS and SAMS. This involves the extratropics and the excitation of the Pacific North and South American Rossby wave trains or teleconnection patterns. For the SAMS, the influence is related to the subtropical eddy-circulation mentioned in section 3, which also involves the SACZ and anomalies of the opposite sense over the subtropical plains. Interannual and intraseasonal variability over subtropical South America can therefore be an expression of rectified modulation of intrinsic intraseasonal modes of atmospheric variability by remote forcing (Grimm *et al.* 2004).

Observational evidence indicates the springtime snowpack modulates the amplitude of the NAMS (Ropelewski *et al.* 2004). No similar relationship involving the snowpack has been proposed, to our knowledge, for the SAMS.

5. A Discussion on the Dynamics of the NAMS and SAMS

The distributions of continental masses, orography and SSTs combine to define the characteristics of the American monsoon systems. The dynamical balances that characterize NAMS and SAMS share key similarities at the continental scale, as well as some important differences. Over both North and South America, the summertime upper-level anticyclone/low-level heat low that characterizes a monsoon circulation is in spatial quadrature in longitude with ascent on the eastern side of the continent and subsidence on the western side (Chen 2003). This configuration of phase allows for a largely Sverdrup-type balance between the vorticity source associated with diabatically-forced continental-scale vertical motion, and advection of planetary vorticity. (The balance differs from that in the larger-scale Asian SW monsoon in a somewhat stronger contribution by zonal vorticity advection.) On their eastern flanks, both the NAMS and SAMS are associated with large-scale subtropical convergence zones (STCZs). Here, the low-level poleward motion associated with the Sverdrup balance feeds warm moist air into the convergence zone in a positive feedback. In the case of NAMS, the STCZ emanates from the eastern Pacific ITCZ across Central America and the Caribbean, extending northeastward along the eastern seaboard of North America. During the SAMS, the STCZ comprises convection over Amazonia and the SACZ. Thus, while NAMS is associated with convection over warm SSTs, i.e. the western hemisphere warm pool, the SAMS convection is largely over land. The latter difference is more formal than substantial since the Amazon is sometimes referred to as an "inland sea," in connection with its atmospheric impact.

On their western flanks, both NAMS and SAMS are associated with subsidence over the cool waters of the eastern Pacific and their accompanying stratocumulus decks. Here Sverdrup balance dictates equatorward winds and hence surface wind stresses that will lead to Ekman pumping of cool waters to the surface. The highest incidence of California stratocumulus clouds peaks in the warm season. However, the highest incidence of stratocumulus clouds off the coast of Perú-Ecuador occurs in the southern spring, indicating that the monsoon is not the only mechanism that determines their seasonal cycle. The upper-level anticyclones over North and South America develop at different distances from the equator. The low-level jet (LLJ) over the Gulf of California lies near 30N, while the South American LLJ is situated near 15S. Given the importance of the planetary-scale zonal temperature gradient in setting up the monsoonal divergent circulation, the distribution of SSTs in the eastern Pacific, with cold water off the California coast and warm water to the south, would appear to constrain NAMS to be situated further poleward than SAMS.

Baroclinic Rossby wave dynamics is largely responsible for setting up the planetary-scale features of the American monsoonal circulations presented above. Rossby waves generated by deep convection generate an area of subsidence westward and poleward of their source. The term "interactive Rodwell-Hoskins mechanism" (IRH) coined by J. D. Neelin and collaborators describes the way Rossby wave-induced subsidence to the west of monsoonal heating interacts with the convergence zone. The adjective "interactive" in IRH stresses that the spatial pattern of the monsoon heating itself is determined interactively with the subsidence regions. The subsidence itself interacts with the mid-latitude westerlies, descending adiabatically and equatorward down the sloping isentropes, with west-coast orography tending to localize the region of descent. An application of these arguments to NAMS and SAMS is consistent with Rossby wave descent over the eastern subtropical Pacific, where a persistent stratocumulus deck develops.

Returning to the issue of meridional extent of the American monsoons, the alignment of the continents could potentially allow both NAMS and SAMS to extend far poleward, and there is evidence that the STCZs on the eastern side of the continents extend into the midlatitudes in both cases. The monsoons extend poleward until the midlatitudes dynamical regime takes over, in which horizontal temperature advection by the westerlies is able to balance surface heat flux.

In addition to the planetary-scale wave dynamics discussed above, both American monsoon systems have smaller-scale features embedded within them, again with baroclinic Rossby wave dynamics. Within NAMS, there is an inverse relationship between precipitation in the core NAMS region/southeastern U.S. and that over the Great Plains of North America. Similarly, within SAMS the activity of the SACZ and precipitation over southeastern South America are inversely correlated. In both cases, to balance a stronger zone of regional convergence requires a regional strengthening of the upper-level anticyclone/low-level heat low structure that is then superposed on the planetary scale monsoon circulation. The associated scale interactions remain largely unexplored, but Fig. 4 presents an example of their importance.

6. Predictability

On intraseasonal time scales, accurate forecasts of MJO activity could be expected to lead significant improvements in the skill of warm season precipitation forecasts. The improvement for NAMS was suggested by Higgins *et al.* (2000), while that for SAMS results from the apparent MJO-type variability found in the SACZ (Nogués-Paegle *et al.* 2002). On seasonal to interannual time scales, the potential for prediction resides in the possible effects on the atmosphere of slowly-varying surface conditions, such as those in the oceans or land.

Predictability studies with global models generally indicate very modest levels of seasonal-mean precipitation skill over the NAMS and SAMS domains. Hindcast experiments have been carried out in both a two-tier context, in which the SST is predicted first, as well as in fully coupled GCMs. These predictions are made with ensembles of GCM runs, and expressed probabilistically, typically in terms of the ranked probability skill score (RPSS) of tercile-categorical predictions.

An evaluation has been made of IRI's real-time 1-month lead forecasts since their inception in 1997 (over the 1997-2001 interval; Goddard *et al.* 2003). Over this short period, the January-March seasonal-average precipitation predictions contain useful skill over the equatorial part of the SAMS domain, as well as over the subtropical plains, near 30°S. The skill is very low at intermediate latitudes. Over the NAMS region during July-September, the IRI's real-time forecasts show some skill over northwest Mexico, but skill levels are generally lower than for the SAMS.

The reasons for the low hindcast skills, especially for the NAMS, can be attributed to the weak impact of ENSO described in section 4. The higher predictability in the north and south of the SAMS domain can also be attributed to ENSO; the spatial distribution of precipitation probabilities associated with ENSO show a similar pattern, for reasons discussed in section 4. However, the IRI seasonal forecasts do not use an initialization of land surface conditions, which may in certain situations lead to useful seasonal predictive skill.

Another reason for the low predictability of the American monsoon systems can be attributed to the importance discussed in section 4 of land atmosphere-land interactions. Increased surface heating by insolation increases towards the end of the dry season weaken the static stability of the overlying atmosphere and contribute to set up favorable conditions for the onset of the wet season. There are other contributors, however. On the one hand, regional soil conditions influence the intensity of surface warming. On the other hand, remote climate anomalies such as in SST influence conditions in the atmosphere. Since NAMS and SAMS regions appear to be marginally sensitive to oceanic anomalies, one could argue that regional variations in surface conditions are more important to the onset of the wet season than those in remote SST. It is well know that the successful simulation of land surface processes is one of the major current challenges for numerical modeling of climate variability.

7. Final Remarks

Both the NAMS and SAMS comprise an upper-level anticyclone/low-level heat low structure; large-scale convergence zones with ascent to the east and descent to the west over the ocean where stratocumulus clouds enhanced by subsidence and upwelling develop. The distribution of continental masses, orography and SSTs contribute to define the characteristics of the monsoon systems. A major difference between the NAMS and SAMS is that the former is farther away from the equator than the latter. The intraseasonal (interannual and even interdecadal) variations of the NAMS and SAMS appear to be associated with continental-scale modes in which stronger precipitation in the core monsoon regions is associated with drier conditions to the northeast. The relative roles of internal atmospheric dynamics remote forcing (particularly SST) local and regional land surface forcing in the development, maintenance and decay of NAMS and SAMS are a matter of current debate. Consistently with the marginal influence of slowly varying surface conditions reported so far, predictability of the NAMS and SAMS variations is low.

Internationally organized research on the American Monsoons has been accelerated in recent years by the WCRP/CLIVAR on the Variability of American Monsoon Systems (VAMOS). Specifically, VAMOS has encouraged the realization of the SALLJ experiment (SALLJEX) in 2003 and the North American Monsoon Experiment (NAME) in 2004. One major goal of VAMOS and its projects is to increase the prediction skill for warm season rainfall over the Americas. In order to achieve this goal it is necessary to improve the observing system over North and South America, particularly over the latter. In addition, prediction systems must be improved in many respects, including improved models able to produce a better simulation of the diurnal cycle and land surface processes. Last, but not least, multinational scientific collaboration and coordination have to be strengthened across the Americas.

References

Chen, T.-C., 2003: Maintenance of summer monsoon circulations: A planetary scale perspective. *J. Climate*, **16**, 2022-20037.

Grimm A. M., C. Vera, and C. R. Mechoso, 2004: The South American Monsoon Systems. This volume.

Herdies, D. L., A. Da Silva, M. A. Silva Dias, and R. Nieto-Ferreira, 2002: Moisture budget of the bimodal pattern of the summer circulation over South America. *J. Geophys. Res*., **107**, 42/1-42/10.

Higgins, W. and co-Authors, 2003a: Progress in Pan American CLIVAR Research: The North American Monsoon System. *Atmosfera*, **16**, 29-65.

Nogués-Paegle, J., C. R. Mechoso, and coauthors, 2002: Progress in Pan American CLIVAR Research: Understanding the South American Monsoon. *Meteorologica*, **27**, 3-30.

Ramage, C. S., 1971: Monsoon Meteorology. Academic Press, 295 pp.

Robertson, A. W., C. R. Mechoso and Y. -J. Kim, 2000: The influence of Atlantic sea surface temperature anomalies on the North Atlantic Oscillation. *J. Climate*, **13**, 122-138.

Ropelewski, C. F., D. Gutzler, E. W. Higgins, and C. R. Mechoso, 2004: The North American Monsoon Systems. This volume.

Vera, C.,W. Higgins, and co-Authors, 2004: A Unified View of the American Monsoons. *J. Climate*. Submitted.

THE NORTH AMERICAN MONSOON SYSTEM

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Abstract

This paper examines major features of the North American Monsoon System (NAMS). The summertime circulation shows upper-level anticyclone/low-level heat low structures. In a companion papers Mechoso *et al.*, (2004) compare the North and South American monsoons and to Grimm *et al.*, (2004) discuss the South American monsoon. This paper reviews recent research on the NAMS and summarizes hypothesized mechanisms associated with NAMS rainfall and sources of moisture for this rainfall. No single rainfall mechanism is seen to be dominant. Numerical model simulations indicate modest potential predictability when forced by observed sea surface temperature. We suggest that the large scale boundary forcing provides some degree of "conditioning" to all or the majority of these mechanisms. From this perspective the large-scale boundary forcing associated with SST anomalies can provide some modest shifts to the probability distribution of the occurrence or effectiveness of the phenomena related to NAMS rainfall.

1. Introduction

The North American Monsoon System (NAMS) contains all of the elements of the much larger and stronger Asian Monsoon system, Mechoso *et al.* (2004), but on a smaller scale. Other major seasonal circulation features, primarily the evolution of the Bermuda high and its westward extension, also vie for importance in shaping the character of summer precipitation in the North America. Nonetheless the NAMS is a critical feature of the climate in western Mexico and parts of the southwestern United States. The monsoon accounts for at least 50% of the summer rainfall for much of western Mexico from near 20 N through the states of Nayarit, Sinaloa and Sanora, Adams and Comrie (1997), and nearly 40% of seasonal rainfall in southern Arizona and New Mexico in the United States, Fig 1. Some locations in Mexico receive as much as 70% or their annual rainfall associated with the NAMS during July, August and September (e.g. Douglas et al 1993).

While the bulk of the monsoon rains falls from July through September agriculture in southwestern Mexico is sensitive to the start of the monsoon rains in June. The NAMS region has experienced a dramatic population growth over the past 20 years along with an increase in manufacturing and land under cultivation making the entire area more sensitive to year-to-year variations in monsoon rainfall and more vulnerable to monsoon failures. Variability in the NAMS rains have also been related to variations in the threat of wildfires as well as being a factor in public health including Valley Fever and more recently dengue, Ray *et al.* (2003). Thus NAMS, like its larger and stronger counterparts, has a profound impact over a substantial population influenced by its evolution.

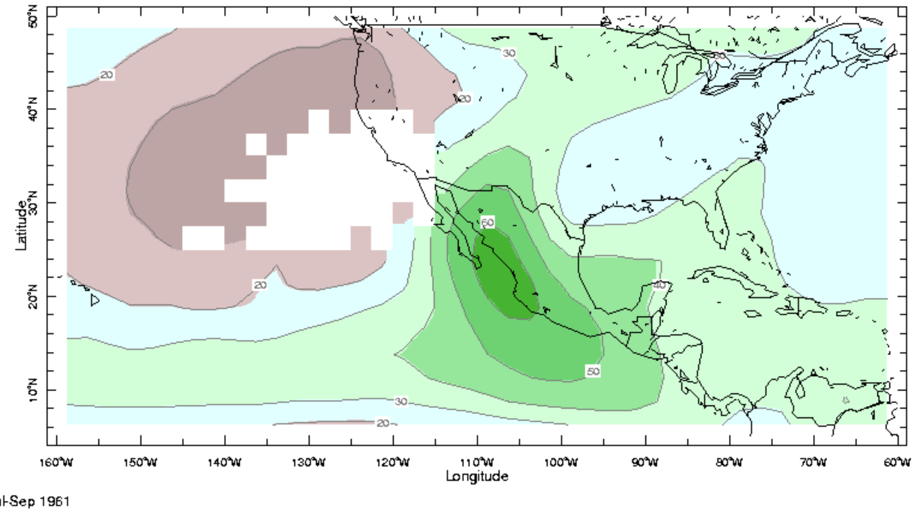

Fig 1. Percent of annual rainfall falling in the heart of the North American Monsoon season (July, August, September). The 50 % level is shown in dark green, CAMS_OPI data, Xie and Arkin (1997).

2. Characteristics of the North American Monsoon System (NAMS)

The NAMS shares many of the characteristics of its larger and more powerful monsoonal siblings in Asia and Africa and its Western Hemisphere counterpart in South America. On each of these continents the monsoon rainfall occurs during the summer season, has a distinct onset and a less distinct demise, and is accompanied by significant changes in large-scale circulation fields. While the NAMS rains are not as impressive as those associated with the monsoons in Africa and Asia (accounting for only 50% to 70% of the annual total rainfall compared to large areas that experience 80% of the seasonal totals in the latter regions) the NAMS rains are indeed important. Some earlier monsoon studies (e.g. Ramage 1971) did not include The NAMS and its South American counterpart (see Grimm *et al.*, 2004) have many circulation components in common with the Asian and African monsoon systems and most researchers include the North and South American monsoons in their definitions of the phenomenon. Most notably each of the monsoon areas are, to first approximation, a manifestation of the reversals of low level temperature gradients generally associated with the seasonal shifts in insolation i.e. for the most part they are a result of the changes in differential heating between a land mass and adjacent oceans associated with the changes in seasons. The concomitant development of a surface low-pressure system is also accompanied by the development of an upper level monsoon anticyclone and seasonal summer rainfall.

While there are numerous similarities among Earth's monsoonal climate regimes, plate tectonics have provided climate scientists with a set of four very different land-sea configurations, which can be viewed as interesting natural experiments, to help develop our understanding of monsoon systems. Both the African and Asian Monsoon systems have water to the south and land to the north, with no significant elevated heat source (mountains) in the former case and the largest mountains on earth in the latter. In the Western Hemisphere monsoon systems the mountains are oriented primarily north-south. Moisture for the South American monsoon has its origins in easterly trade wind circulations rooted in the Atlantic Ocean (Grimm *et al.* 2004). The oceanic pole of the North American monsoon circulation is by comparison extremely complex, with important roles played by the Gulf of Mexico, Pacific Ocean, and much smaller Gulf of California. This complexity is associated with an unresolved debate over the oceanic sources of atmospheric moisture for the NAMS, as discussed below. Thus study of the similarities and differences among the monsoon systems can provide clues to the physical mechanisms associated with monsoon generation, demise and interannual variability.

A. Evolution of the NAMS Mean Circulation Features

1.) The Upper Level Circulation

The evolution of the NAMS atmospheric circulation patterns is influenced by the complex topography over the region, Fig 2. Prominent geographic features include: the Gulf of California, defined to the west by the Baja California peninsula, and to the east by the Mexican mainland and Sierra Madre mountains which rise to above 2000m throughout much of central Mexico. The western slopes of the Sierra Madre Occidental effectively channel and confine low-level flow from the south-southeast up the Gulf of California towards the north-northwest into the southwest United States. The eastern mountain slopes (Sierra Madre Oriental) provide a barrier to low-level flow and direct low level moisture transport from the Gulf of Mexico into the monsoon region. The high topography of central Mexico also promotes vertical transport of moisture into the middle troposphere via the deep convective mixing associated with orographically triggered thunderstorms.

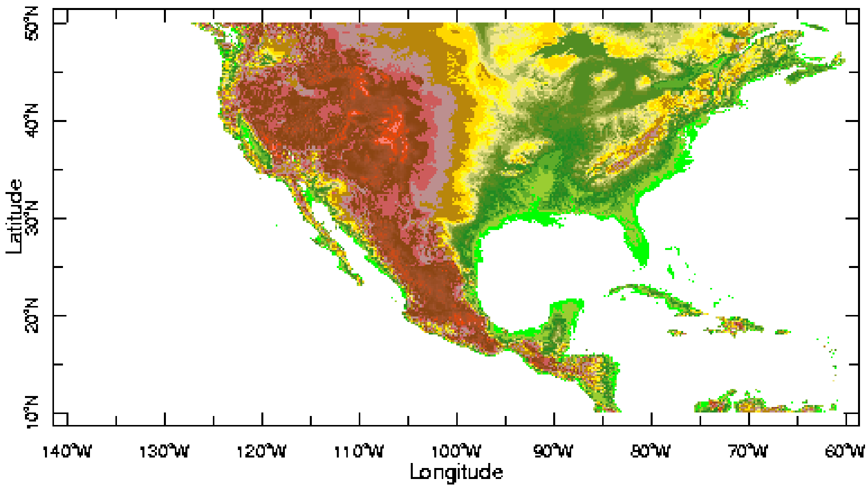

Fig.2 Digital elevation map for the North American Monsoon system region. Based on data from the GLOBE data set, Hastings and Dunbar (1998).

Because of the complex terrain over the land areas under the influence of the NAMS the evolution of the monsoon is most evident in the 200-hPa circulation. Prior to the start of the monsoon the flow is primarily zonal. During late May to early June the zonal flow begins to evolve with the formation of a monsoon-like anticyclone, centered near 15°N just to the south of Mexico, Fig 3a. Even at these early stages the 200-hPa ridge axis, though weak, extends northward into Canada. By July the mean position of the anticyclone migrates north and is centered over the Sonora desert near the border of Mexico and the United States, Fig 3b. In a 500 hPa composite analysis Higgins et al (1997) suggest that the formation of a downstream trough to the lee of the Rocky Mountains and extending into the central United States is a part of the NAMS. During August the monsoon anticyclone continues to strengthen, in the mean, with a stronger ridge to the north and a suggestion of upper-level confluent flow northward to the Canadian border, Fig 3c. By August the mean 200-hPa anticyclone dominates the climatological circulation pattern from the Pacific Coast eastward through the Gulf of Mexico. In September the 200-hPa anticyclone moves southward and diminishes in size signaling the demise of the monsoon, Fig 3d. By October (not shown) the mean flow becomes essentially zonal over the NAMS region and remains so until the beginning of the next monsoon season. While these features are quite evident in the mean 200-hPa circulation, the evolution of the NAMS for a particular individual year shows considerable interannual variability.

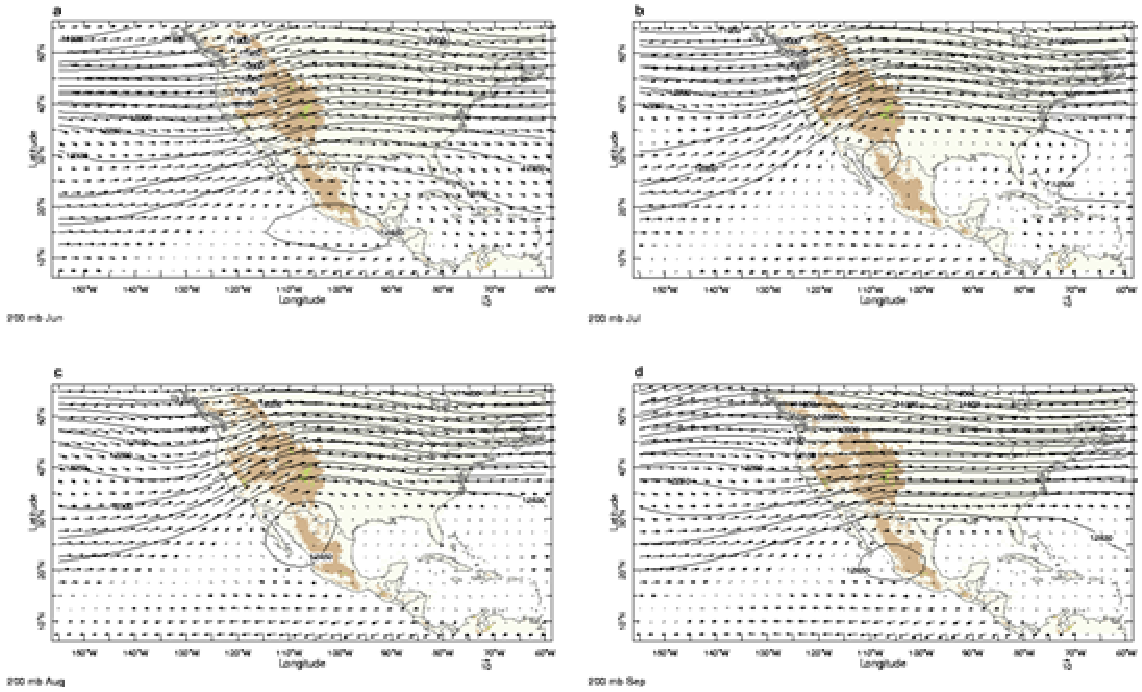

Figure 3. The mean 200 hPa vector winds (mps) and geopotential heights (gpm), 1961-1990 averages based on the NCEP/NCAR Reanalysis, Kalnay *et al.*, 1996. a) June, b) July, c) August and d) September. Topographic features above 2000m are indicated by shading.

2.) Mean Low-level Circulation

As mentioned above, Baja California to the west, the Sierra Madres to the east and the Rockies, the Sierra Nevadas and Colorado Plateau to the north complicate the low-level circulation patterns over much of the NAMS region (see Fig. 2). This string of mountain ranges and the high plateau form barriers to most circulation features below 700 hPa where most of the atmospheric moisture resides. This has led to some debate, summarized in Adams and Comrie (1997), as to the source of the moisture for the NAMS rainfall. We will return to this question below. In this section we simply describe the mean circulation features of the lower atmosphere.

The area of the Pacific Ocean immediately to the west of the NAMS region is generally characterized by high pressure and anticyclonic circulation that form part of the mean summer subtropical high pressure belt. Sinking motion to the west of the NAMS precipitation may enhance the strength of the high pressure as discussed in Mechoso *et al.* (2004). Northerly winds on the eastern flank of the anticyclonic mid-latitude circulation are associated with coastal upwelling along western North America. The northerly flow extends to Baja California through the Northern Hemisphere winter but diminishes in strength during the spring and summer. By June the mean northerlies are well off the coast and the low level mean circulation becomes relatively weak along the Mexican coast for the remainder of the summer, Fig 4a through 4d. Conversely, in the Gulf of Mexico strong low-level easterly and southeasterly flow develops during the spring, feeding moisture into the low level jet over the Great Plains to the lee of the Rockies, Fig 4a. During the late spring and early summer a relative low pressure area develops under the 200-hPa anticyclone discussed above. This is manifest in the NCEP/NCAR Reanalysis pressure fields by a col over the Sonora Desert (not shown) and by light and variable winds at 850 hPa, Fig 4a.

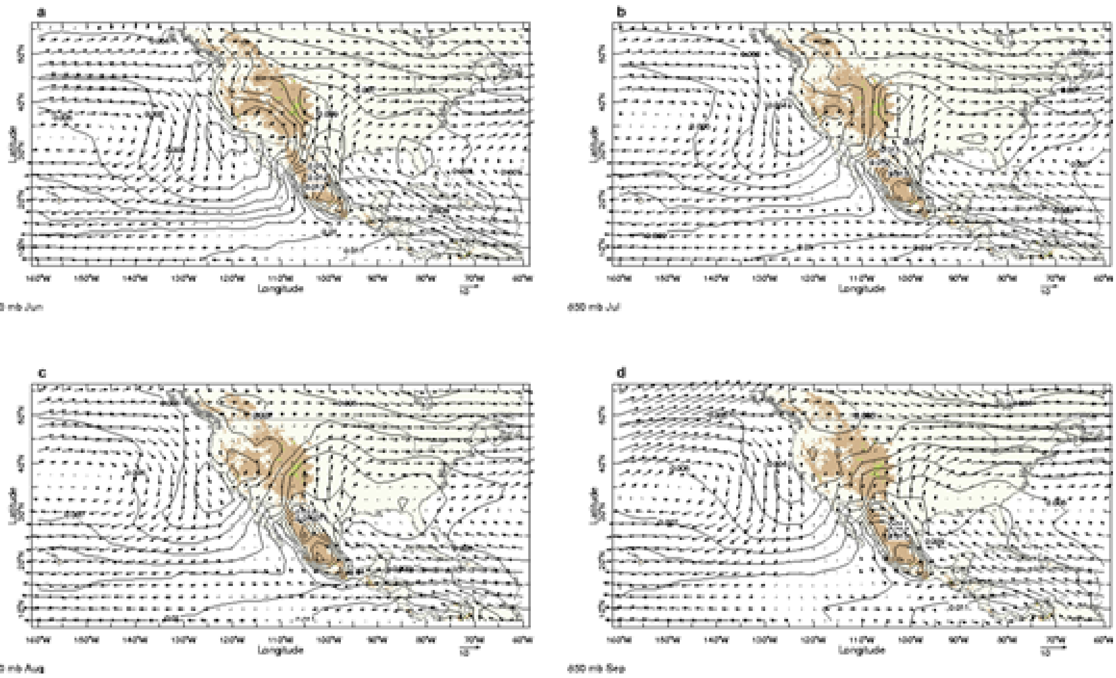

Figure 4. The mean 850 hPa vector winds (mps) and specific humidity (gm/kg), 1961-1990, based on the NCEP/NCAR Reanalysis, Kalnay *et al.*, 1996. a) June, b) July, c) August and d) September. Topographic features above 2000m are indicated by shading.

The mean 850-hPa wind and moisture fields illustrate some of the fundamental challenges in understanding the moisture transport mechanisms associated with NAMS rainfall. The strongest low level winds and by implication the strongest low level moisture transports are to the east of the main NAMS monsoon rain areas. The mean low level winds are strongest in regions of relatively weak moisture gradients i.e. to the east of the Sierra Madre mountains and, in contrast, strong low level moisture gradients appear along the west coast of Mexico in a region where the NCEP/NCAR Reanalysis shows the mean wind fields to be near zero during each month of the monsoon season, Figs 4a-d. If the moisture for monsoon rainfall comes from the Pacific this suggests that the NAMS precipitation results primarily from transients. (We note that some analyses suggest a strong mean low level southerly jet extending along the entire length of the Gulf of California but there is not a strong observational consensus with respect to the detailed spatial extent and temporal variability of this low level jet. This feature is the subject of ongoing research.)

A close look at the topography, moisture gradients and winds in northern Mexico lends support to the view that some portion of the NAMS precipitation may be fed by moisture from the east. (Note, however, that only a relatively small area is below 2000m along the Mexico-United States border in Figs 4a-d). Schmitz and Mullen (1996) using ECMWF analyses and Higgins *et al.* (1997) using NCEP/NCAR Reanalysis suggest that moisture enters into the NAMS region from the east i.e., from the Gulf of Mexico above 850 hPa. On the other hand, given the sparseness of input data and the scale of the NCEP/NCAR and ECMWF analyses the details of the wind and moisture fields may be somewhat uncertain. Current mesoscale modeling efforts are expected to help better define the source regions of the moisture needed to sustain the monsoon rains.

B. Mean Evolution of NAMS Monsoon Rainfall

The bulk of the summer (July to September) rainfall in the region falls over the ocean with maximum amounts centered near 10°N and 105°W, Yu and Wallace (2000), Higgins and Shi (2001). This rainfall is associated with the northward progression of the Inter-Tropical Convergence Zone (ITCZ) and can be viewed in that context. However, an argument can be made that the contrast between the relatively high sea surface temperatures (SSTs), coinciding with the areas of maximum rainfall, and the relatively cold waters to the south, constitute a “classic” thermally direct monsoon system, but with a major component over water. Presumably the associated subsidence region(s) would be on the northern flank of precipitation maximum, over the stratus decks off of the California coast and/or on the southern flank of the precipitation maximum off of the west coast of South America. Most studies of NAMS precipitation concentrate on the precipitation that falls over the land areas of North America while acknowledging that the continental rainfall is a relatively small portion of the total rainfall regime in the NAMS. The description below follows this convention.

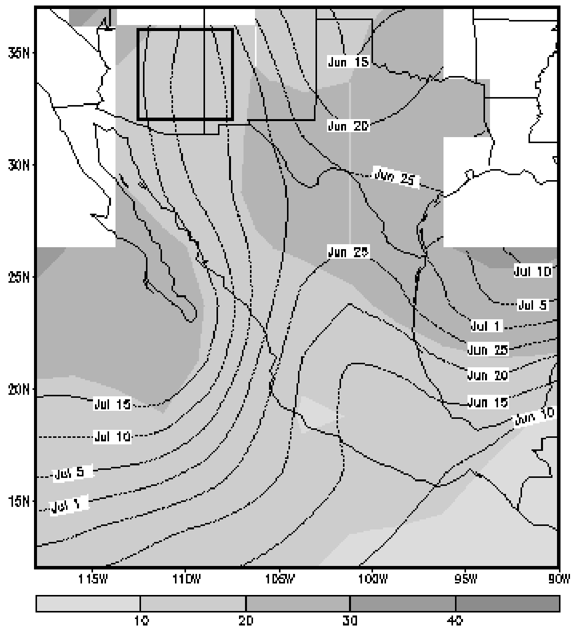

Figure 5. Mean North American Monsoon onset date based on 5-day satellite estimated rainfall, from Janowiak and Xie (2003). The shading represents the standard deviation of the onset dates in days.

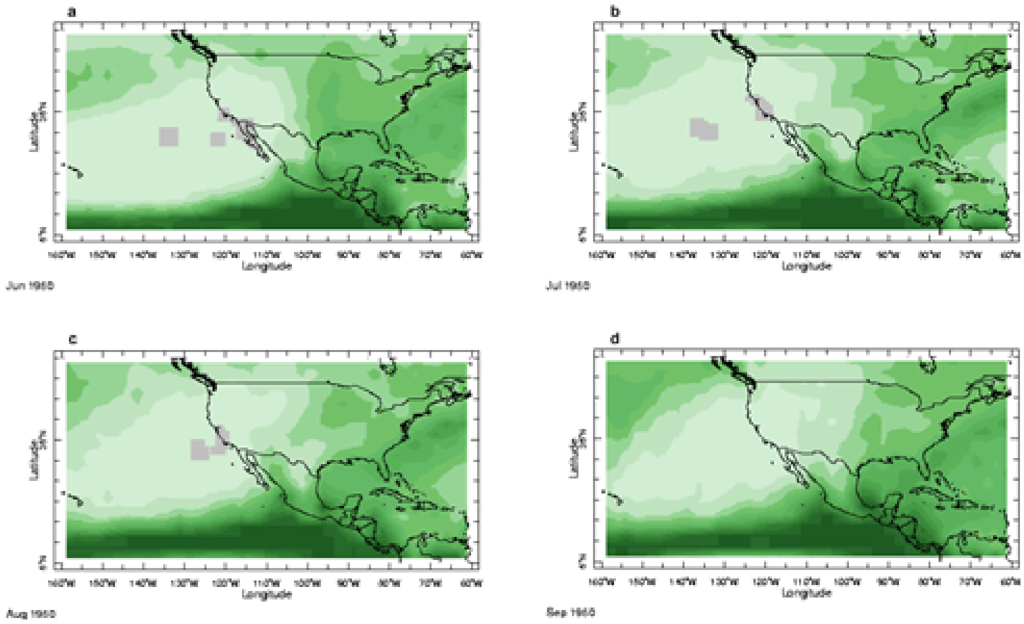

Figure 6. Evolution of the North American Monsoon rainfall based on CMAP, Janowiak and Xie (1999) (Mike Bell's version) estimated range 0 to 10 mm/day, 0 to 300mm/month.

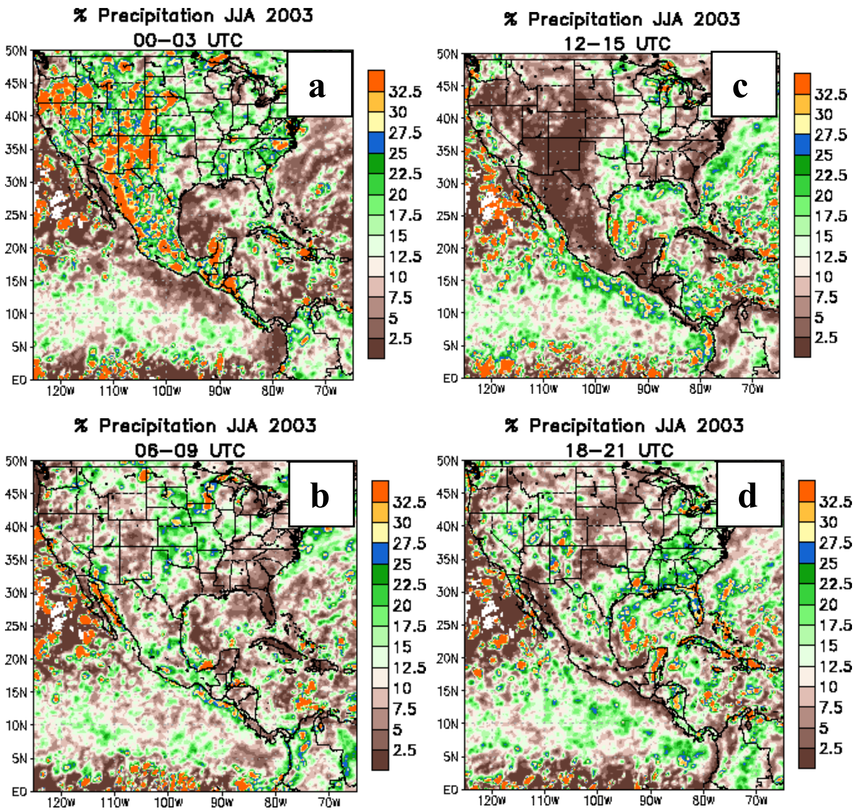

Figure 7. Mean Diurnal cycle of rainfall June to August 2003, Joyce *et al*. 2004.

By mid-to-late June the continental monsoon rain is generally evident near 20°N in Nayarit State in Mexico and proceeds fairly rapidly to the north. Satellite-based precipitation analysis suggests that once underway the monsoon precipitation progresses from the east along the Sierra Madre Occidental, towards the west, Fig 5, Janowiak and Xie (2003). The same analysis suggests that the monsoon rainy season has about a 100-day duration, lasting essentially from late June through September, over much of the core monsoon region. The June monsoon rainfall is relatively modest even in Mexico, Fig 6a, but by July the rainfall has lined up from west central Mexico northward into the United States, Fig 6b. August is the rainiest monsoon month over much of the region, Fig 6c, but by September the rains have substantially retreated to the south, Fig 6d. Generally drier conditions return to the region by October (not shown). Barlow et al (1998) note that the rapid June to July increase in monsoon rainfall is coincident with dramatic decreases in the U.S. central plains. This suggests that the influence of the NAMS may extend into the central U.S., far beyond the areas directly involved in evolution of the monsoon rainfall.

The monsoon rainfall has a large diurnal component over the Sierra Madre Occidental, Fig 7. Satellite estimates suggests that rainfall amounts peak in the mountains in late afternoon with the time of maximum rainfall becoming later to the west and a tendency for late evening to near midnight local maxima for locations on the Gulf of California. The diurnal variability is not completely understood and is likely related to the rainfall mechanisms discussed in the following section.

C. Rainfall Mechanisms

Studies of the continental monsoon rainfall have suggested several dynamical mechanisms that may modulate rainfall, including pressure “surges”, easterly waves, tropical storms, and intra-seasonal variability associated with the Madden-Julian Oscillation (MJO). None of these mechanisms appear, by itself, to explain the bulk of the monsoon rainfall and its variability. There does appear to be an emerging consensus that rainfall over the continental portions of the monsoon regions is associated with transients rather than the mean flow. Berbery (2001) analyzed the Eta Model’s seasonal mean and transient moisture flux at 950 hPa and suggested that the mean flow may actually be northerly along the southern half of the Gulf of California while the transients, though weaker, show northward transports of moisture flux into the monsoon rainfall regions. If the transients are the primary NAMS rainfall producers, the relative climatological contributions of various transient phenomena (Gulf surges, easterly waves, tropical storms, intraseasonal variability like the MJO) are still not clear.

Tropical storms are one transient rainfall phenomenon that certainly brings abundant rainfall to the region (Englehart and Douglas 2001). There is no doubt that substantial amounts of summer monsoon rainfall in western Mexico and into the southwest United States is associated with tropical storms during some years. However, these storms are episodic and do not affect the entire NAMS region during every monsoon season.

Gulf surges are pulses of southerly winds that transport moisture up the Gulf of California. While the surge phenomenon is well documented (Stensrud *et al.* 1997). The link between Gulf surges and rainfall, particularly along the northern extent of the monsoon is not strong e.g.Mechoso *et al.*, (2004). It has been hypothesized that the surges may be related to the occurrence of easterly waves propagating across the Gulf of Mexico. Fuller and Stensrud (2000) further suggest that the easterly waves are most effective in producing surges if they are properly in phase with the passage of mid-latitude troughs. While their study suggests that this mechanism may account for some of the monsoon rainfall, the required phasing of all of the elements suggests that it can’t be the sole, or even primary, mechanism.

Thus there is no strong evidence for the dominance of any of these transient phenomena in producing NAMS rainfall. It is possible that there is no overall dominant transient rainfall mechanism but that each mechanism contributes to the rainfall and the relative contributions for any particular season is random.

As mentioned above the continental component of the monsoon rains is a relatively small fraction of the total rainfall associated with this phenomenon. Studies that include the portion of the NAMS rainfall over oceans suggest that the both of El Nino/Southern Oscillation and the Madden-Julian Oscillation (MJO) i.e., intraseasonal variability may influence rainfall over the NAMS region. Higgins and Shi (2001) show correlations between seasonal monsoon rainfall and ENSO as well as with intraseasonal variability. On the other hand, in an analysis based on monthly data, Yu and Wallace (2000) could find no strong link between the oceanic component of the rainfall and ENSO (the El Nino/Southern Oscillation) phenomenon except for a tendency towards broadening of the principal rainfall band during warm, El Nino, conditions and contraction or sharpening of that band during cold, La Nina, conditions. To the extent that ENSO has some influence over the NAMS precipitation, its influence appears to be very modest over continental regions, confined to the southern portion of the NAMS domain (Gutzler 2004).

3. Modeling Studies

As mentioned above, the North American Monsoon System (NAMS) is modest in the extent and amplitude of its precipitation maximum, and exhibits a much less pronounced seasonal wind reversal, compared to Earth's other monsoons. Nevertheless the seasonal evolution of the NAMS serves to organize warm season precipitation across the entire North American continent including areas not usually associated with the core monsoon development such as the U. S. Great Plains. The continental geometry of North America, featuring distinct oceanic moisture sources both to the east and west of the monsoon precipitation maximum with steep, high topography in between, forces extremely sharp gradients of precipitation around the periphery of the NAMS domain. The Gulf of Mexico on the eastern side of the domain feeds moisture both into the highly elevated southwestern region of the North American continent (the monsoonal domain), as well northward into the heartland of the United States, where the continent is quite flat and the seasonal precipitation maximum occurs in springtime.

Capturing these climatological features poses an extreme challenge for dynamical models. Proper simulation of the warm season precipitation regime across North America must:

1) Include a realistic description of the seasonal evolution and spatial distribution of precipitation in the core of the monsoon region in northwestern Mexico. The presence of high topography very close to the coastline tends to generate circulations with an extremely high-amplitude diurnal cycle.

2) Capture the observed evolution of continental scale features around the periphery of the NAMS domain. These include the diminution of moisture transport and precipitation in the central U.S. (associated with shifts in the Atlantic subtropical High), and the shifting and strengthening of the Pacific subtropical High, as the NAMS ramps up.

3) Reproduce the proper linkages to synoptic and large-scale tropical features to the south and east of the monsoon domain, including interannual shifts in the amplitude and position of the Pacific ITCZ, and synoptic-scale circulation anomalies such as the Madden-Julian Oscillation (MJO), easterly waves and tropical cyclones.

Regarding point (1), the sharp spatial gradients in precipitation associated with the NAMS have proven especially difficult to model (Gutzler *et al.* 2004). In the core of the monsoon region, the summer precipitation maximum occurs along the slopes of the Sierra Madre Occidental, on the eastern side of the Sea of Cortes (the Gulf of California). The interaction of steep orography, diurnal land-ocean circulations, and atmospheric buoyancy is quite complicated and generally not well simulated by existing treatments of deep convection in models. On the western side of the Gulf of California, not much more than 100 km away, summer precipitation is very sparse, the Pacific subtropical High dominates the circulation, and most precipitation falls in winter. Simulated precipitation rates in the core of the NAMS domain have been shown to be sensitive to choice of convective parameterization and boundary layer treatment in atmospheric models, as well as being very sensitive to both land surface treatment and SST in the Gulf of California (e.g. Mitchell *et al.* 2002; Gochis *et al.* 2002; Matsui *et al.* 2003; Kanamitsu and Mo 2003)

Large-scale models have demonstrated some fidelity in simulating the continental features listed in point (2), but these features are sensitive to the precipitation rates (hence tropospheric heating) in the monsoon core. Improvements in continental-scale simulation, and thereby dynamical seasonal-interannual climate prediction, would therefore seem to hinge on getting the core monsoon precipitation modeled properly.

With regard to point (3), model improvements and empirical research must be linked, as the relationships between tropical ocean anomalies and the NAMS are still being explored in observations. Robust links between interannual ENSO-related anomalies (and associated meridional shifts in the eastern Pacific ITCZ) and the NAMS are still elusive, as discussed further in the following section. Low frequency intraseasonal variability in the NAMS is very pronounced; some of this variability has tentatively been identified with coherent tropical synoptic variability associated with either the MJO (propagating eastward) and/or easterly waves (propagating westward). Recent empirical studies suggest that tropical cyclones imprint a significant signal on coastal precipitation, especially late in the monsoon season (Englehart and Douglas 2001).

4. Predictability

Even though ENSO tends to be primary source of climate predictability over many regions of the world it does not seem to offer much to predictability over the continental regions of the NAMS domain during the summer, Yu and Wallace (2000), Higgins and Shi (2001), Gutzler (2004). Nevertheless, several numerical models forced by observed sea surface temperatures show modest correlation skill between "forecast" and observed rainfall over the NAMS domain during the July to September season. In particular, the ECHAM3.6, NCEP-MRF9, CCM3.2 and the NASA/Goddard NSIPP models each show NAMS regions with correlations in the order of 0.4 and greater for the 33-year period 1965 to 1997 (on the web, http://iri.columbia.edu/forecast/climate/skill/SkillMap.html). The sources of this modest potential predictability has not been diagnosed but the consistency among the numerical models suggest that at least a part of this predictability is real.

Empirical studies have explored possible relationships between sea surface temperatures in the Gulf of California e.g., Mitchell *et al.* (2002). In general, these studies rely on sea surface temperature analyses that resolve smaller spatial scales for shorter duration than analyses that are typically available for global models, and thus are limited to a few samples or case studies.

Empirical studies have also suggested that some predictability over portions of the NAMS domain may be associated with winter and spring snow cover, Gutzler and Preston (1997), Gutzler (2000), Lo and Clark (2001) and Matsui *et al.*, (2003). These studies suggest some very limited predictability of NAMS summer precipitation in New Mexico and Arizona associated with snow water equivalent or proxies for this variable. However, these studies relate only to the relatively modest monsoon precipitation on the very northern boundaries on the overall NAMS precipitation regime.

In the absence of any single dominant mechanism associated with NAMS precipitation, but faced with indications of modest potential predictability from models forced by observed sea surface temperature, one might conclude that the large scale boundary forcing provides some degree of "conditioning" to all or the majority of these mechanisms. From this perspective the large-scale boundary forcing associated with SST anomalies can provide some modest shifts to the probability distribution of the occurrence or effectiveness of these phenomena related to NAMS rainfall.

5. Final Remarks

To accelerate progress on the issues outlined above, an international process study called the North American Monsoon Experiment (NAME) has been organized. NAME seeks to improve understanding and predictability of warm season precipitation fluctuations across the continent. The NAME field campaign is taking place in summer 2004 to make enhanced observations in the heart of the NAMS across Southwest North America. A primary goal of the experiment and the field campaign is to reach better understanding of NAMS rainfall mechanisms and their predictability. In addition, in conjunction with the NAME observational campaign, a focused set of modeling activities is being undertaken to address the simulation challenges outlined above.

A series of complementary activities are also being conducted to better understand the South American Monsoon System (SAMS), e.g., see Grimm *et al.*, 2004. Coordinated study of the similarities and differences between these two monsoon systems may provide insights into the nature of monsoons that could not be obtained through study of either in isolation.

Acknowledgements

We wish to thank Mike Bell, IRI, for his assistance with many of the figures. We are also grateful to Vern Kousky, CPC/NWS/NOAA, for providing the diurnal variability figure and Hugo Berbery, University of Maryland, for the Eta-model based mean and transient flux analyses. The IRI was established as a cooperative agreement between the NOAA Office of Global Programs and Columbia University.

References

Adams, D. K and A. C. Comrie, 1997: The North American Monsoon. *Bull Amer. Met Soc*, **78**, 2197-2213.

Barlow, M, S. Nigam and E. H. Berbery, 1998: Evolution of the North American Monsoon System. *J. Climate*, **11**, 2238-2257.

Berbery, E. H., 2001: Mesoscale moisture analysis of the North American Monsoon. *J. Climate*, **14**, 121-137.

Douglas M. W., R. Maddox, K. Howard and S. Reyes, 1993: The Mexican Monsoon. *J. Climate*, **6**, 1665-1667.

Englehart, P. J., and A. V. Douglas, 2001: The role of Eastern North Pacific tropical storms in the rainfall climatology of western Mexico. Int. J. Climatology, **21**, 1357-1370.

Fuller, R. D. and D. J. Stensrud, 2000:The relationship between tropical easterly waves and surges over the Gulf of California during the North American Monsoon. *Mon. Wea. Rev*, **128**, 2983-2989.

Gochis, David J., W. J. Shuttleworth, and Z. Yang. 2002: Sensitivity of the Modeled North American Monsoon Regional Climate to Convective Parameterization**.** *Mon. Wea. Rev.* **130**, 1282–1298.

Grimm A, C Vera and C.R. Mechoso, 2004: American Monsoon: South America, (This volume)

Gutzler, D. S., 2000: Covariability of spring snowpack and summer rainfall across the Southwest United States. *J. Climate*, **13**, 4018-4027.

-----, 2004: An index of interannual precipitation variability in the core of the North American monsoon region. *J. Climate*, **17**, in press.

-----, and J. W. Preston, 1997: Evidence for a relationship between spring snow cover and summer rainfall in New Mexico. *Geophys. Res. Lett.*, **24**, 2207-2210.

-----, H.-K. Kim, R. W. Higgins, H. Juang, M. Kanamitsu, K. Mitchell, K. Mo, P. Pegion, E. Ritchie, J.-K. Schemm, S. Schubert, Y. Song, and R. Yang., 2004: The North American monsoon Model Assessment Project (NAMAP). NCEP/Climate Prediction Center ATLAS No. 11, 32 pp. (Available from Climate Prediction Center, 5200 Auth Road, Room 605, Camp Springs, MD, 20746 and online at URL:

http://www.cpc.ncep.noaa.gov/research_papers/ncep_cpc_atlas/11/atlas11.htm

Hastings, D. A. and P.K. Dunbar, 1998: Development & Assessment of the Global Land One-km Base Elevation Digital Elevation Model (GLOBE). International Society of Photogrammetry and Remote Sensing Archives, Vol.32, number 4, pp. 218-221.

Higgins, R. W., Y. Yao and X.L. Wang, 1997: Influence of the North American Monsoon System on the U.S. Summer Precipitation Regime. *J. Climate*, **10**, 2600-2622.

------------- and W. Shi, 2001: Intercomparison of the principal modes of interannual and intraseasonal variability of the North American Monsoon System. . *J. Climate*, **14**, 403-417.

------------ and co-authors, 2003: Progress in Pan American CLIVAR research: The North American Monsoon System, *Atmosfera*, **16**, 29-65.

Janowiak, J. E. and P. Xie, 2003: A Global-Scale Examination of Monsoon-Related Precipitation. *J. Climate*, **16**, 4121-4133.

------------- and P. Xie, 1999: CAMS_OPI: A Global Satellite-Rain Gauge

Merged Product for Real-Time Precipitation Monitoring Applications. J. Climate, 12, 3335-3342.

Joyce, R. J., J. E. Janowiak, P. A. Arkin and P. Xie, 2004: CMORPH: A method that produces precipitation estimates from passive microwave and infrared data at high spatial and temporal resolution. J. Hydromet., 5, 487-503.

Kalnay E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M Iredell, S. Saha, G.White, J. Wollen, Y Zhu, M. Chelliah, W. Ebisuzaki, W. Higgins, J. Janowiak, K.C. Mo, C. Ropelewski, J. Wang. A. Leetmaa, R. Reynolds, Roy Jenne and Dennis Joseph, 1996: The NCEP/NCAR 40-year reanalysis project. *Bull. Amer. Met Soc.* **77**, 437-471.

Kanamitsu M., and K. C. Mo 2003: Dynamical effect of land surface processes on summer precipitation over the southwestern United States. *J. Climate*, **16**, 496-509.

Lo, F. and M.P. Clark, 2002:Relationships between spring snow mass and summer precipitation in the Southwestern United states associated with the North American Monsoon System. . *J. Climate*, **15**, 1378-1385.

Matsui, T., V. Lakshmi and E. Small, 2003: Links between snow cover, surface skin temperature, and rainfall variability in the North American Monsoon System. J. Climate, 16, 1821-1829.

Mechoso, C. R, A. W. Robertson, C. F. Ropelewski and A. Grimm, 2004: The American Monsoon Systems (This volume)

Mitchell, D. L., D. Ivanova, R. Rabin, T. J. Brown and K. Redmond, 2002: Gulf of California sea surface temperatures and the North American Monsoon: Mechanistic Implications from observations. . *J. Climate*, 17, 2261-2281.

Ramage, C. S., 1971: *Monsoon Meteorology*, Academic Press, 296 pp

Ray, A. J., N. Schmidt, B. J. Morehouse, and R. S. Webb, 2003: Report on Research Opportunities for Climate and Society Interactions in the North American Monsoon Region. Proceedings of the Workshop on Applications and Human Dimensions of Monsoon Research 18-20 June 2001, Tucson, AZ, Institute for the Study of Planet Earth, University of Arizona, 42

Schmitz, J. T., and S. Mullen, 1996: Water vapor transport associated with the summertime North American monsoon as depicted by ECMWF analyses. *J. Climate*, **9**, 1621-1634.

Stensrud, D. J., R. L. Gall, and M. K. Norquist, 1997: Surges over the Gulf of California during the Mexican monsoon. *Mon. Wea. Rev*, **125**, 17-437.

Xie, P. and P. A. Arkin, 1997: Global precipitation: A 17-year monthly analysis based on gauge observations, satellite estimates, and numerical model outputs. *Bull. Amer. Meteor. Soc.*, **78**, 2539-2558.

______, J. E. Janowiak, P. A. Arkin, R. Adler, A. Gruber, R. Ferraro, G. J. Huffman, and S. Curtis, 2003: GPCP Pentad Precipitation Analyses: An experimental dataset based on gauge observations and satellite estimates. *J. Climate*, **16**, 2197-2214.

Yu, B., and J. M. Wallace, 2000: The principal mode of interannual variability of the North American Monsoon System. *J. Climate*, **13**, 2794-2800.

THE SOUTH AMERICAN MONSOON SYSTEM

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1. Introduction

Although the seasonal reversal of the surface winds in a “classical” monsoon regime is not apparent over South America, the warm season in the region shows features that are typical of a monsoon climate. For example, if the annual mean is removed, the surface wind does reverse in association with the strong diabatic heating in the subtropical highlands (Zhou and Lau, 1998). Also, the seasonal cycle of precipitation over most of South America is monsoon-like (Fig. 1), with great contrasts between the winter and the summer (Rao *et al.* 1996; Kousky and Ropelewski 1997; Grimm 2003; Gan *et al.* 2004). For instance, in Central-West Brazil it rains more than 1000 mm in summer (DJF) and less than 100 mm in the winter (JJA). Therefore we are justified in referring to a South American Monsoon system (SAMS). In South America, the monsoon regime prevails even in the subtropics. Therefore, understanding the SAMS and being able to forecast its variability is important in many aspects.

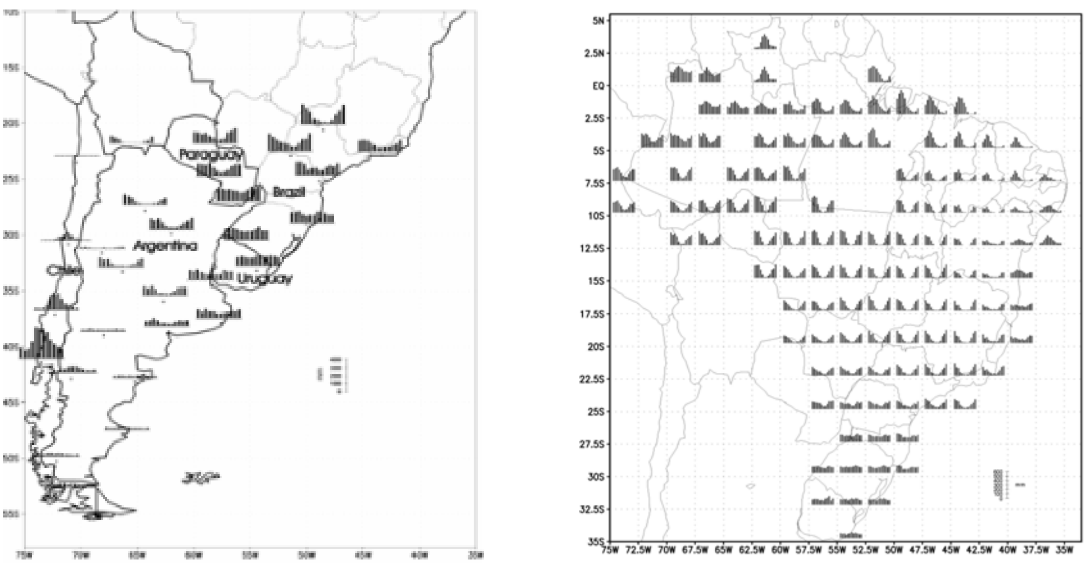


Fig. 1. Annual cycles of precipitation over Brazil (left) and southern South America (right) for the period 1956-1992 (from Grimm 2003; Grimm *et al.* 2000).

This paper examines major features of the SAMS. Section 2 presents climatological aspects, such as evolution, heating distribution and the associated circulation. Variability in all time-scales is outlined in section 3. Sections 4 and 5 review modeling studies and discuss predictability aspects. In companion papers Mechoso *et al.* (2004) compare the North and South American monsoons and Ropelewski *et al.* (2004) discuss the North American monsoon. Other relevant sources of information on the SAMS are the papers by Paegle *et al.* (2002) and Vera *et al.* (2004).

2. Climatological Aspects

a. General Features

In austral summer, as the major heating zone migrates to the subtropics, a thermal low-pressure system develops over the Chaco region, in central South America. The low pressure system over northern Argentina and western Paraguay is a climatological feature present throughout the year, but strongest during the summer.

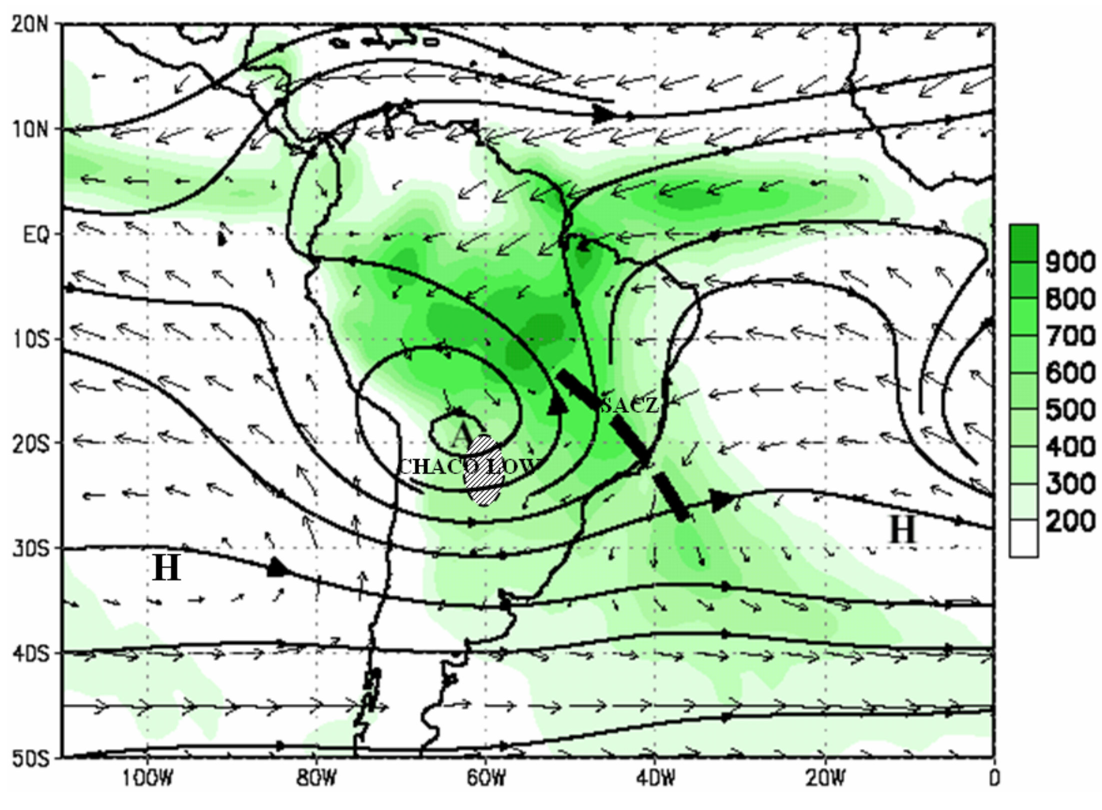


Fig. 2. Main features of the SAMS. December-February mean (1979-1995) 925 hPa vector wind and 200 hPa streamlines from the NCEP/NCAR reanalysis archive, and merged satellite estimates and station observations of precipitation (mm, shading). The position of the Bolivian High (A) and the subtropical Atlantic and Pacific surface high pressure centers (H) are indicated. The approximate axis of the South Atlantic Convergence Zone is indicated by the heavy dashed line (adapted from V. Kousky and M. Halpert).

The southwest-northeast inter-hemispheric pressure gradient between the South American low and the northwestern Sahara strengthens, enhancing the tropical northeasterly trade winds (Fig. 2). Anomalous (cross equatorial flow penetrates the continent, carrying moisture. The flow becomes northwesterly, is channeled southward by the Andes mountains, and turns clockwise around the

Chaco low. Low-level wind and moisture convergence associated with the interaction of the continental low with the South Atlantic high and the northeasterly trade winds result in enhanced precipitation in the Amazon, and Central and Southeast Brazil (Lenters and Cook, 1995) (Fig. 3). The southeastward extension of cloudiness and precipitation towards the Atlantic Ocean is referred to as South Atlantic Convergence Zone (SACZ) (Kodama 1992). As the SACZ enters its most active stage (DJF), the upper level anticyclonic center moves southward from the Amazon, setting up the "Bolivian High". East of this high and over the Atlantic Ocean close to the coast of Northeast Brazil, there is a through known as the "Nordeste trough" (Virji 1981; Kousky and Ropelewski 1997).

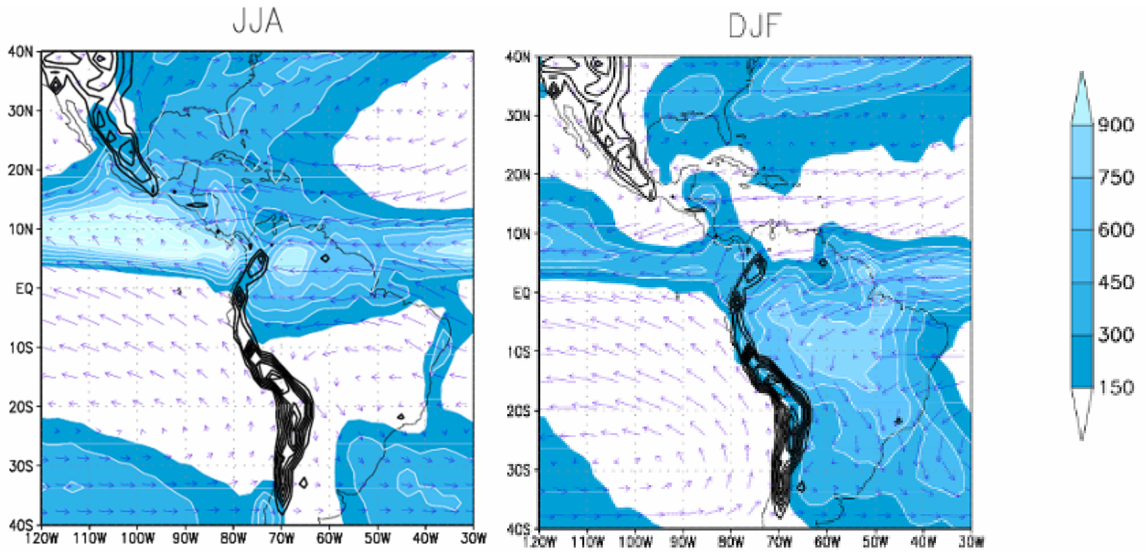


Fig. 3. Climatological mean accumulated precipitation and vertically-averaged climatological mean moisture fluxes for a) JJA and b) DJF (from Vera *et al.* 2004).

The formation of the Bolivian High has been attributed to the latent heat of thunderstorms and sensible heating over the Altiplano Plateau of Peru-Bolivia (15°-21°S) (Gutman and Schwerdtfeger 1965; Rao and Erdogan 1989), and to a Rossby wave response to the latent heat release from the convection over the Amazon basin (Silva Dias *et al.* 1987; Figueroa *et al.* 1995). Lenters and Cook (1997) argued that the Bolivian high-Nordeste trough system is generated in response to the main SAMS precipitation centers (Amazon, SACZ, and Altiplano, the Amazonian precipitation being most influential), with important participation of African precipitation in the formation of the Nordeste low. Chen *et al.* (1999) examined the Bolivian high-Nordeste trough system as a quasistationary wave regime and found that is maintained by South America local heating and remote Africa heating as well as by that over the western tropical Pacific.

Embedded within the northwesterly winds along the Andes Mountains is the South American low level jet (SALLJ). The SALLJ plays an important role in the transport of moisture from the Amazon to the subtropics, producing enhanced rainfall in its exit region. The strongest winds are near Santa Cruz, Bolivia.

The SALLJ, unlike LLJs in other parts of the world, is present throughout the year (Berbery and Barros, 2002; Marengo *et al.* 2004). The explanation for this characteristic resides in the mechanical blocking effect of the Andes orography, which causes stationary Rossby waves in the zonal circulation (Byerle and Paegle 2002; Campetella and Vera 2002). This mechanical effect tends to produce an orographically bound cyclone throughout the year, with poleward flow east of the mountains. Its variability may be partly explained by changes in the zonal circulation (Byerle and Paegle 2002). Changes in sensible and latent heating also modulate the SALLJ and are important in explaining the observed diurnal cycle (Berbery and Collini 2000).

The diurnal cycle of convective cloudiness during the summer rainy season, documented by satellite products, is tied to the diurnal march of the insolation, but also influenced by regional factors (Garreaud and Wallace 1997; Sorooshian *et al.* 2002). The peak is observed at afternoon/early evening in most of the monsoon region, consistent with the more conducive thermodynamic conditions during this part of the day, while a nocturnal precipitation maximum is observed at the subtropical plains, which might be ascribed to the diurnal cycle of the SALLJ (Berbery and Collini 2000) and the decrease of the intensity of the compensating subsidence (Silva Dias *et al.* 1987).

b. Evolution

Although the wet season in tropical South America can be initiated rapidly by synoptic systems, probabilistically the onset of the convection is controlled by changes in the thermodynamic structure, mainly related to the moistening of the planetary boundary layer and the lowering of temperature at its top (Fu *et al.* 1999; Marengo *et al.* 2001). These changes are brought about by changes in large-scale circulation that enhance low-level moisture convergence into the region, particularly a southward reversal of the cross-equatorial flow (Fig. 3) (Wang and Fu 2002; Li and Fu 2004), for the seasonal changes of surface evapotranspiration are one order of magnitude smaller than those of low-level moisture convergence. Notwithstanding, the increase of surface evapotranspiration and local water recycling are key for initiating the transition from the dry to wet seasons (Li and Fu 2004).

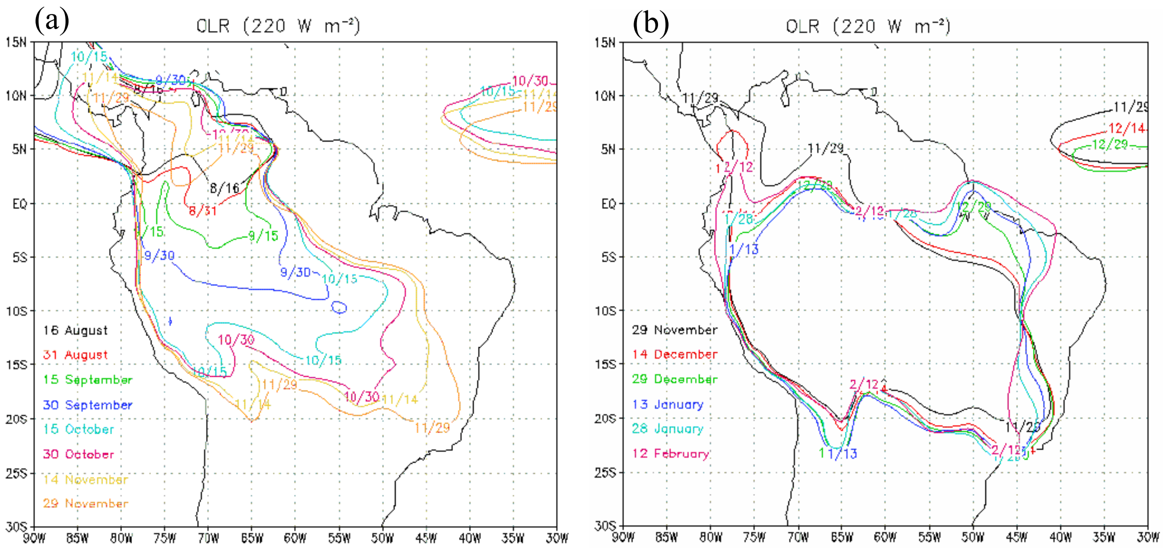
In the southern Amazon/Central Brazil the land surface warming destabilizes the lower tropospheric lapse rate from winter to spring, but a significant increase of the convection happens from October to December, when more moisture is transported into the region. The land surface warming increases the gradient of land-ocean temperature and drives the seasonal changes of circulation. On the other hand, in the equatorial Amazon, changes in the local land surface temperature are only 25-50% of those in southern Amazon. Therefore, changes of large-scale circulations are largely controlled by changes of the surface temperature in the adjacent oceans and southern Amazon (Fu *et al.* 1999, 2001; Wang and Fu 2002). Once favorable large-scale thermodynamic conditions are established, the transition to the wet season in Central Brazil may be rapid and connected to synoptic or intraseasonal variations.

During pre-monsoon season turbulent sensible heating dominates the warming of the subtropics and is confined to the lower atmosphere. This heating is maximum before mid-November (~5 K/day). When the deep convection reaches the highlands in Southeast Brazil latent heat release becomes the dominant heating component, being maximum in the middle and upper troposphere (~5 K/day) (Zhou and Lau 1998). These two heat sources are essential in shaping the climatological SAMS and influential in its year-to-year variability.

The wet season starts in the western Amazon in spring (September) and then spreads to the south and southeast (Fig. 4a), reaching southeast Brazil in October. The average onset date of the rainy season in each region depends on the criteria used for defining it (e.g., Kousky 1988, Marengo *et al.* 2001, Gan *et al.* 2004). By late November, deep convection covers most of central South America from the equator to 20°S, but is absent over the eastern Amazon Basin and Northeast Brazil. Throughout this period, deep convection associated with the ITCZ is confined to the central Atlantic between 5°N-8°N.

During the mature phase of the SAMS, from late November through late February, the main convective activity is centered over central-west Brazil. The SACZ is fully established and the heavy rainfall zone extends over the Altiplano (Fig. 1 and 4b). At upper levels the Bolivian high and the Nordeste trough are observed (Kousky and Ropelewski 1997). There is little change in the areal extension of the deep convection, except over eastern Amazon basin, which experiences an increase in deep convection throughout the period. As deep convection extends into the eastern Amazon, the Atlantic ITCZ weakens.

Beginning in March, the SAMS weakens as the area of deep convection retreats northwestward, faster over central and western sections (Fig. 4c). Over the north coastal regions of Brazil the deep convection only weakens after late April. During the demise phase the Atlantic ITCZ remains weak. In Northeast Brazil the rainy season takes place during April through June, when the ITCZ is in its southernmost position.



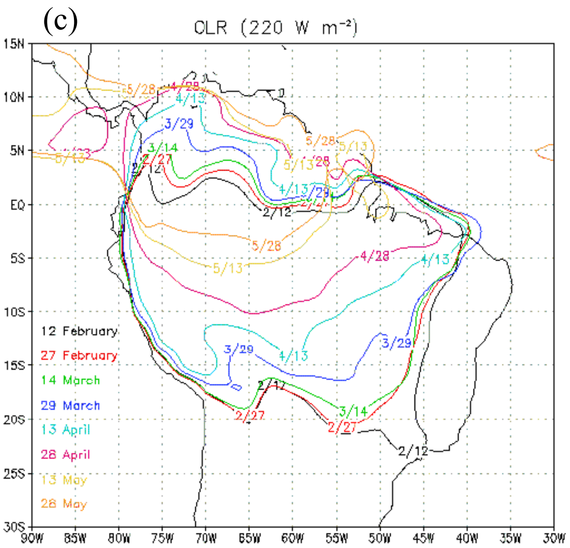
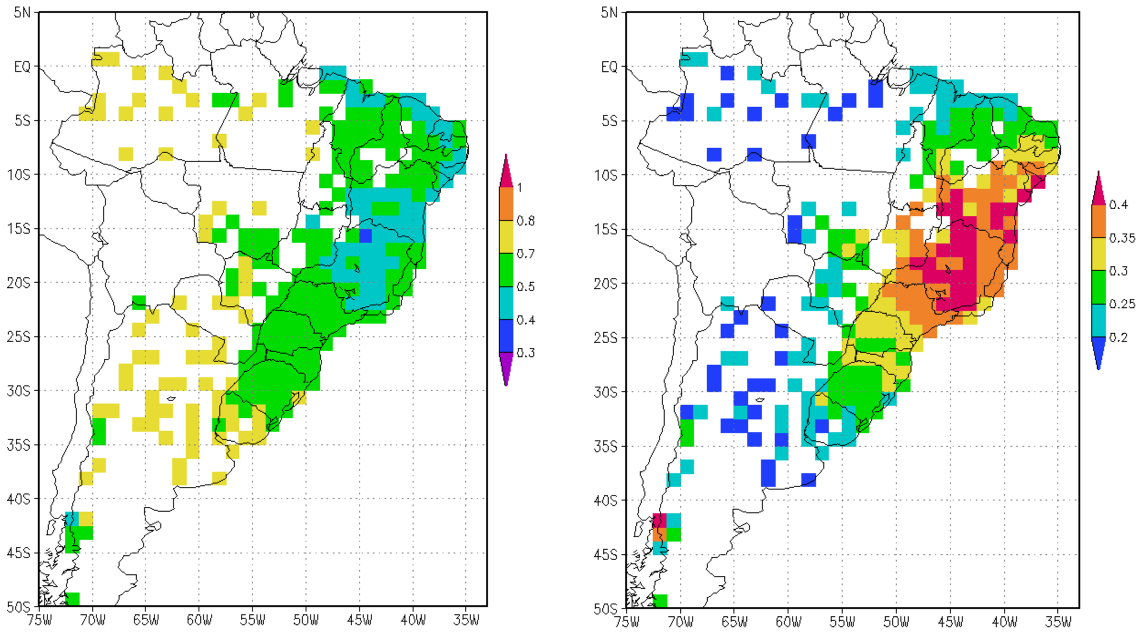


Fig. 4. Mean evolution of the 220 W/m2 OLR contour for the (a) onset, (b) mature phase, and (c) decay phase of the South American Monsoon (from V. Kousky).

3. Variability

Summer precipitation in South America undergoes variability in several time-scales. The spatial distribution of the contribution of synoptic, intraseasonal and interannual/interdecadal variability to the summer (November through March) precipitation variability is shown in Fig. 5 (Ferraz 2004). The region with highest intraseasonal contribution (10-100 day periods) is approximately the same where the synoptic variability shows its minimum, although the contribution of the synoptic variability is still the highest almost everywhere.



(a) (b)

Fig. 5. Contribution of (a) synoptic (b) intraseasonal variability to total variance of precipitation (from Ferraz 2004).

a. Mesoscale and Synoptic Variability

1) Mesoscale Variability

Deep convection during the South America monsoon season frequently undergoes mesoscale organization in certain regions, depending on regional factors. Mesoscale systems are strongly modulated by the diurnal cycle and by transient synoptic systems. Mesoscale Convective Complexes (MCCs) occur frequently in southeastern South America (SESA), between 20°S and 40°S (western South Brazil, Northeast Argentina and Paraguay) (Velasco y Fritsch, 1987; Silva Días, 1987). The MCCs are approximately circular systems of deep cold clouds with diameter of few hundreds kilometers (average area around 5×10^5 km^2 in SESA) and a longer life-time than isolated convective systems (at least 6 hours, 12 hours on average in SESA). Intensification of these convective complexes is associated with the position of the upper-level subtropical jet, which in autumn and spring is over that region, and its interaction with the low-level warm and moist northerly wind. They preferentially initiate in afternoon and mature at night time, which might be partially explained by the diurnal variability of the SALLJ, with late afternoon-evening maximum. The environment of MCCs during initiation and maturity includes a strong low-level jet (e.g., Maddox 1983).

In SESA, highly precipitating mesoscale convective systems during October-April initiate to the east of the Andes, move toward E, NE, and preferentially SE, being associated with a northerly low-level jet and related moisture flux convergence (Machado *et al.* 1998; Nicolini *et al.* 2002). More than 80% of these systems occur during SALLJ events that penetrate farther south from the location of the SALLJ mean maximum (15°S- 20°) (Nicolini *et al.* 2002).

Over Amazonia, most convective systems are smaller (average area of less than 1×10^5 km^2) and have shorter life-time (3-6 hours) than MCCs (Carvalho and Jones 2001; Nieto-Ferreira *et al.* 2003). Convection tends to be aligned to the northeastern coast of South America as result of the inland propagation of coastal squall lines forced by onshore low-level flow, documented by Cohen *et al.* (1995). The coastal band of convective cloudiness increases to a maximum in the late afternoon and weakens during nighttime. After inland propagation, it is reactivated in the afternoon of the next day (Garreaud and Wallace 1997).

2) Synoptic Variability

The day-to-day variability of rainfall over the subtropical South America and western Amazon basin is largely explained by frequent northward incursions of mid-latitude systems to the east of the Andes. Although synoptic disturbances are particularly large and frequent in winter, they are also present during summer, and often reach sufficiently low latitudes to affect the American Monsoon systems (e.g., Garreaud and Wallace 1998; Seluchi and Marengo 2000; Garreaud 2000). The deep northward intrusion of midlatitude systems is the result of their interaction with the Andes topography, which has a significant dynamical impact on the structure and evolution of the synoptic pressure systems that cross South America. In particular, cold fronts tend to be directed northward immediately to the east of the Andes, fostering the advance of cold air incursions (cold surges) well into subtropical (and sometimes tropical) latitudes. During wintertime the major effects of the incursions are temperature drops and strong meridional winds. In summer the largest impact is on the precipitation, through the equatorward propagation (~10 ms^{-1}) of a northwest-southeast oriented band of enhanced convection ahead of the leading edge of the cool air, which tends to be followed by an area of suppressed convection. This synoptic scale banded structure, which maintains its identity for about 5 days, is the dominant mode of the day-to-day variability of the deep convection, contributing with ~25% of summer precipitation in the central Amazonia and ~50% over subtropical South America (Fig. 6, Garreaud and Wallace 1998). These bands also influence convection in the SACZ, lending support to the role of transient disturbances in the maintenance of the SACZ (Lenters and Cook 1995). As in wintertime, these incursions occur with periodicity of about 7 days.

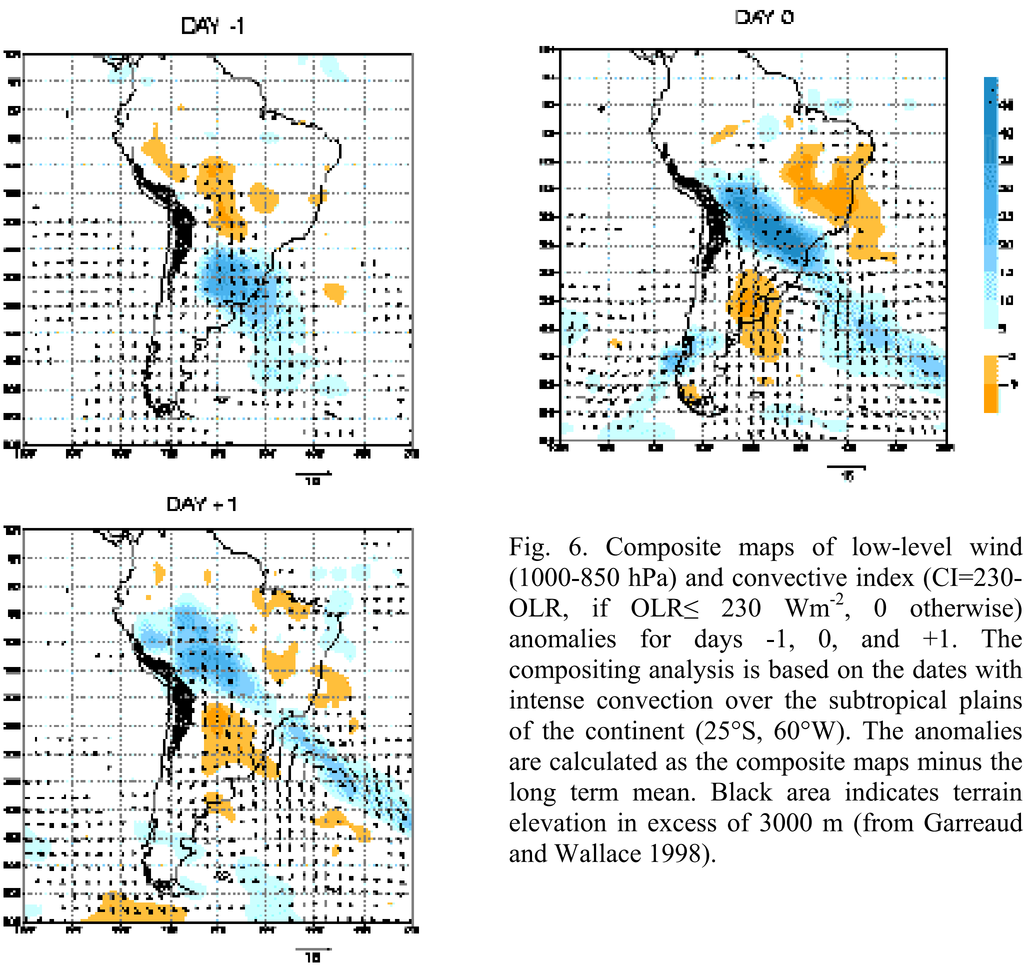


Fig. 6. Composite maps of low-level wind (1000-850 hPa) and convective index (CI=230-OLR, if OLR≤ 230 Wm^{-2}, 0 otherwise) anomalies for days -1, 0, and +1. The compositing analysis is based on the dates with intense convection over the subtropical plains of the continent (25°S, 60°W). The anomalies are calculated as the composite maps minus the long term mean. Black area indicates terrain elevation in excess of 3000 m (from Garreaud and Wallace 1998).

The SALLJ events are conditioned by synoptic variability, and may be separated into two groups with different synoptic evolution: 1) events in which the jet extends farther south, at least to 25°S, 2) events in which the jet leading edge is north of this threshold (Nicolini *et al.* 2002). The LLJs in the first category are stronger, and associated with high (low) moisture convergence and precipitation in SESA (SACZ), while those in the second one are weaker and associated with enhanced (suppressed) precipitation in the SACZ (SESA). During summer (DJF) the first category is less frequent than the second one. The second category is associated with increased probability of extreme events of rainfall and warm temperature in the subtropical region (Penalba and Rusticucci 2004).

b. Intraseasonal Variability

The maximum contribution of the intraseasonal variability (periods in the 10-100 day band) to the total variance is concentrated in Central-East Brazil, including the SACZ, while the lowest values are in western part (Fig. 5).

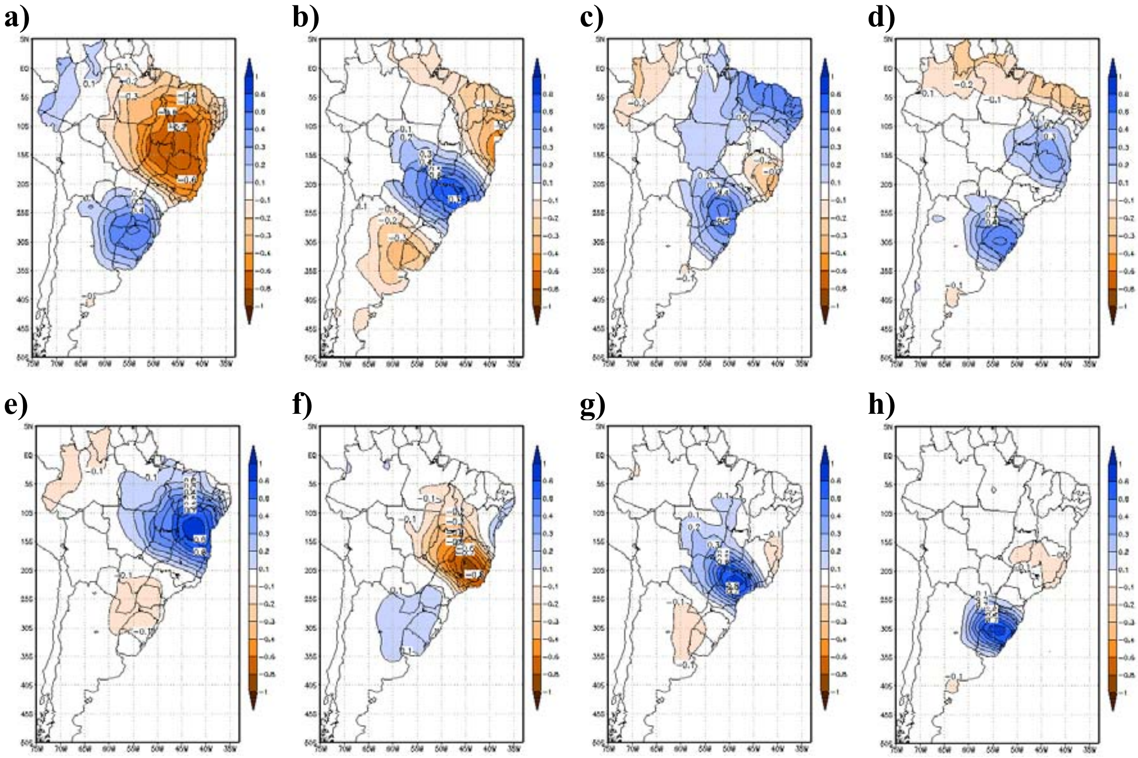


Fig. 7. (a-d) First four EOFs of precipitation, which explain 37,3 % of the intraseasonal variance. (e-h) Factor loadings of the (e) first, (f) second, (g) third, and (h) fifth rotated EOFs, which explain 27.7 % of the intraseasonal variance (from Ferraz 2004).

Ferraz (2004) identified the intraseasonal modes of summer (November through March) precipitation variability. The first non-rotated EOF, which explains 16.3 % of the variance, features a strong center in Central-East Brazil, with the SACZ on its southern edge, and weaker oscillations of opposite signs in the subtropical plains (Fig. 7a). Fig. 5b confirms that the intraseasonal variability is not equally important in both regions. The rotation of modes separates the variability in Central-East Brazil from that over the SACZ (Fig. 7e, f). The first rotated mode (10% of the variance) represents oscillations in Central-East Brazil, while the second one is focused on the SACZ (7.3 % of the variance). Both modes feature anomalies in the subtropical plains that are opposite to those in the main center, but much weaker. Notwithstanding, there is a significant “dipole”-like relationship between precipitation anomalies in Central-East Brazil/SACZ and the subtropics to the south, although with very different magnitude in both centers. The dipole structure appears stronger in OLR analyses and in this case probably merges the first and second modes (e.g. Casarin and Kousky 1986; Nogués-Paegle and Mo 1997, Nogués-Paegle *et al.* 2000).

The local circulation anomalies associated with the first two modes are consistent with the seesaw pattern in precipitation. In one extreme phase, a cyclonic anomaly around 20°S, 50°W (25°S, 45°W) directs the northwesterly moisture flux into Central-East Brazil (SACZ) and decreases the southward transport (Fig. 8a, b). In the opposite phase, an anticyclonic anomaly enhances the moisture flux towards the subtropical plains (Fig. 8c, d). Southwest of this strong circulation anomaly, there is a weaker anomaly of opposite sign. Coherently with this pattern, low-level zonal westerly (easterly) winds over tropical Brazil during summer are associated with an active (inactive) SACZ and net moisture divergence (convergence) over SSA, implying a weak (strong) SALLJ (Herdies *et al.* 2002). Other studies have also reported that similar intraseasonal variations in summer low-level wind regimes over Central Brazil are linked to breaks and active phases of the SAMS (e.g., Jones and Carvalho 2002; Gan *et al.* 2004). The “dipole” structure present in the above mentioned modes does also appear when the analysis is focused on a specific region, like Uruguay (Diaz and Aceituno 2003). Even on interannual time scales similar structure is present (Robertson and Mechoso 2000).

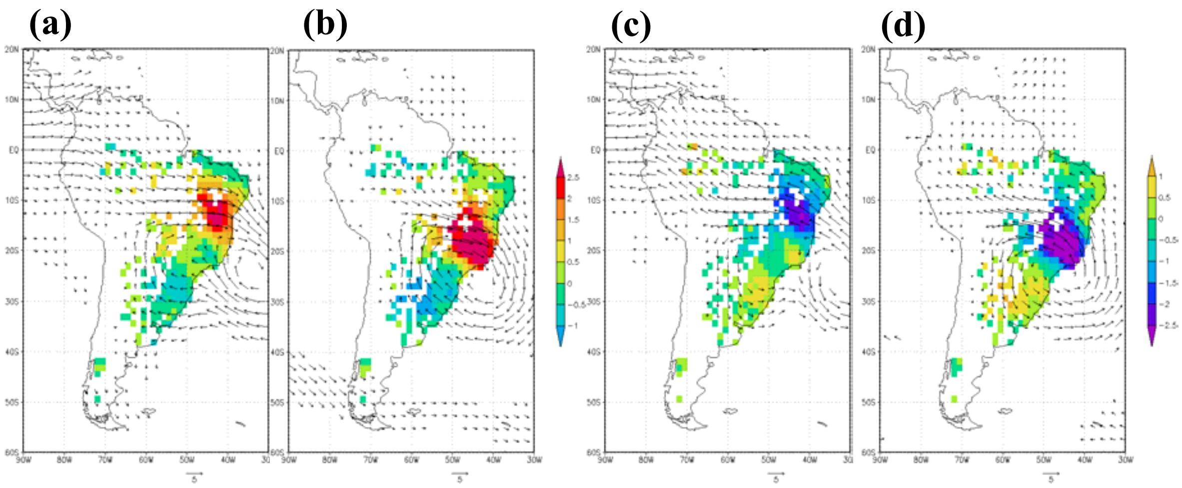


Fig. 8. Composites of rainfall anomalies and vertically integrated moisture flux for wet phases of the (a) first and (b) second rotated principal components, and for dry phases of the (c) first and (d) second rotated principal components in the 30/70 day band (from Ferraz 2004).

Intraseasonal summer precipitation variability is modulated by different time-scales in South America. Spectral analyses of precipitation and OLR show distinct peaks in the intraseasonal band at 20-25 day and 30-70 day (e. g. D’Almeida, 1997; Liebmann *et al.* 1999). Peaks are also observed in the 10-20 day band. The four rotated modes shown in Fig. 7 appear among the five most important modes in several intraseasonal frequency bands (periods of 10-20, 20-30, and 30-70 day). This means that precipitation over South America results from a complex interaction of different time scales (Nogués-Paegle *et al.* 2000).

The origin of the circulation anomalies associated with the first modes of precipitation variability on intraseasonal time-scales over South America is not yet completely understood. They seem to be related to wave-trains propagating southeastward from West or Central Pacific, rounding the southern tip of South America and turning toward the northeast, as part of larger scale systems, whose associated convection in West and Central Pacific may originate or modify those wave trains (Grimm and Silva Dias 1995; D'Almeida 1997; Nogués-Paegle *et al.* 2000). Fig. 9 shows that the first mode in the 30-70 day band seems to originate from the convective anomalies associated with the Madden-Julian Oscillation (MJO) in west Pacific and in central Pacific, where the SPCZ is enhanced and shifted eastward. Grimm and Silva Dias (1995) have shown that the circulation anomalies leading to enhanced rainfall in Central-East Brazil/SACZ can be produced by convective anomalies in the Central Pacific. Liebmann et al (2004a) have shown that SACZ rain events preferentially occur around 20 days prior to the peak in MJO convection in the western tropical Pacific, while events downstream of the SALLJ tend to occur around 2 days after the peak in convection. Also, Carvalho *et al.* (2004) found that the MJO plays an important role in modulating persistence of intense SACZ events.

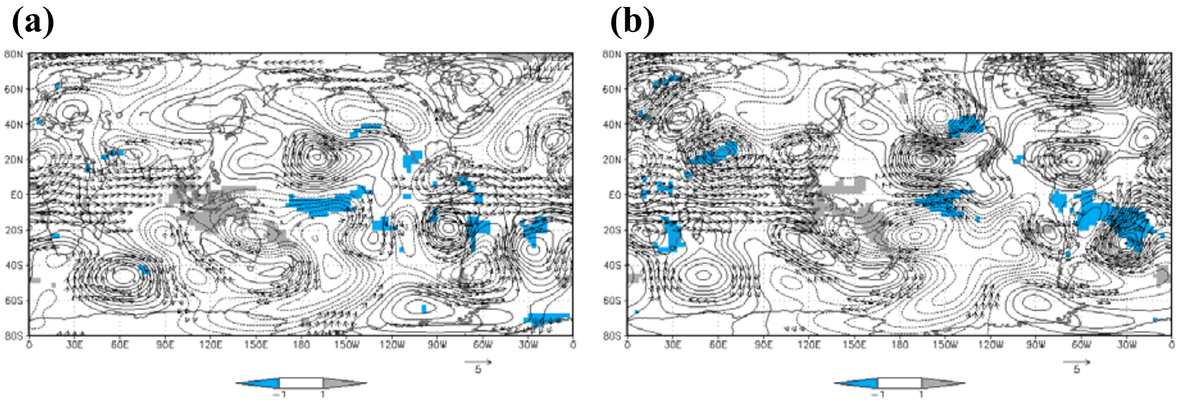


Fig. 9. Composites of OLR and 200 hPa streamfunction anomalies associated with the days when the first rotated principal component in the 30/70 day band (Fig. 7e) reaches the threshold of 1.4 standard deviations. (a) Day -4, (b) day 0. Only statistically significant OLR anomalies are represented (blue shade for negative and gray shade for positive anomalies) (from Ferraz 2004).

Kiladis and Weickmann (1992) associated tropical convection in the Pacific Ocean with circulation anomalies that propagate first poleward and then equatorward over South America, and analyzed cases in which SACZ variations (in the 6-30 day time scales) are forced by westerly perturbations originating in the extratropics. Enhanced convection is activated by upper-level troughs and the convection occurs in the upward motion induced by the advection of cyclonic vorticity ahead of the trough axis, as in a midlatitude baroclinic wave. The troughs are accompanied by the intrusion of cold fronts into the tropics from higher southern latitudes. Regions in which upper-level westerly flow lies near a tropical convergence zone (as the SPCZ and the SACZ) are prone to larger interaction between westerly disturbances and tropical convection.

Besides the remote influence, intraseasonal variability of the summer monsoon may also be influenced by regional factors. Regional circulation, like the SALLJ, may be modulated by intraseasonal fluctuations of zonal flow above the Andes and consequent fluctuations of the orographically bound cyclone east of the Andes. This relationship could be used in a forecast scheme of SALLJ variations (Byerle and Paegle 2002; Wang and Fu 2004).

c. Interannual Variability

Nogués-Paegle and Mo (2002) isolated the main modes of interannual variability of summer (DJF) precipitation over South America through rotated empirical orthogonal functions of the seasonal rainfall anomalies using reconstructed rainfall data. They found three fundamental patterns of variability shown in Fig. 10. The dominant mode (REOF1) is associated with ENSO, with negative rainfall anomalies during warm ENSO events in northern South America, and positive anomalies south of 25 S. REOF2 is influenced by Atlantic sea surface temperatures, with warm tropical South Atlantic SST associated with positive rainfall anomalies in the eastern half of the continent centered at the Equator. Both the Atlantic and the Pacific Ocean influence REOF3. The pattern is similar to that of REOF2, but anomalies are displaced about 10°S and there is an additional center at about 30°S. These two patterns were in phase between 1950–1962 and between 1983 to 1993, with a dominant Atlantic influence during those years. The amplitude of these two modes was small during mid- 50s to mid-60s and out of phase from 1968–70, resulting in persistent dry conditions over the upper basin of La Plata River. This is corroborated by the extent and level of the Pantanal, South American wetlands, with anomalously low values during this period.

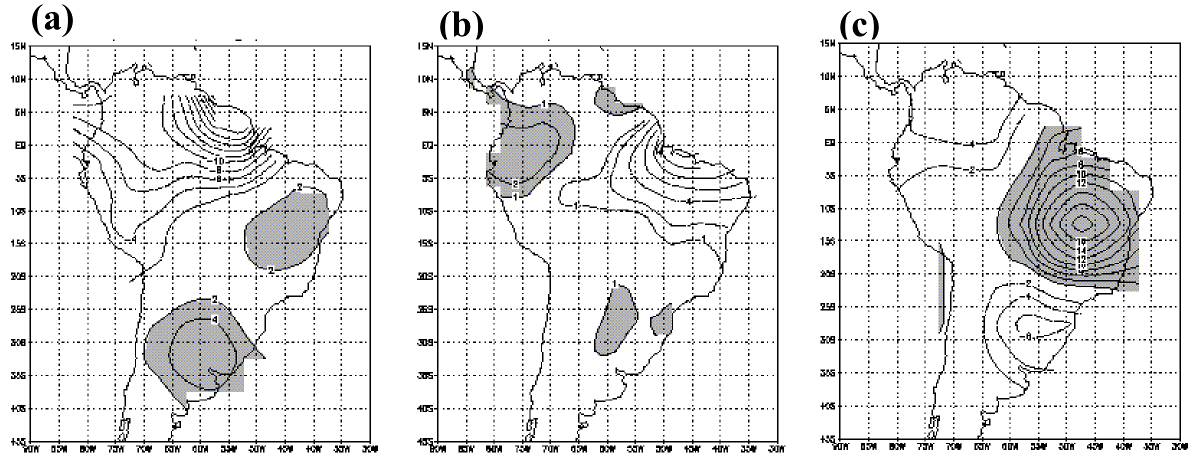


Fig. 10. (a) REOF1, (b) REOF2 and (c) REOF3 for DJF rainfall over South America, explaining 12%, 10.8% and 7.2% of the total variance by each REOF respectively. Contour interval is 2 non-dimensional units. Zero contours are omitted. Contours -1 and 1 are added in (b). (From Nogués-Paegle and Mo 2002)

The main source of interannual variability of precipitation during the summer monsoon season is El Niño-Southern Oscillation (ENSO). ENSO events (El Niño/La Niña) have significant impact on rainfall (Ropelewski and Halpert 1987; Aceituno 1988; Grimm *et al.* 2000; Grimm 2003, 2004). The impact of ENSO shows strong regional characteristics and strong intraseasonal changes, suggesting the prevalence of regional processes over remote influences during part of the season. Due to these intraseasonal changes, the seasonal analysis of ENSO impacts usually smoothes out consistent and strong anomalies that appear for shorter periods.

The anomalous tropical heat sources associated with ENSO events perturb the Walker and Hadley circulation over South America, and generate Rossby wave trains that produce important effects in the subtropics and extratropics of South America. This response appears in the leading EOFs of circulation anomalies, which are referred to as Pacific-South American modes (PSA1 and PSA2) (Kidson 1999, Mo 2000). The PSA1 and PSA2 patterns appear in various time-scales. While PSA1 appears to be related to ENSO, PSA2 appears to be associated with the quasi-biennial component of ENSO.

During El Niño events, in the early summer monsoon season (November) remotely produced atmospheric perturbations prevail over Brazil (Grimm 2003). Anticyclonic low-level anomalies predominate over central-east Brazil, in the tropics and subtropics, due to the enhanced subsidence over the Amazon and to Rossby waves in the subtropics. Easterly moisture inflow from the equatorial Atlantic is favored, but diverted towards northern South America and south Brazil. There are negative precipitation anomalies in north and central-east Brazil and positive ones in south Brazil (Fig. 11). These precipitation anomalies are favored by the perturbation in the Walker and Hadley circulation over East Pacific and South America, and by a Rossby wave-train over southern South America that originates in the eastern Pacific.

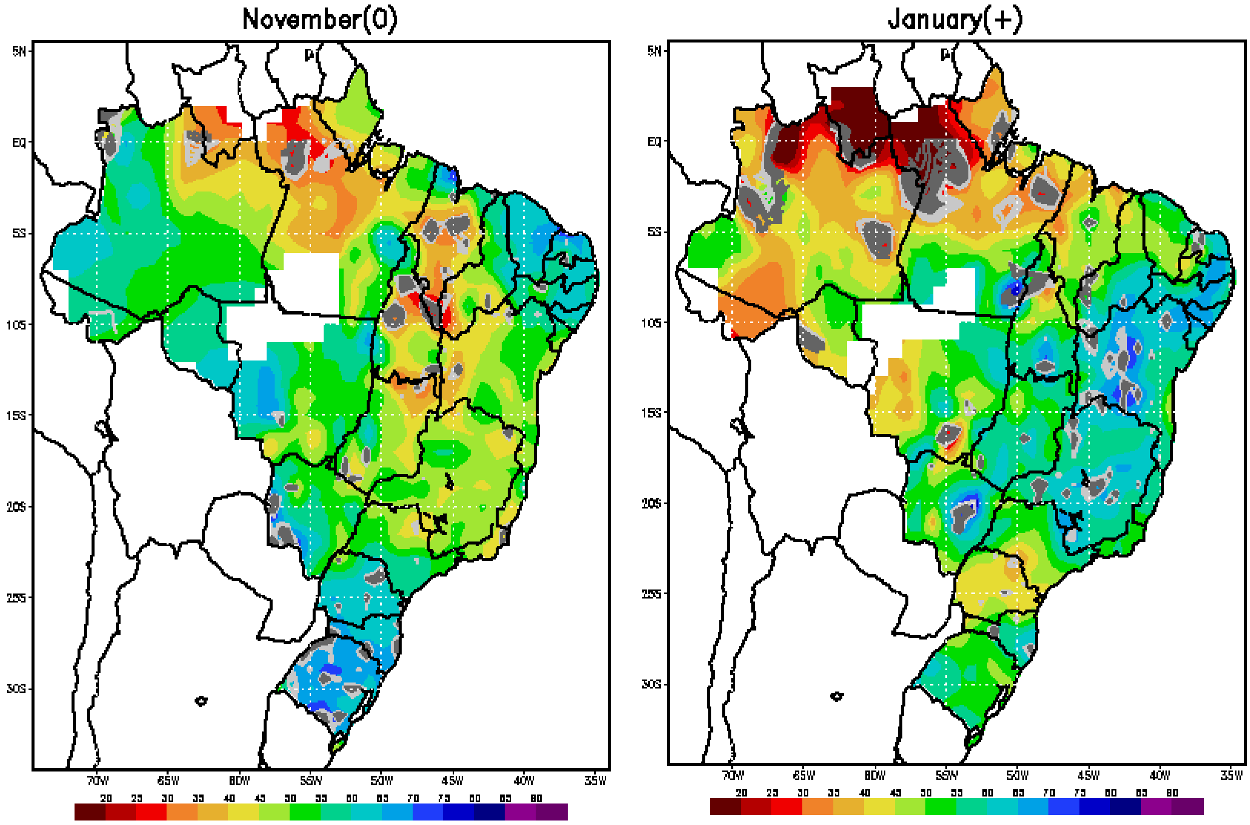


Fig. 11. Monthly mean precipitation percentiles expected for the indicated month of El Niño events, Shadowed areas have precipitation anomalies consistent over 90% confidence level (from Grimm 2003).

In January, with the enhancement of the continental subtropical heat low by anomalous surface heating during the spring, there is anomalous low-level convergence and cyclonic circulation over southeast Brazil, while at upper-levels anomalies of divergence and anticyclonic circulation prevail. This anomalous circulation directs moisture flux towards central-east Brazil, causing moisture convergence in this region. A favorable thermodynamic structure enhances precipitation over central-east Brazil, the dry anomalies in north Brazil are displaced northward, and the anomalies in south Brazil almost disappear (Fig. 11). In February, after the above normal precipitation of January, the surface temperature anomalies turn negative and the precipitation diminishes in central-east. There are negative rainfall anomalies in north Brazil and in the SACZ, and positive ones in south Brazil.

During La Niña events the circulation and precipitation anomalies are fairly opposite to those described for El Niño events, sometimes with little shifts in the position of the strongest anomalies, and in the magnitude of the anomalies (Grimm 2004).

The impact of ENSO over rainfall in the Altiplano is related to the strengthening (weakening) of the upper-level westerly winds at subtropical latitudes east of the central Andes. This leads to a decrease (increase) in the moisture transport from the continental lowlands into the Altiplano (Garreaud and Aceituno 2001).

Although the described impacts are consistent during ENSO events, there is still significant inter-events variability, associated with differences in the SST anomalies in the subtropical South Pacific (Barros and Silvestri 2002), which produce different atmospheric teleconnections (Vera *et al.* 2004).

Besides the influence of the ENSO-related SST anomalies, there are other connections between precipitation and SST anomalies, although it is not always easy to separate cause and effect in the statistical relationship. Enhanced (suppressed) precipitation in the SACZ is related with colder (warmer) SST in southwestern subtropical Atlantic, near the SACZ (Doyle and Barros 2002; Robertson and Mechoso 2000). Grimm (2003) showed that January rainfall in Central-East Brazil is positively correlated with November SST in the oceanic SACZ, off the southeast coast of Brazil, while negatively correlated with January SST in the same region. Anomalies of precipitation and circulation in the region, like those associated with El Niño events in November, favor increased shortwave radiation and set up warm SST anomalies. On the other hand, the enhanced convection and rainfall in January lead to negative SST anomalies. In this case, the atmosphere controls the ocean, but one might speculate that the warmer SST in November helps trigger the regional circulation anomalies that lead to enhanced precipitation in January. Although in the relationship ZCAS-SST, the SST anomalies seem to be result of the convection anomalies in the ZACS, there are possible feedback mechanisms between SST and the atmosphere (Robertson *et al.* 2003; Chaves and Nobre 2004).

The mechanism suggested by Grimm (2003) to explain the intraseasonal changes of circulation and precipitation anomalies during ENSO events, involving land-surface conditions, could also explain changes during other years, through the feedback of the initial springtime soil moisture and vegetative cover upon the peak summer climate (Higgins *et al.* 1998).

d. Decadal/Interdecadal Variability

Analyses of variability of South American precipitation in longer time-scales have shown that there are modes with interdecadal variability, in connection with regional or global SST variations (e.g., Robertson and Mechoso 2000; Zhou and Lau 2001; Nogués-Paegle and Mo 2002). Long-term variations of rainfall in Northeast Brazil and eastern Amazonia have been related with variations of inter-hemispheric SST gradient in the tropical Atlantic, associated with increase in South Atlantic SSTs, while variations in SESA seem related with SST both in the Atlantic and Pacific Ocean.

Southeastern South America rainfall shows interdecadal variability in connection with global modes of interdecadal non-ENSO SST variability, like the Pacific Decadal Oscillation (PDO) and the North Atlantic Oscillation (NAO) (Grimm and Canestraro 2003). The first mode shows a change of phase in the mid 70's, when there is a change of phase in the PDO. Other studies have shown a substantial increase in southern/southeastern Brazil rainfall after the 70's (e.g., Liebmann *et al.* 2004b). Even the spatial patterns of intraseasonal variability of rainfall undergo interdecadal modulation (Ferraz and Grimm 2004).

Evidences of interdecadal variations also appear in the river flows. Robertson and Mechoso (2000) found a near cyclic 15-17 year component in the SACZ variability that is also present in river flows of the La Plata Basin. An 8-9 year component was also identified in the Paraná and Paraguay rivers, apparently related with NAO (Robertson and Mechoso 1998). The substantial increase of rainfall in southeastern South America after the 70's does also appear in the river flows (Genta et al 1998). In western and central Amazonia there was negative trend, while in eastern Amazonia the trend was positive from 1960s to 1980s, according to Dias de Paiva and Clarke (1995). In the upper Paraguay Basin the river flows were much lower during the period 1960-70 than in the periods before and after (Collischonn *et al.* 2001).

e. Paleo Climate

It is believed that full glacial climates throughout South America were colder than today by about 5°C with moisture patterns showing distinct regional differences. Observations in South America based on paleodata for the Last Glacial Maximum (LGM) show a predominant cooling and drying of the region. The Tropical half of the continent was drier (with the exception of Brazil's Nordeste) and the subtropical part was wetter. The available paleodata were compared to simulation results from the Paleoclimate version of the National Center for Atmospheric Research coupled climate system model (CCSM1.4) to see if the simulations are able to capture the spatial distribution of humid and arid climate and attempt to explain the physical mechanisms behind the moisture distribution. Although there are discrepancies between the model and observations, some major features are captured. In the tropical part, the summer ITCZ during the LGM doesn't stretch all the way to the continent and this prevents moisture inflow from the adjacent Atlantic Ocean into the continental area. In the subtropics, LGM low-level westerlies are weaker and data show characteristics of more humid climate (Wainer *et al.* 2004 and references therein).

4. Modeling Studies

Atmospheric general circulation models (AGCMs) have been used to investigate the individual roles played by the shape and location of continents, topography, and SST distributions in the characteristics of the SAMS. The results of Lenters and Cook (1995), for example, suggest that the presence of the South American continent alone, without topography or longitudinal structure in the SST field, suffices to obtain the summer precipitation maxima in the Amazon, SACZ, and northwestern South America. Topography, however, induces precipitation maxima on the eastern flank of the central Andes and the western flank of the southern Andes, sharpens the SACZ, and strengthens and repositions the Amazonian precipitation maximum. SST affects moderately the position of the Amazonian precipitation. Fu *et al.* (2001) suggest that the seasonality of the land surface dominates that of the precipitation in the western Amazon throughout the year and that in the eastern Amazon during the solstices via a direct thermal circulation and propagation of stationary Rossby waves.

An AGCM with enhanced horizontal resolution over South America can capture the nocturnal precipitation maximum, as over the Great Plains of North America (Wang *et al.* 1999). Ensemble simulations by Nieto-Ferreira *et al.* (2000) using an AGCM with enhanced resolution over the Amazon show more intra-ensemble variability in the SACZ region than over the Amazon, which is consistent with lower predictability in the former than in the latter region.

Recent work has shown that AGCMs have difficulties in capturing the diurnal cycle of precipitation over regions of strong monsoon circulations. For example, Betts and Jakob (2002) demonstrate that precipitation simulated by the ECMWF forecast over the Amazon starts about 2 hours after sunrise, which is several hours earlier than observed. This precipitation structure appears as a large-scale feature over the Amazon region as the model produces deep convection and convective rain as soon as the surface heating starts deepening the planetary boundary layer (PBL). Yang and Slingo (2001) find similar difficulties with the climate version of the U.K. Met. Office (UM, version HadAM3). The reasons for these difficulties are a topic of current research.

The SAMS climatology, as simulated by six different AGCMs, was evaluated by Zhou and Lau (2002). They simulate reasonably well the large-scale features of the SAMS. However surface pressure is overestimated, resulting in an excessively strong SAMS. The SALLJ is not well resolved. There are large rainfall errors in association with the Andes and the Atlantic ITCZ, indicating problems with steep mountains and parameterization of convective processes.

Regional models have shown promise for the study of climate over South America. Misra *et al.* (2002) examine simulations of the austral summer season for different phases of the El Niño-Southern Oscillation cycle with the Regional Spectral Model (RSM) developed at the National Centers for Environmental Prediction (NCEP). The simulated interannual variability of precipitation over the

Amazon Basin and other parts of South America compare reasonably well with observations. A detailed moisture budget reveals that moisture flux convergence determines most of the interannual variability of precipitation over the Amazon Basin and the Nordeste region of Brazil, and that both surface evaporation and surface moisture flux convergence are critical in determining the interannual variability of precipitation over the Gran Chaco area and SACZ. Also Seth and Rojas (2003), using RegCM (ICTP), were able to simulate the dramatically different large-scale circulations, as well as the resulting rainfall differences during summers in opposite phases of ENSO. However, the rainfall in Amazonia is underestimated, as is the low-level moisture transport from the Atlantic. Simulation of January precipitation for six years with the MM5 model (Penn State/NCAR) by Menendez (2004) also shows good reproduction of interannual variations, although the skill to reproduce the precipitation amounts varies much according to the region. The main mean precipitation patterns are captured, but precipitation over Uruguay and central Argentina is underestimated.

The role of SALLJ in tropical-subtropical/extratropical interactions and the development of the monsoonal precipitation have been investigated using the mesoscale Eta model. The warm season SALLJ appears to be topographically bounded and diurnally modulated, with a nighttime maximum intensity. The model forecasts also show that the diurnal cycle of the SALLJ favors increased nighttime moisture flux convergence at its exit region in southeastern South America, associated with nighttime increased precipitation (Berbery and Collini 2000; Saulo *et al.* 2000).

5. Predictability

The consistent impact of ENSO events lends some predictability on interannual time scales. Even so, there are uncertainties associated with the inter-event variability. Besides, this predictability based on tropical Pacific SST is restricted to specific times of the year and certain regions (e.g., Montecinos *et al.* 2000). The influence of the Atlantic SSTs is not well known yet, except on Northeast Brazil. There might be some predictability on the decadal/interdecadal time scales, but the causes of the interdecadal variability in the region are still not well known, which limits its prediction. The intraseasonal time scale shows relationship with quasi-periodic oscillations, like the Madden-Julian Oscillation, which allows some predictability in this band. Notwithstanding, there are other shorter intraseasonal oscillations not well understood, that interact/superimpose on the MJO and may modify them substantially. There are studies indicating that MJO activity might increase predictability over South America (e.g., Jones and Schemm 2000; Wheeler and Weickmann, 2001; Mo 2001). Regarding the monsoon onset, potential precursors for onset of the wet season over tropical South America may help to establish a prediction scheme of early or later onsets (Li and Fu 2004).

Predictability studies with global models have not supported the expectation of increased predictability generated by the relationships documented between SAMS and slowly varying surface conditions in SST and land surface conditions. Goddard *et al.* (2003) show that regions such as southern Amazonia and the SAMS core regions exhibit relatively low climate predictability on seasonal-to-interannual time scales, since circulation and rainfall anomalies in these regions are more dependent on regional forcing than remote forcing

One AGCM study (Zhou and Lau 2002) shows higher skill in the prediction of El Niño impacts over tropical than over subtropical South America. In the former region anomalies are governed by the Walker cell shift that is directly induced by the central-eastern Pacific warming, while in the latter anomalies have large uncertainties due to poorly resolved orographic relief and surface conditions. Grimm *et al.* (2000) showed that the consistent impact of ENSO events on precipitation over southeastern South America in a long AGCM run is only reasonably realistic during spring, when it is at its maximum.

A study with a different AGCM (Marengo *et al.* 2003) shows that mean precipitation in the

SAMS region is more successfully simulated than in the NAMS, which misses the characteristic double peaked structure. Nevertheless, the simulated interannual variability is very different from the observed for both monsoon systems. During summer (DJF) the model shows relatively good skill in northern South America and in a region including part of southern Brazil, southern Paraguay and northern Argentina (where MCCs are more frequent). Another study (Misra 2004) compares the skill of an AGCM over South America for the austral summer in seasonal and interannual predictions with observed SSTs. The seasonal precipitation climatology is vastly superior in the seasonal runs except over the Nordeste region where the multiannual runs show a marginal improvement. The seasonal runs outperform the multiannual model integrations both in deterministic and probabilistic skill. All model predictions clearly beat persistence in regard to the interannual precipitation anomalies over the Amazon River basin, Nordeste, SACZ, and subtropical South America.

References

Aceituno, P., 1988: On the functioning of the Southern Oscillation in the South American sector. Part I: Surface climate. *Mon. Wea. Rev*., **116**, 505-524.

Barros, V. R., and G. E. Silvestri, 2002: The relation between sea surface temperature at the subtropical South-Central Pacific and precipitation in Southeastern South America. *J. Climate*, **15**, 251–267.

Berbery, E. H., and E. Collini, 2000: Springtime precipitation and water vapor flux convergence over Southeastern South America. *Mon. Wea. Rev*., **128**, 1328-1346.

_____, and V. R. Barros, 2002: The hydrologic cycle of the La Plata basin in South America. *J. Hydromet.*, **3**, 630-645.

Betts, A. K. and C. Jakob, 2002: Evaluation of the diurnal cycle of precipitation, surface thermodynamics, and surface heat fluxes in the ECMWF model using LBA data. *J. Geophys. Res.*, **107**, 12-1/12-7.

Byerle, L. A. and J. Paegle, 2002: Description of the seasonal cycle of low-level flows flanking the Andes and their interannual variability. *Meteorologica,* **27**, 71-88.

Campetella, C. and C. Vera, 2002: The influence of the Andes mountains on the South American low-level flow. *Geophys. Res. Lett*., **29** (17), 1826, doi: 10.1029/2002GL015451.

Carvalho, L. M. V., and C. Jones, 2001: A satellite method to identify structural properties of mesoscale convective systems based on the maximum spatial correlation tracking technique (MASCOTTE). *J. of Appl. Meteor*. **40**, 1683-1701.

_____, C. Jones, and B. Liebmann, 2004: The South Atlantic Convergence Zone: intensity, form, persistence, and relationships with intraseasonal to interannual activity and extreme rainfall. *J. Climate*, **17**, 88–108.

Casarin, D. P. and V. E. Kousky, 1986: Anomalias de precipitação no sul do Brasil e variações na circulação atmosférica. *Rev. Bras. Met*., **1**, 83-90.

Chaves, R. R. and P. Nobre, 2004: Interactions between sea surface temperature over the South Atlantic Ocean and the South Atlantic Convergence Zone. Geophysical Research Letters, v. 31, n. 3, L0320410.1029/2003GL018647.

Chen, T.-C., S. P. Weng, and S. Schubert, 1999: Maintenance of austral summer upper-tropospheric circulation over tropical South America: the Bolivian high-Nordeste low system. *J. Atmos. Sci*., **56**, 2081-2100.

Cohen J., M. Silva Dias, and C. Nobre, 1995: Environmental conditions associated with Amazonian squall lines: A case study. *Mon. Wea. Rev*., **123**, 3163-3174.

Collishchonn, W., C. E. M. Tucci, and R. T. Clarke, 2001: Further evidence of changes in the hidrological regime of the River Paraguay: part of a wider phenomenon of climate change? *J. Hydrology*, **245**, 218-238.

Dias de Paiva, E. M. C., and R. T. Clarke, 1995: Time trends in rainfall records in Amazonia. *Bull. Am. Meteorol. Soc*., **76**, 2203-2209.

Diaz, A., and P. Aceituno, 2003: Atmospheric circulation anomalies during episodes of enhanced and reduced convective cloudiness over Uruguay. *J. Climate,* **16**, No. 19, 3171–3185.

D'Almeida, C., 1997: Oscilações intrasazonais de precipitação na estação chuvosa em São Paulo e condições atmosféricas associadas. MSc. Dissertation. Department of Atmospheric Sciences, University of São Paulo.

Doyle, M. E., and V. R. Barros, 2002: Midsummer low-level circulation and precipitation in subtropical South America and related sea surface temperature anomalies in the South Atlantic. *J. Climate*, **15**, 3394–3410.

Ferraz, S. E. T., and A. M. Grimm, 2004: Intraseasonal variability in Summer: modulation by the interdecadal variability. Proceedings of the 13th Brazilian Meteorological Congress, Fortaleza, Brazil.

Ferraz, S. E. T., 2004: Intraseasonal variations of the summer precipitation over South America. PhD. Thesis. Department of Atmospheric Sciences, University of São Paulo.

Figueroa, S. N., P. Satyamurty, and P. L. Silva Dias, 1995: Simulations of the summer circulation over the South American region with an eta coordinate model. *J. Atmos. Sci.*, **52**, 1573-1584.

Fu, R., B. Zhu, and R. E. Dickinson, 1999: How do atmosphere and land surface influence seasonal changes of convection in the tropical Amazon? *J. Climate*, **12**, 1306-1321.

_____, R. E. Dickinson, M. Chen, and H.Wang, 2001: How do tropical sea surface temperaturas influence the seasonal distribution of precipitation in the equatorial Amazon? *J. Climate*, **14**, 4003-4026.

Gan, M. A., V. E. Kousky, and C. F. Ropelewski, 2004: The South America monsoon circulation and its relationship to rainfall over West-Central Brazil. *J. Climate*, **17**, 47–66.

Garreaud, R. D., 2000: Cold air incursions over subtropical South America: mean structure and dynamics. *Mon. Wea. Rev.,* **128**, 2544–2559.

_____, and J. M. Wallace, 1997: The diurnal march of convective cloudiness over the Americas. *Mon. Wea. Rev*., **125**, 3157-3171.

_____, and _____, 1998: Summertime incursions of midlatitude air into subtropical and tropical South America. *Mon. Wea Rev*., **126**, 2713-1733.

_____, and P. Aceituno, 2001: Interannual rainfall variability over the South American Altiplano. *J. Climate*, **14**, 2779–2789.

Genta, J. L., G. Perez-Iribarren and C. R. Mechoso, 1998: A recent increasing trend in the streamflow of rivers in Southeastern South America. *J. Climate*, **11**, 2858-2862.

Goddard, L., A. G. Barnston, S. J. Mason, 2003: Evaluation of the IRI's "Net Assessment" Seasonal Climate Forecasts: 1997–2001. *Bull. Amer. Met. Soc*., **84**, 1761–1781.

Grimm, A. M., and P. L. Silva Dias, 1995: Analysis of tropical–extratropical interactions with influence functions of a barotropic model. *J. Atmos. Sci*., **52**, 3538–3555.

_____, Grimm, A. M., C. F. Ropelewsky, e S. Mason, 2000: Does a climate model reproduce consistent ENSO precipitation signals in Southern South America? *Proceedings of the 15th Conference on Hydrology,* Long Beach, California, American Meteorological Society, 186-189.

_____, V. R. Barros, M. E. Doyle, 2000: Climate variability in Southern South America associated with El Niño and La Niña events. *J. Climate*, **13**, 35–58.

_____, C. D. Canestraro, 2003: Interdecadal oscillations in southeastern South America and their relationship with sea surface temperature. Abstracts, International Symposium on Climate Change (ISCC), Beijing, China, 166.

_____, 2003: The El Niño impact on the summer monsoon in Brazil: regional processes versus remote influences. J. Climate,16, 263–280.

_____, 2004: How do La Niña events disturb the summer monsoon system in Brazil? Climate Dynamics, 22, n. 2-3, 123-138.

Gutman, G. J., and W. Schwerdtfeger, The role of latent and sensible heat for the development of a high pressure system over the subtropical Andes, in the summer. *Meteor. Rundsch*., **18**, 69-75.

Herdies, D. L., A. Da Silva, M. A. Silva Dias, and R. Nieto-Ferreira, 2002: Moisture budget of the bimodal pattern of the summer circulation over South America. *J. Geophys. Res*., **107**, 42/1-42/10.

Higgins, W., K. C. Mo, Y. Yao, 1998: Interannual variability of the US summer precipitation regime with emphasis on the southwestern monsoon. *J. Climate*, **11,** 2582-2606.

Jones, C., and J. Schemm, 2000: The influence of intraseasonal variations on medium-to extended-range weather forecasts over South America. *Mon. Wea. Rev*., **128**, 486–494.

_____, and L. M. Carvalho, 2002: Active and break phases in the South American Monsoon System. *J. Climate*, **15**, 905–914.

Kiladis, G. N., and K. M. Weickmann, 1992: Circulation anomalies associated with tropical convection during northern winter. *Mon. Wea. Rev.*, **120**, 1900-1923.

Kidson, J. W., 1999: Principal modes of Southern Hemisphere low frequency variability obtained from NCEP/NCAR reanalysis. *J. Climate*, **12**, 2808-2823

Kousky, V. E., 1988: Pentad outgoing longwave radiation climatology for the South American sector. *Rev. Bras. Meteo.*, **3**, 217-231.

_____, and C. F. Ropelewski, 1997: The tropospheric seasonally varying mean climate over the Western Hemisphere (1975-1995). NCEP/Climate Prediction Center ATLAS N°3, 135 pp.

Kodama, Y-M., 1992: Large-scale common features of sub-tropical precipitation zones (the Baiu Frontal Zone, the SPCZ, and the SACZ). Part I: characteristics of subtropical frontal zones. *J. Meteor. Soc. Japan*, **70**, 813-835.

Lenters, J. D., K. H. Cook, 1999: Summertime precipitation variability over South America: role of the large-scale circulation. *Mon. Wea. Rev.,* **127**, 409–431.

Li, W., and R. Fu, 2004: Transition of the large-scale atmospheric and land surface conditions from the dry to the wet season over Amazonia as diagnosed by the ECMWF re-analysis. J. Climate, **17**, 2637-2651.

Liebmann, B., G. N. Kiladis, J. A. Marengo, T. Ambrizzi, J. D. Glick, 1999: Submonthly convective variability over South America and the South Atlantic Convergence Zone. *J. Climate*, **12**, 1877–1891.

_____, G. N. Kiladis, C. S. Vera, A. C. Saulo, 2004a: Subseasonal variations of rainfall in South America in the vicinity of the low-level jet east of the Andes and comparison to those in the South Atlantic Convergence Zone. *Accepted in J. Climate.*

_____, C. Vera, L. M.V. Carvalho, I. A. Camilloni, M. Hoerling, D. Allured, V. Barros, J. Baez, M. Bidegain, 2004b: An observed trend in Central South American precipitation. *Accepted in J. Climate.*

Machado, L. A. T., W. B. Rossow, R. L. Guedes, and A. W. Walker, 1998: Life cycle variations of mesoscale convective systems over the Américas. *Mon. Wea. Rev*., **126**, 1630-1654.

Maddox, R. A., 1983: Large-scale meteorological conditions associated with midlatitude, mesoscale convective complexes. *Mon. Wea. Rev*., 1475-1493.

Marengo, J. A., B. Liebmann, V. Kousky, N. Filizola, and I. Wainer, 2001: On the onset and end of the rainy season in the Brazilian Amazon Basin., *J. Climate*, **14**, 833-852.

_____, Cavalcanti, I. F. A., P. Satyamurty., I. Trosnikov, C. A. Nobre., J. P. Bonatti., H. Camargo, G. Sampaio., M. B. Sanches, A. O. Manzi, C. A. C. Castro, C. D'Almeida., L. P. Pezzi, and L. Candido, 2003: Assessment of regional seasonal rainfall predictability using the CPTEC/COLA atmospheric GCM. *Climate Dynamics*, **21**, 459–475

_____, W. R. Soares, C. Saulo, and M. Nicolini 2004a: Climatology of the Low-Level Jet east of the Andes as derived from the NCEP reanalyses. *J. Climate*, **17**, 2261-2280

Menendez, C., M. F. Cabré, and M. N. Nuñez, 2004: Interannual and diurnal variability of January precipitation over subtropical South America simulated by a regional climate model. Available under http://www.clivar.org/publications/exchanges/ex29/ex29_cont.htm

Misra, V., P. A. Dirmeyer, B. P. Kirtman, H.-M. H. Juang, and M. Kanamitsu (2002), Regional simulation of interannual variability over South America, *J. Geophys. Res.*, *107*(D20), 8036, doi:10.1029/2001JD900216.

_____, 2004: An evaluation of the predictability of austral summer season precipitation over South America. *Journal of Climate*, **17**, 1161-1175.

Mo, K. C., 2001: Adaptive filtering and prediction of intraseasonal oscillations. *Mon. Wea. Rev.,* **129**, 802–817.

_____, 2000: Relationships between low-frequency variability in the Southern Hemisphere and sea surface temperature anomalies. *J. Climate*, **13**, 3599-3610.

Montecinos, A., A. Diaz, and P. Aceituno, 2000: Seasonal diagnostic and predictability of rainfall in subtropical South América based on tropical Pacific SST. *J. Climate*, **13**, 746-758.

Nicolini, M., A. C. Saulo, J. C. Torres, and P. Salio, 2002: Enhanced precipitation over southeastern South América related to strong low-level jet events during austral warm season. *Meteorologica*, **27**, 59-69.

Nieto-Ferreira, R. N., M. J. Suarez, and L. Tacaks, 2000: Two-way interacting regional climate model simulations for the Amazon region. Procedings of the 25th Annual Climate Diagnostics and Prediction Workshop, 23-27 October, 2000, Palisades, NY.

_____, T. M. Rickenbach, D. L. Herdies, and L. M. V. Carvalho, 2003: Variability of South American Convective Cloud Systems and Tropospheric Circulation during January-March 1998 and 1999. *Mon. Wea. Rev*., **131**, 961-973

Nogués-Paegle, J., C. R. Mechoso, and coauthors, 2002: Progress in Pan American CLIVAR Research: Understanding the South American Monsoon. *Meteorologica*, **27**, 3-30.

_____, L. A. Byerle, and K. C. Mo, 2000: Intraseasonal modulation of South American summer precipitation. *Mon. Wea. Rev*., **128**, 837–850.

_____, and K.-C. Mo, 1997: Alternating wet and dry conditions over South America during summer. *Mon. Wea. Rev*., **125**, 279-291.

_____, and K. Mo, 2002: Linkages between summer rainfall variability over South America and sea surface temperature anomalies. *J. Climate*. **15**, 1389–1407.

Rao, G. V., and S. Erdogan, 1989: The atmospheric heat source over the Bolivian plateu for a mean January. *Bound.-Layer-Meteor.*, **46**, 13-33.

Rao, V. B., I. F. A. Cavalcanti, and K. Hada, 1996: Annual variations of rainfall over Brazil and water vapor characteristics over South América. *J. Geophys. Res.*, **101**, 26539-26551.

Robertson, A. W., and C. R. Mechoso, 1998: Interannual and decadal cycles in river flows of southeastern South America. *J. Climate*, **11**, 2570-2581.

_____ and _____, 2000: Interannual and interdecadal variability of the South Atlantic Convergence Zone. Mon. Wea. Rev., 128, 2947-2957.

_____, J. D. Farrara, and C. R. Mechoso, 2003: Simulations of the atmospheric response to South Atlantic sea surface temperature anomalies. *J. Climate*, **16**, 2540-2551.

Ropelewski, C.F., and M.S. Halpert, 1987: Global and regional scale precipitation patterns associated with the El Niño/Southern Oscillation. *Mon. Wea. Rev.*, **115**, 1606-1626.

Saulo, A..C., M. Nicolini, and S.C. Chou, 2000: Model characterization of the South American Iow-level flow during the 1997-1998 spring-summer season. *Climate Dynamics*,**16**, 867-881.

Seluchi., M., and J. Marengo, 2000: Tropical-Mid Latitude Exchange of Air Masses during Summer and Winter in South America: Climatic aspects and extreme events, *Int. J. Climatol*, **20**, 1167-1190.

Seth, A., M. Rojas, 2003: Simulation and sensitivity in a nested modeling system for South America. Part I: Reanalyses boundary forcing. J. Climate, **16**, 2437-63.

Silva Días, M. A. F. 1987: Sistemas de mesoescala y previsão de tempo a curto prazo. *Rev. Bras. Meteor.*, **2**,133-150.

Silva Dias, P. L., J. P. Bonatti and V. E. Kousky, l987: Diurnally forced tropical tropospheric circulation over South America. *Mon. Wea. Rev.,***115**, 1465-1478.

Sorooshian, S., X Gao, K Hsu, R A. Maddox, Y. Hong, H. V. Gupta, and B. Imam, 2002: Diurnal Variability of Tropical Rainfall Retrieved from Combined GOES and TRMM Satellite Information. *J. Climate*, **15**, 983-1001.

Velasco, I., and J. M. Fritsch, 1987: Mesoscale Convective Complexes in the Americas. *J. Geophys. Res.*, **92**, 9591-9613

Vera, C. S., G. Silvestri, V. Barros, A. Carril, 2004: Differences in El Niño response over the Southern Hemisphere. *J. Climate*, **17**, 1741-1753.

_____, W. Higgins, and co-authors, 2004: A unified view of the American monsoon system. Submitted to *J. Climate*.

Virji, H., 1981: A preliminary study of summertime tropospheric circulation patterns over South America estimated from cloud winds. *Mon. Wea. Rev.*, **109**, 599-610.

Wainer, I., G. Clauzet, M.-P. Ledru, and E. Brady, 2004: Last Glacial Maximum in South America: Proxies and model results. Submitted to *Geophys. Res. Letters*.

Wang, M., J. Paegle, and S. DeSordi, 1999: Global variable resolution simulations of Mississippi Basin rains of summer 1993. *J. Geophys. Res*.,**104**, 19,399-19,414.

Wang, H., and R. Fu, 2002: Cross-Equatorial Flow and Seasonal Cycle of Precipitation over South America. *J. Climate,* **15**, 1591–1608.

____, and R. Fu, 2004: Influence of cross-Andes flow on the South American low-level jet. J. Climate, 17, 1247-1262.

Wheeler, M. and K. M. Weickmann, 2001: Real-Time Monitoring and Prediction of Modes of Coherent Synoptic to Intraseasonal Tropical Variability. *Mon. Wea. Rev.*, **129**, 2677–2694.

Yang, G.-Y. and J. Slingo, 2001: The Diurnal Cycle in the Tropics. *Mon. Wea. Rev.*, **129**, 784-801.

Zhou, J. and W. K.–M. Lau, 1998: Does a monsoon climate exist over South America. *J. Climate*, **11**, 1020-1040.

_____, and _____, 2001: Principal modes of interannual and decadal variability of summer rainfall over South America, *Int. J Climatol*., **21**, 1623-1644.

_____, and _____, 2002: Intercomparison of model simulations of the impact of 1997/98 El Niño on South American summer monsoon. *Meteorologica*, **27**, 99-116.