Toward a Unified View of the American Monsoon Systems

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ABSTRACT

An important goal of the Climate Variability and Predictability (CLIVAR) research on the American monsoon systems is to determine the sources and limits of predictability of warm season precipitation, with emphasis on weekly to interannual time scales. This paper reviews recent progress in the understanding of the American monsoon systems and identifies some of the future challenges that remain to improve warm season climate prediction. Much of the recent progress is derived from complementary international programs in North and South America, namely, the North American Monsoon Experiment (NAME) and the Monsoon Experiment South America (MESA), with the following common objectives: 1) to understand the key components of the American monsoon systems and their variability, 2) to determine the role of these systems in the global water cycle, 3) to improve observational datasets, and 4) to improve simulation and monthly-to-seasonal prediction of the monsoons and regional water resources. Among the recent observational advances highlighted in this paper are new insights into moisture transport processes, description of the structure and variability of the South American low-level jet, and resolution of the diurnal cycle of precipitation in the core monsoon regions. NAME and MESA are also driving major efforts in model development and hydrologic applications. Incorporated into the postfield phases of these projects are assessments of atmosphere–land surface interactions and model-based climate predictability experiments. As CLIVAR research on American monsoon systems evolves, a unified view of the climatic processes modulating continental warm season precipitation is beginning to emerge.

1. Introduction

Monsoon circulation systems, which develop over low-latitude continental regions in response to seasonal changes in the thermal contrast between the continent and adjacent oceanic regions, are a major component of continental warm season precipitation regimes. Both North and South America are characterized by such systems [hereafter referred to as the North American Monsoon System (NAMS) and the South American Monsoon System (SAMS), respectively]. The NAMS and SAMS provide a useful framework for describing and diagnosing warm season climate controls, and the nature and causes of year-to-year variability. A number of studies during the past decade have revealed the major elements of these systems, including their context within the annual cycle, and some aspects of their variability.

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Due in large part to the success of the World Climate Research Programme/Climate Variability and Predictability/Variability of the American Monsoon Systems (WCRP/CLIVAR/VAMOS) program, a unifying view of the NAMS and SAMS is beginning to emerge. In particular, CLIVAR/VAMOS has implemented complementary international programs in North and South America, namely the North American Monsoon Experiment (NAME) and the Monsoon Experiment South America (MESA). These programs have common objectives: 1) to understand the key components of the American monsoon systems and their variability, 2) to determine the role of these systems in the global water cycle, 3) to improve observational datasets, and 4) to improve simulation and monthly-to-seasonal prediction of the monsoon and of regional water resources.

Several recent publications summarize the accomplishments of the scientific community toward achieving a better understanding of the monsoon systems of the Americas. These include a review of the South American Monsoon System (Nogués-Paegle et al. 2002) and a review of the North American Monsoon System (Higgins et al. 2003). The intention of this paper is to discuss recent advances in our understanding of the NAMS and SAMS in an integrated framework, highlighting recent papers that illustrate progress made during the initial phase of CLIVAR. More complete historical reference lists are found in the overview papers cited above.

The basic features of the NAMS and SAMS and their variability within the context of the land surface–atmosphere–ocean annual cycle are discussed in sections 2 and 3, respectively. The role of land surface memory in the variability and predictability of the NAMS and SAMS is considered in section 4. Hydrologic characteristics of the monsoon systems are described in section 5. Outstanding science questions associated with gaps in our understanding are reviewed in section 6.

2. Basic features

Both the NAMS and the SAMS exhibit many of the features of their Asian counterpart, including large-scale land–sea temperature contrast, a large-scale thermally direct circulation with a continental rising branch and an oceanic sinking branch, land–atmosphere interactions associated with elevated terrain and land surface conditions, surface low pressure and an upper level anticyclone, intense low-level inflow of moisture to the continent, and associated seasonal changes in precipitation (both increases and decreases). Moreover, the poleward extension of the summer convection in the tropical Americas seems to be associated with similar mechanisms (such as the ventilation processes) to those acting in the Asian monsoon system (Chou and Neelin 2003).

Both the NAMS and the SAMS receive more than 50% of total annual precipitation during the respective summer monsoons, though the SAMS precipitation amounts are considerably greater (Figueroa and Nobre 1990; Higgins et al. 1997). The SAMS exhibits somewhat distinct characteristics compared to the other monsoon systems, given that most of South America is situated in the Tropics, and seasonal temperature differences are less pronounced than in subtropical monsoon regimes.

A clear annual cycle characterizes the convection over the tropical Americas that exhibits a seasonal regularity and degree of symmetry with respect to the equator (Horel et al. 1989). The NAMS and SAMS can be interpreted as the two extremes of the same cycle (Fig. 1) and their corresponding life cycle can be described using terms that have been traditionally reserved for the Asian summer monsoon system, namely, onset, mature, and decay phases.

Seasonal evolution of the convection

During May–June (the onset phase of the North American monsoon), heavy rains spread northward along the western slopes of the Sierra Madre Occidental (SMO; Fig. 2; Douglas et al. 1993; Stensrud et al. 1995; Adams and Comrie 1997). Precipitation increases over northwestern Mexico coincide with increased vertical transport of moisture by convection (Douglas et al. 1993) and southerly winds flowing along the Gulf of California. During this season, weather in the NAMS region changes abruptly from relatively hot, dry conditions to cool, rainy ones (Mock 1996; Adams and Comrie 1997).

Increases in precipitation over the southwestern United States occur abruptly around the beginning of July (e.g., Mock 1996; Higgins et al. 1997) and coincide with the development of a pronounced anticyclone at the jet stream level (e.g., Okabe 1995), the development of a thermally induced trough in the desert Southwest (Rowson and Colucci 1992), northward displacements of the Pacific and Bermuda highs (Carleton 1987), the formation of southerly low-level jets over the Gulf of California (Douglas 1995), and the formation of the Arizona monsoon boundary (Adang and Gall 1989). The onset of the summer monsoon rains over southwestern North America has been linked to an increase of rainfall along the East Coast of the United States (e.g., Higgins et al. 1997) and to a decrease of rainfall over the Great Plains of the United States (e.g., Douglas et al. 1993; Mock 1996; Higgins et al. 1997).

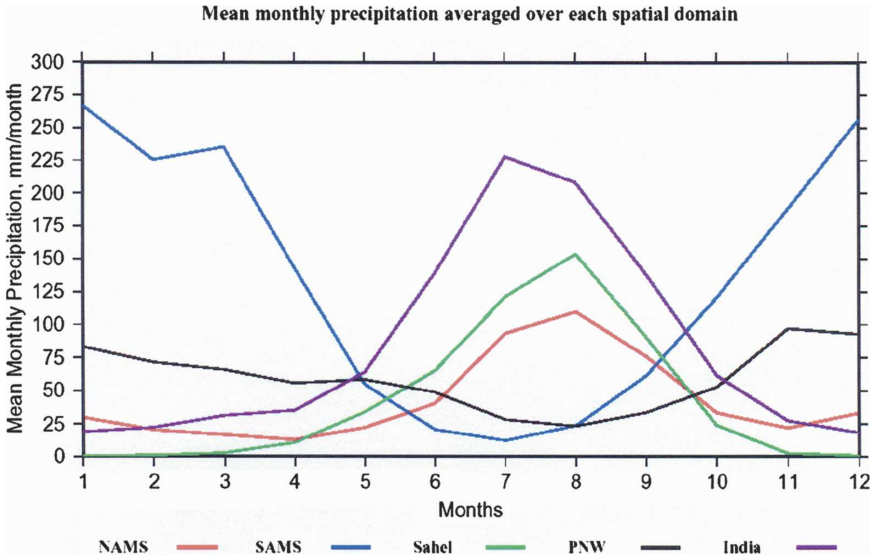

FIG. 1. Mean annual cycle of precipitation over several major monsoon areas [NAMS (20°–37°N, 248°–257°E); SAMS 40°–60°W); India (6°–37°N, 68°–98°E); Sahel (10°–20°N, 15°–15°W)]. For comparison, one nonmonsoon region with a large annual cycle is also shown [Pacific Northwest (PNW: 42°–50°N, 112°–124°W)].

Moisture transport onto the North American continent is accomplished by boundary layer flow from the south (Gulf of California), and in the middle troposphere via southeasterly flow from the Gulf of Mexico (Schmitz and Mullen 1996; Fig. 3a).

During July–August–early September, the NAMS is fully developed. The heaviest precipitation is west of the SMO and near the Bay of Campeche (Fig. 4b). The northern edge of the monsoon extends into Arizona and New Mexico (e.g., Douglas et al. 1993), but the rainfall is much lighter and more directly influenced by midlatitude effects. Precipitation also occurs in Central America and northwestern South America, with a relative maximum of precipitation in southeast (SE) subtropical South America (Fig. 4a). The diabatic heating released by the NAMS in combination with the orographic forcing induces descent over the eastern North Pacific while it promotes the development of the North Atlantic subtropical high (Rodwell and Hoskins 2001).

Surges of maritime tropical air move northward along the Gulf of California and are linked to bursts and breaks of the monsoon rains over the deserts of Arizona and California (Stensrud et al. 1995). Gulf surges are triggered by a variety of synoptic-scale and mesoscale disturbances, including tropical easterly waves, tropical cyclones, mesoscale convective systems, and upper-level inverted troughs (e.g., Stensrud et al. 1995; Higgins et al. 2004). Equatorward of the tropic of Cancer the monsoon exhibits a double-peak structure in precipitation and diurnal temperature range. From south-central Mexico into Central America, this midsummer dry spell or "canicula" is sufficiently regular as to appear in climatological averages (Magaña et al. 1999). Recent evidence indicates that the trade winds, evaporation, and precipitation patterns over the warm pool region to the southwest of Mexico modulate sea surface temperatures (SSTs) in a manner consistent with the double-peak structure in precipitation (Magaña et al. 1999).

During late September–October the decay phase of the NAMS occurs. The ridge over the western United States weakens as the monsoon high retreats southward and precipitation diminishes. Simultaneously the midlatitude westerly regime shifts equatorward, and precipitation events are more frequently associated with synoptic-scale frontal systems rather than with localized convective instability.

By September, the convection migrates from Central America into South America, and the onset of the wet season over South America starts first in the equatorial Amazon and then spreads quickly to the east and southeast (Fig. 2b). The onset across the Amazon basin lasts about one month (e.g., Kousky 1988; Horel et al. 1989; Marengo et al. 2001; Liebmann and Marengo 2001) and is followed by abundant rainfall. The onset (demise) of the Amazon rainy season is preceded by an increase in the frequency of northerly (southerly) cross-equatorial flow over South America (Marengo et al. 2001; Wang and Fu 2002). Changes in the moistening of the planetary boundary layer and changes in the tem-

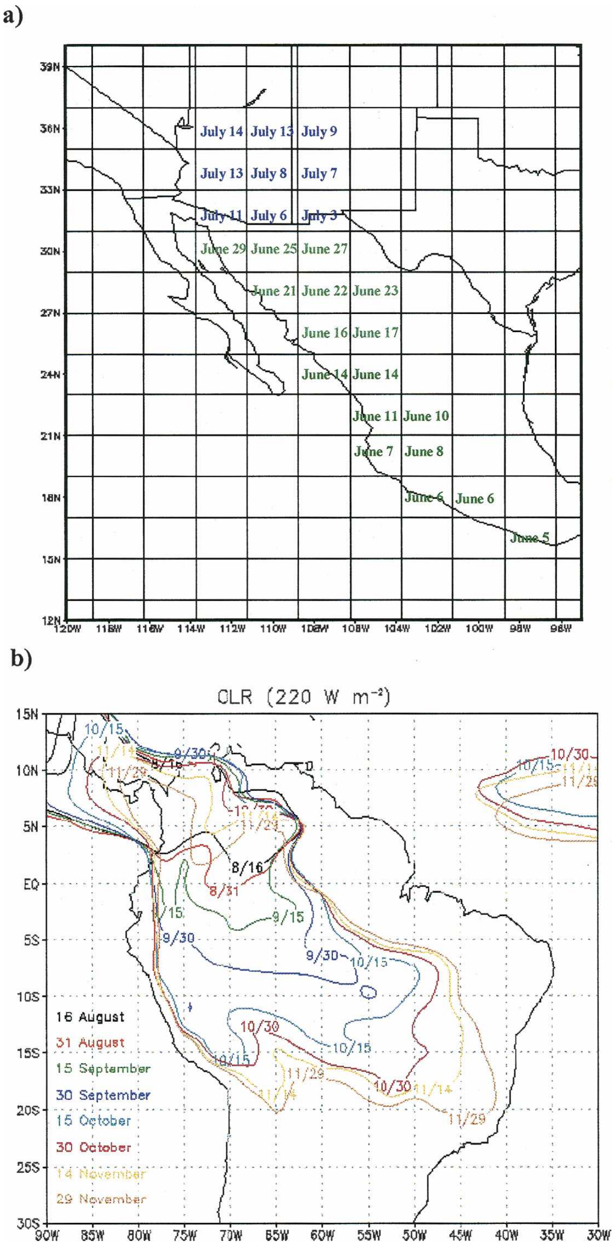

FIG. 2. Mean calendar date of onset for (a) the North American monsoon (from Higgins et al. 1999), and (b) the South American Monsoon (courtesy of V. Kousky).

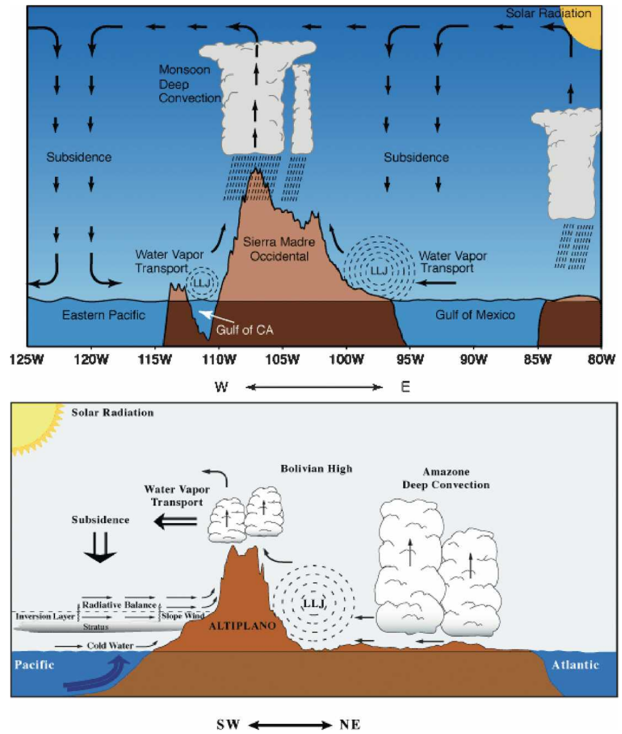

FIG. 3. (a) Schematic vertical (longitude–pressure) cross section through the NAMS at 27.5°N. Topography data were used to establish the horizontal scale and NCEP–National Center for Atmospheric Research (NCAR) reanalysis wind and divergence fields were used to establish the vertical circulations (from Higgins and NAME Science Working Group 2003). (b) Schematic vertical section across South America displaying the major large-scale elements affecting the SAMS (from CLIVAR Web site online at http://www.clivar.org).

perature at the top of the boundary layer seem to explain the seasonal changes of convection in tropical South America (Fu et al. 1999; Marengo et al. 2001; Liebmann and Marengo 2001).

The onset of the wet season in central and southeastern Brazil typically occurs between the end of September and early October and may sometimes occupy only a single 5-day period (Sugahara 1991). Intraseasonal oscillations may promote rapid onset over central Brazil (Vera and Nobre 1999). 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 intertropical convergence zone (ITCZ) is confined to the central Atlantic between 5° and 8°N (Zhou and Lau 1998).

During late November through late February (the mature phase of the SAMS), the main convective activity is centered over central Brazil and linked with a southeastward band of cloudiness and precipitation extending from southern Amazonia toward southeastern Brazil and the surrounding Atlantic Ocean (Fig. 4b). That convection band, known as the South Atlantic convergence zone (SACZ), is a distinctive feature of the SAMS (Kodama 1992). Also, the heavy rainfall zone extends over the Altiplano Plateau and the southernmost Brazilian highland.

The upper-level circulation during the South American summer includes well-defined regional circulation

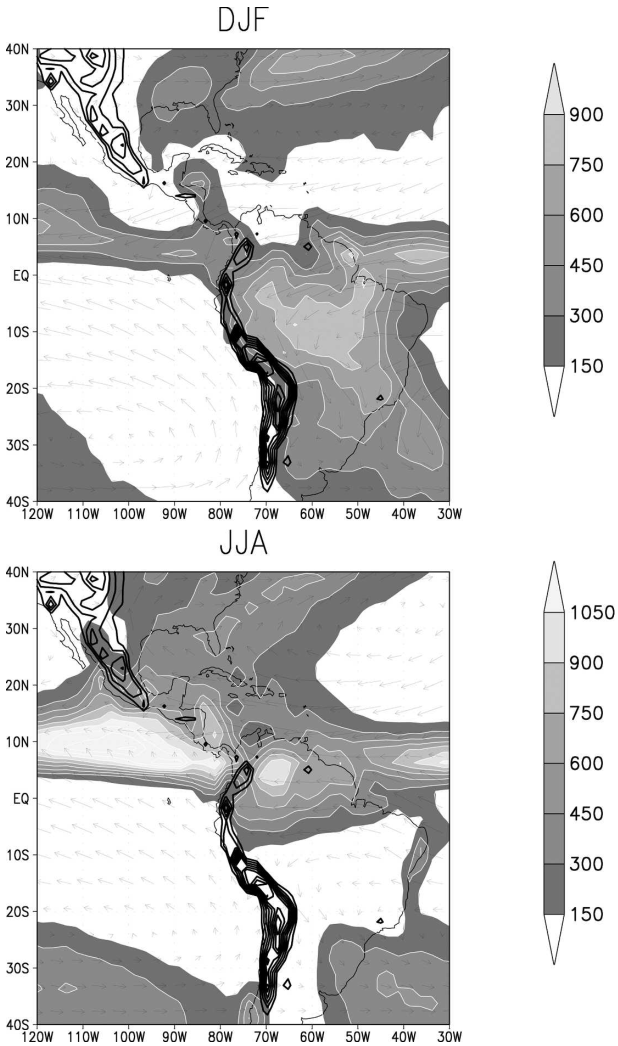

FIG. 4. Climatological mean accumulated precipitation (from Xie and Arkin 1997) and vertically averaged climatological mean moisture fluxes (from NCEP–NCAR reanalysis) for (a) DJF and (b) June–August (JJA). Contour interval is 150 mm. Vector units are kg (m s)$^{-1}$. Orography higher than 1000 m is contoured in black.

features, including a large anticyclonic circulation (the "Bolivian High") centered near 15°S, 65°W and an upper-level trough near the coast of northeast Brazil (not shown). At low levels, the "Chaco Low" is the most conspicuous feature downslope of the Andes and can be considered together with the Bolivian High as the regional response of the tropospheric circulation to the strong convective heating over the Amazon–central Brazil. The Andes effect reinforces the strength of the Chaco Low through the barrier role of the mountains (Gandu and Silva Dias 1998, and references therein). A continental-scale gyre transports moisture westward from the tropical Atlantic Ocean to the Amazon basin, and then southward toward the extratropics of South America (Fig. 4a). The diabatic heating released in the SAMS region seems to promote that gyre, and the maintenance of the South Atlantic subtropical high during austral summer (Rodwell and Hoskins 2001). A regional intensification of this gyre circulation to the east of the Andes Mountains is due to the South American low-level jet (SALLJ), with strongest winds in Bolivia near Santa Cruz (18°S, 63°W). The SALLJ transports considerable moisture between the Amazon and the La Plata basins and is present throughout the year (e.g., Berbery and Barros, 2002); it can be explained using simple, adiabatic models in which orography provides dynamical, rather than thermodynamical modification of the zonally averaged circulation (Byerle and Paegle 2002; Campetella and Vera 2002). During the warm season, thermodynamic processes associated with precipitation either over the SACZ region or southeastern South America (SESA) modulate the low-level flow in tropical regions (Berbery and Collini 2000).

Between March and May, the SAMS decay phase begins, as regions of heavy precipitation over the southern Amazon and central Brazil decrease and gradually migrate northwestward toward the equator and as the rainy season along the eastern coast of northeast (NE) Brazil gets under way and continues from April through June (Rao and Hada 1990). Throughout the decay phase of the SAMS, deep convection associated with the Atlantic ITCZ is relatively weak.

3. Climate variability

a. Diurnal and mesoscale variability

1) DIURNAL CYCLE

The timing of daily maximum convection in the Americas is location specific and closely related to topographic features such as mountain ranges and coast lines. In the NAMS there are large-scale shifts in the regions of deep convection during the day from over land during the afternoon and evening (especially along the western slopes of the SMO) to offshore locations during the morning hours (Figs. 5a,b; e.g., Higgins et al. 2003). In general, onshore flow begins during the morning transporting moist air from the Gulf of California inland and up the slopes of the SMO. Enhanced by inland mountain valley circulations, convection initiates near midday along the western slopes and high ridges of the SMO (Gochis et al. 2003; Fig. 6). During the evening and nighttime upslope flow reverses and generates a westward-propagating zone of convergence that migrates downslope across the coastal plains and out over

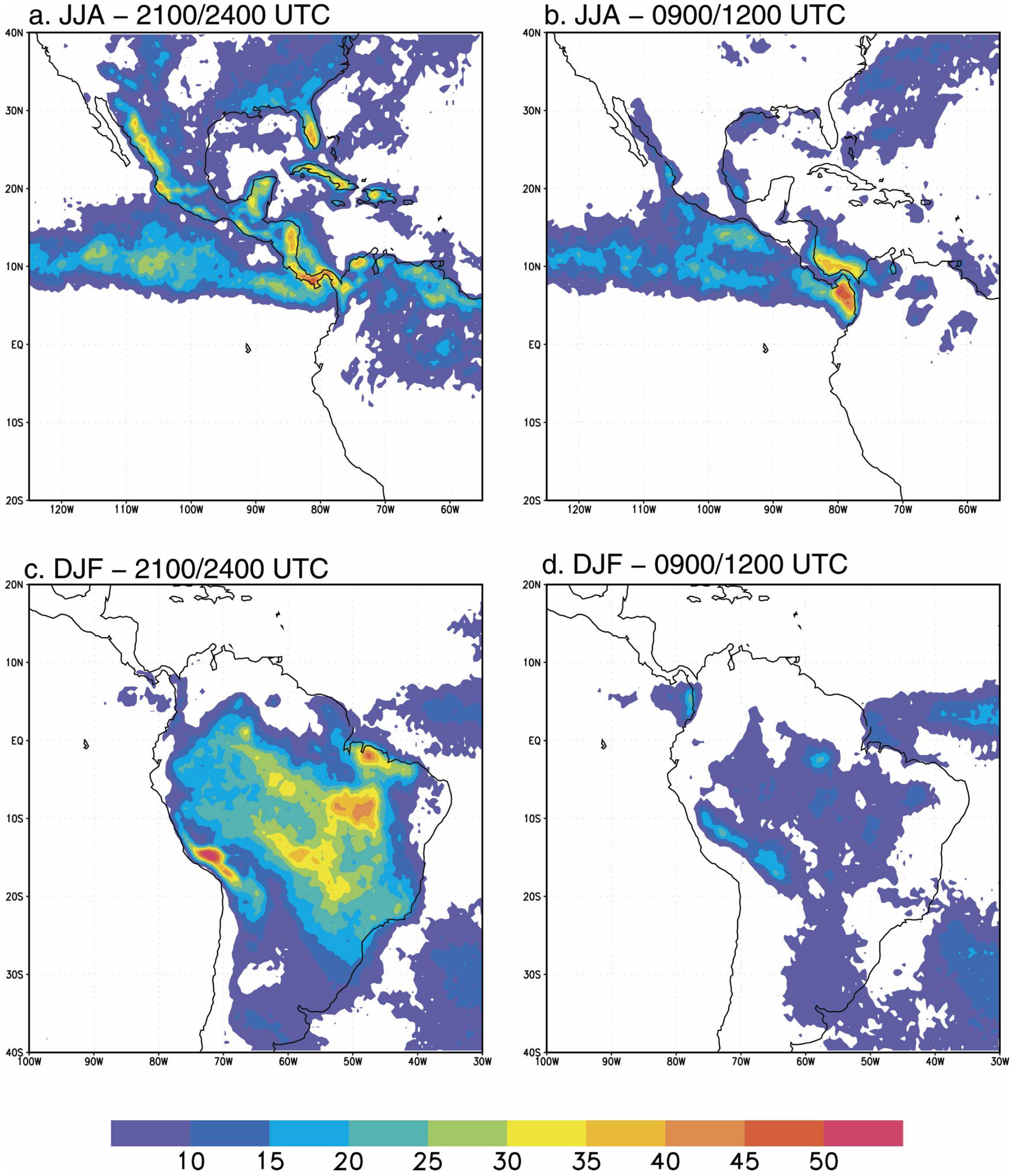

FIG. 5. Evening (2100–2400 UTC) temporal frequency of cold clouds (infrared brightness temperature Tb < 235 K) during the (a) boreal summer (JJA) and (c) austral summer (DJF). Pixel resolution is 0.5° × 0.5°. The period of analysis extends from 1983 to 1991. (b), (d) Same as in (a) and (c), respectively, but for late night and early morning (0900–1200 UTC).

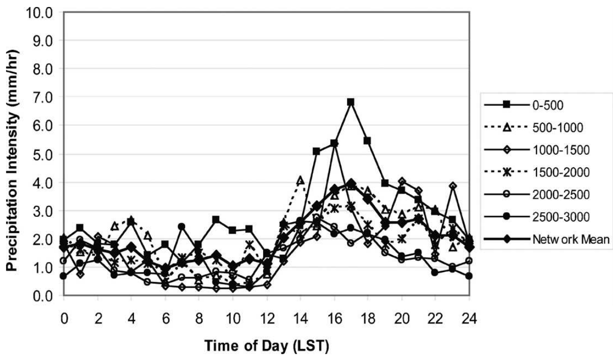

FIG. 6. Wet-day hourly rain rates for various elevation bands in the SMO from the NAME Event Raingauge Network (NERN; from Gochis et al. 2003).

the Gulf of California by early morning. Convection, in the form of small cloud systems, develops in coastal zones of the narrow Isthmus of Central America during the afternoon, and offshore nocturnal rainfall tends to propagate away through the nighttime and morning hours (Mapes et al. 2003b; Garreaud and Wallace 1997). The mechanism of propagation of the nocturnal rainfall appears to involve gravity waves (Mapes et al. 2003a).

Because of interaction with the slope of the Central American mountains, the diurnal cycle in Central America is strongly dependent upon the magnitude of the trade wind regime, which in turn is seasonally dependent. The Pacific slope of Central America shows a rainfall maximum occurring mainly in the afternoon and early evening, while on the Caribbean side, the diurnal cycle has a late evening–early morning precipitation maxima that changes with altitude, with the maximum occurring at altitudes between 600 and 2000 m in the early evening (Fernandez et al. 1996).

Satellite products have been used to document a coherent diurnal cycle of convective cloudiness over the South American continent during the summer rainy season [December–February (DJF)], and the peak is observed at around 1800 LST (Garreaud and Wallace 1997; Negri et al. 1994; Sorooshian et al. 2002). During the afternoon and evening convective clouds are particularly common over the Altiplano, along the northeast coast of the continent, and in two parallel bands over central Amazonia (Fig. 5c). The first two bands arise from the concurrent timing of the thermodynamic destabilization and the maximum strength of the plain-to-mountain wind convergence over the central Andes and land–sea breeze convergence along the coast, respectively, while the latter have been interpreted as the afternoon reactivation of inland-moving coastal squall lines (e.g., Cohen et al. 1995; Garreaud and Wallace 1997). Marengo et al. (2005) found that the westerly surface wind regime (related to enhanced SACZ) is associated with a precipitation maximum in southwest Amazonia during the afternoon (~1400 LST) followed by a relatively weak maximum in the evening. During the easterly regime, there is an early morning maximum (~0200 LST) followed by a stronger maximum in the afternoon (~1400 LST). Nighttime and early morning convection tends to be more prevalent along the eastern slope of the central Andes and subtropical plains (Fig. 5d; Nesbitt and Zipser 2003), and it can be ascribed to a decrease of the intensity of the compensating subsidence, which provides a fast linkage between convectively active and inactive regions (Silva Dias et al. 1987).

2) Mesoscale variability

While stability of the long-term precipitation climatology over the core monsoon regions is provided by stationary convective forcing mechanisms, regions peripheral to the core regions rely on mesoscale organization and propagation of convective storms or transient incursions of low-level moisture.

Transient disturbances including active or decaying tropical storms from either the eastern Pacific or the Gulf of Mexico, the passage of midlatitude troughs, westward-propagating cyclonic perturbations in the Tropics or various combinations of each of these phenomena perturb specific components of the time mean circulation thereby giving rise to increased mesoscale variability in Central and North America. This variability is manifested in phenomena such as pulsing of the Gulf of California low-level jet (Douglas 1995; Berbery and Fox-Rabinovitz 2003), intensification or reduction of the land–sea and valley–mountain circulations (Berbery 2001; Fawcett et al. 2002; Anderson et al. 2001) and Gulf of California moisture surges (Douglas 1995; Stensrud et al. 1997; Fuller and Stensrud 2000; Douglas and Leal 2003; Higgins et al. 2004). Some climatological features of long-lived, propagating, organized convection in the NAMS have been studied (e.g., McCollum et al. 1995), but the specific mechanisms that promote mesoscale organization and propagation remain elusive.

Velasco and Fritsch (1987) surveyed the mesoscale convective complexes (MCCs) over South and North America using Geostationary Operational Environmental Satellite (GOES) infrared imagery and found a large population of them over eastern South America between 20° and 40°S, especially during December and January. South American MCCs tend to develop during the evening and reach their maximum size after midnight, explaining the nocturnal maximum of convective cloud frequency in this region (Fig. 5c). In comparison to North America, midlatitude South American MCCs are significantly larger (average size of 500 000 km^2 at the time of maximum extent) and last longer (12 h on average). A second cluster of MCCs is found over the Gulf of Panama and along both slopes of the Colombian Andes (see also Mapes et al. 2003b). In contrast, Carvalho and Jones (2001) found that most convective systems (CSs) over Amazonia are relatively small (average size of less than 100 000 km^2), and short-lived (3–6 h), and, therefore, do not meet MCC identification criteria. Tropical Rainfall Measuring Mission (TRMM) satellite products reveal that wet season convection over the Amazon region exhibits a vertical structure intermediate to that observed over tropical oceans (i.e., less vertically developed, less lighting) and other continents (e.g., Nesbitt and Zipser 2003), although intraseasonal variability is large (Petersen et al. 2002). Deep convection is also organized in small CSs over the Altiplano (Garreaud 1999; Falvey and Garreaud 2004).

One of the primary elements associated with the occurrence of subtropical South American CSs is the SALLJ. National Centers for Environmental Prediction (NCEP) and European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis data suggest that the strongest winds are between 15° and 20°S, roughly over the Bolivian lowlands and Paraguay (Salio et al. 2002; Marengo et al. 2004). Recently, pilot balloon observations (Douglas et al. 1999) at Santa Cruz (18°S, 63°W, 30 km east of the Andean foothills) indicated that, when present, the SALLJ exhibits maximum wind speeds in excess of 15 m s^{-1} at about 1.5 km AGL and a vertical shear above that maximum of 5 m s^{-1} km^{-1} (Douglas et al. 1999). The associated maximum of meridional moisture flux occurs just to the east of the Andes, between the surface and 700 hPa, sharply decaying eastward within 500 km of the foothills (Saulo et al. 2000; Berbery and Collini 2000). Recently the South American Low-Level Jet Experiment (SALLJEX) was carried out in the region of sparse upper-air observations over central and subtropical South America during the austral summer 2002/03 (Vera 2004). Preliminary results using SALLJEX observations describe the significant diurnal variability of the SALLJ, with a deep, late afternoon–evening maximum that might partially explain the prevalence of nocturnal convection south of 20°S (Nicolini et al. 2004; Zipser et al. 2004)

b. Synoptic variability

Tropical cyclones (TCs), which account for between 20% and 60% of the rainfall along the Pacific coast of Mexico, also contribute as much as 25%–30% of the seasonal rainfall at western interior locations (Englehart and Douglas 2001), and they explain up to 15% of summer precipitation over the coastal areas in the United States (Larson et al. 2005). Tropical easterly waves propagating into the NAM region from the tropical Atlantic can develop into a TC in the eastern Pacific (Farfán and Zehnder 1997; Molinari et al. 2000). Idealized modeling studies have shown that interactions between the SMO, and easterly waves approaching from the Gulf of Mexico, and the ITCZ can lead to eastern Pacific cyclogenesis (Zehnder 1991; Zehnder et al. 1999; Mozer and Zehnder 1996). Wave amplification tends to occur in regions where the meridional gradient in low-level potential vorticity reverses (Molinari et al. 1997; Molinari and Vollaro 2000). The ITCZ west of the Central American coast can break down into multiple TCs because of a combined barotropic and baroclinic instability, dynamically supported by the reversal in the meridional gradient of potential vorticity (Nieto-Ferreira and Schubert 1997). Moisture surges, which often traverse the length of the Gulf of California, appear to promote increased convective activity in northwest (NW) Mexico and the southwest (SW) United States and are related to the passage of tropical easterly waves across western Mexico (Stensrud et al. 1997; Fuller and Stensrud 2000).

In contrast to the numerous studies on easterly waves in the tropical North Atlantic, there has not been as much consideration of such disturbances in the tropical South Atlantic and tropical South America. Easterly waves from the Atlantic in combination with the tropical heat source in the western Amazon, and with mesoscale features (sea-breeze and cloud-scale circulations), promote squall lines that propagate over the Amazon region (Cohen et al. 1995). Nevertheless, while the daily precipitation distribution over the northern part of South America during austral winter is mainly due to the activity of easterly wave-type tropical disturbances, daily precipitation variability over tropical South America during austral summer results from the combined action of equatorial trades, westward-traveling tropical disturbances, and the equatorward incursions of midlatitude synoptic wave systems (Kayano 2003).

Synoptic-scale disturbances, which are particularly large in the winter hemisphere but are also present during summer, often reach sufficiently low latitudes to affect the American monsoon systems (Fig. 7). Anderson and Roads (2002) investigated the characteristics of summer precipitation variability in the southwestern United States and found that midlatitude airmass intrusions interact with the monsoon ridge, thereby affecting the location and timing of precipitation events near the northern fringe of the NAMS. The synoptic patterns

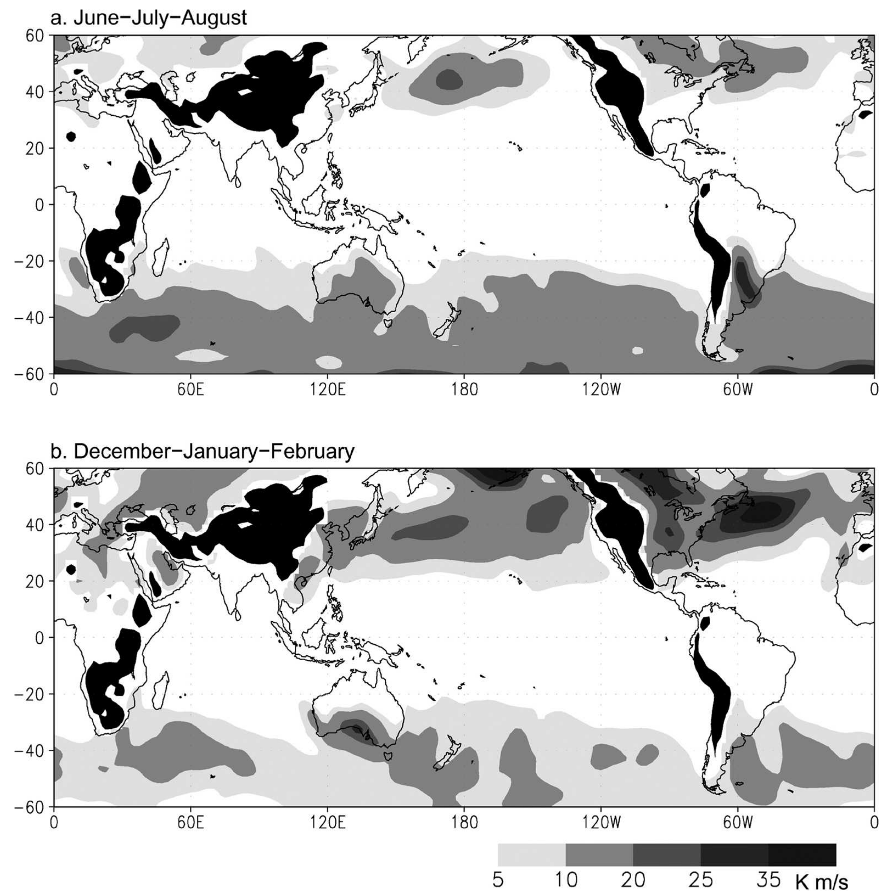

Fig. 7. Meridional heat flux by transient eddies at 925 hPa for (a) JJA and (b) DJF (K m s^{-1}). Black areas indicate terrain elevation in excess of 2000 m ASL.

associated with such shifts in the monsoon ridge, and associated synoptic modulation of severe thunderstorm activity, have been described by Maddox et al. (1995).

Over South America, the origin of the equatorward extension of synoptic activity has been attributed to the dynamical impact of the Andes cordillera on the synoptic-scale circulation. While upper-level synoptic systems tend to maintain their eastward motion as they cross the Andes at low levels, they experience a marked anticyclonic turning and equatorward elongation (e.g., Gan and Rao 1994; Vera et al. 2002). A schematic of the evolution of surface low and high pressure systems moving across South America shows that the persistent subtropical anticyclone over the SE Pacific limits the equatorward extension of low pressure systems approaching southwestern South America (Fig. 8a). As the low pressure system moves inland, strong midlevel westerly winds prevail over the subtropical Andes, resulting in surface troughing east of the Andes (Fig. 8b). The resulting northwestern Argentinean low (NAL; Seluchi et al. 2003) often merges with the monsoonal Chaco Low (centered at about 20°S). The rather unique thermal stratification of the NAL as well as its closed circulation structure was exceptionally diagnosed during SALLJEX, in agreement with previous modeling studies (Saulo et al. 2004). The transient deepening of the NAL favors the northerly transport of moist air to the east of the Andes (Nieto Ferreira et al. 2003; Seluchi et al. 2003), feeding severe storms over the subtropical continental plains or mesoscale convective complexes that contribute to further deepen the surface low as it moves eastward into the Atlantic through the latent heat released (Fig. 8c; Seluchi and Saulo 1998; Vera et al. 2002). Following the cyclone, dry air surges equatorward between the Andes and the Brazilian plateau (Fig. 8d; e.g., Garreaud 1999) that can be accompanied by synoptic-scale bands of deep convection at the leading edge, frequently associated with SACZ intensification (e.g., Garreaud and Wallace 1998).

c. *Intraseasonal variability*

During the NH warm season, the Madden–Julian oscillation (MJO; e.g., Madden and Julian 1994) influ-

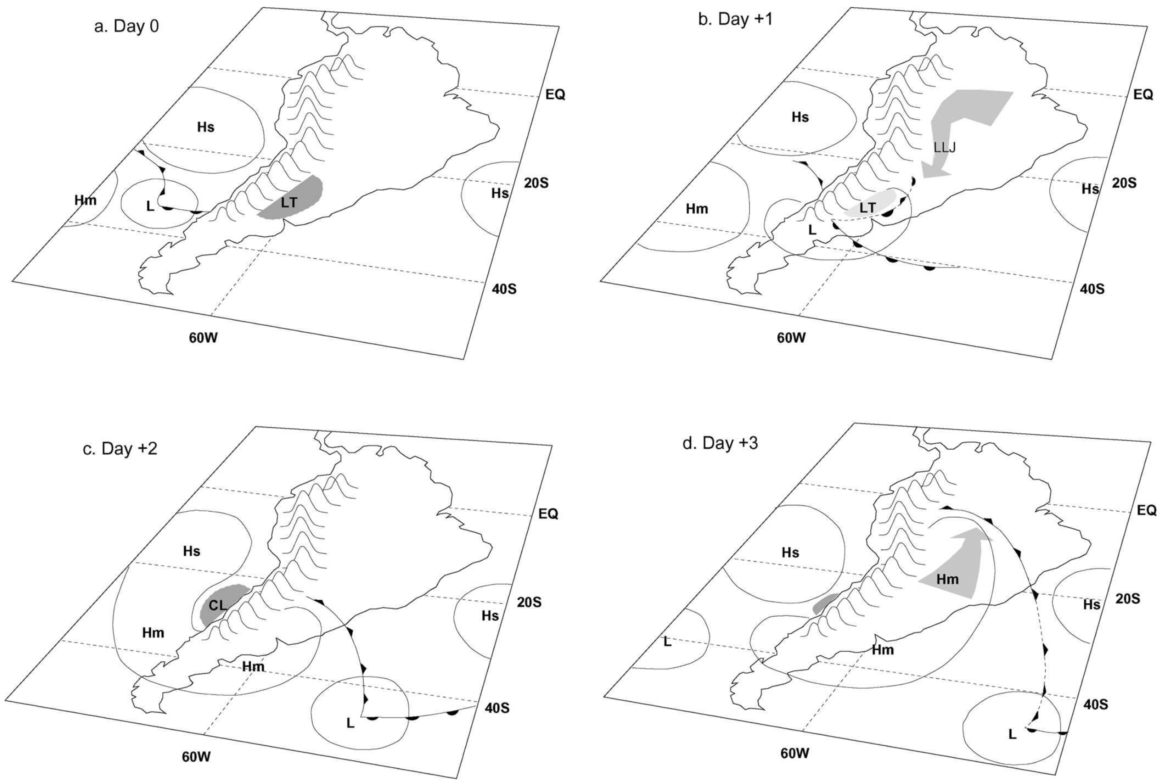

FIG. 8. Schematic representation of the life cycle of surface low and high pressure systems moving across South America. The symbols are as follows: Hs = subtropical high, Hm = migratory high, L = migratory low, and LT = lee trough.

ences a number of different weather phenomena affecting the North American monsoon. For example, there is a strong relationship between the MJO and TC activity in the Pacific and Atlantic Ocean basins (e.g., Liebmann et al. 1994; Maloney and Hartmann 2000). The strongest tropical cyclones tend to develop when the MJO favors enhanced precipitation and as the MJO progresses eastward, the favored region for TC activity also shifts eastward from the western Pacific to the eastern Pacific and finally to the Atlantic (Fig. 9; Higgins and Shi 2001). While this relationship appears robust, the MJO is one of many factors (e.g., SST conditions and vertical wind shear) that contribute to the development of TCs.

Evidence of active and quiescent periods of tropical easterly wave and Gulf of California surge activity in the NAM region motivated studies of relationships to the MJO. Higgins et al. (2004) showed that the relative location of the upper-level monsoon anticyclone in midlatitudes at the time of the gulf surge affects the response to the surge in the southwestern United States (Figs. 10a,b). Wetter-than-normal (drier-than-normal) conditions occur in the Southwest when the ridge axis locates to the east (west) of the region. The response of the ITCZ to the easterly wave forcing can result in gulf surge events, but the response may depend on the phase of the MJO. In particular, the westerly phase of the MJO may result in larger vorticity in the vicinity of the ITCZ, leading to a larger response (Zehnder et al. 1999).

The most distinctive pattern that characterizes intraseasonal rainfall variability over South America is a dipolar one (Casarin and Kousky 1986; Nogués-Paegle and Mo 1997). Enhanced precipitation over the SACZ is accompanied by decreased rainfall in the subtropical plains while the opposite phase is associated with increased southward moisture flux from the Amazon region, and increased rainfall in the subtropical plains (Figs. 10c,d; Diaz and Aceituno 2003). The dipole structure in precipitation is associated with distinct changes in the position and intensity of the Bolivian High (Vera and Vigliarolo 2000). Also, low-level zonal westerly (easterly) winds over tropical Brazil during summer are associated with an active (inactive) SACZ and net moisture divergence (convergence) over SESA, implying a weak (strong) SALLJ (Herdies et al. 2002; Jones and Carvalho 2002). Wave trains extending along the South Pacific link convective pulses in the SPCZ and SACZ regions (Grimm and Silva Dias 1995) and are associated with both phases of the rainfall dipole pat-

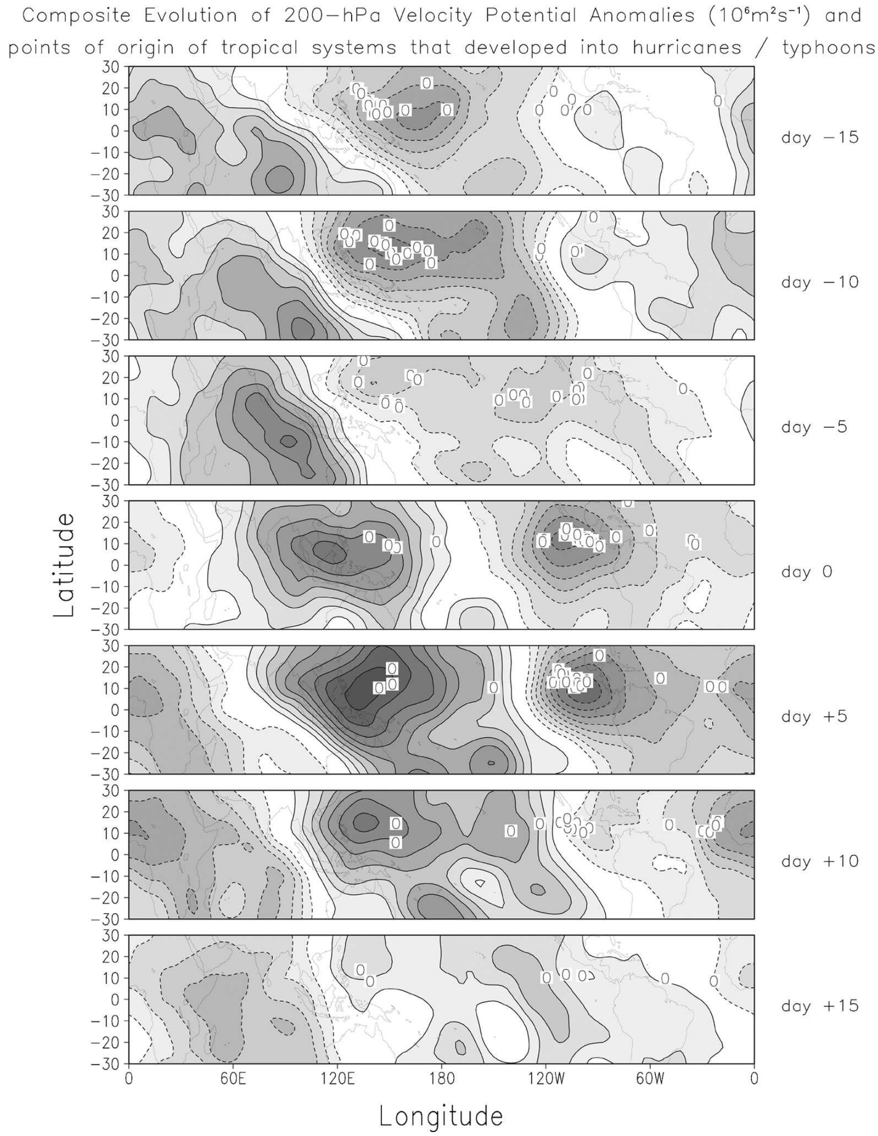

FIG. 9. Composite evolution of 200-hPa velocity potential anomalies (10^6 m^2 s^{-1}) associated with MJO events and points of origin of tropical disturbances that developed into hurricanes or typhoons. Contour intervals are 1×10^6 m^2 s^{-1}, solid contours indicate positive values (convection is suppressed), and dashed contours indicate negative values (convection is enhanced).

tern (right panels of Figs. 10c,d; e.g., Liebmann et al. 2004a).

Carvalho et al. (2004) showed that the MJO plays a role in modulating the persistence of intense SACZ events while Liebmann et al. (2004a) found that rain events in the SACZ (subtropical) region tend to occur 26 days prior to (2 days after) the peak in MJO convection at 10°S, 110°W. Nevertheless, the dipole structure of convection anomalies not only exhibits variability associated with the MJO, but also on time scales around 22–28 days (Liebmann et al. 1999; Nogués-Paegle et al. 2000), which implies that interaction among different intraseasonal fluctuations might control variability in the dipole pattern.

Relative influences of ENSO and the MJO on the monsoon must be separated and understood. Over the NAM region, ENSO-related impacts are linked to relatively broad north–south adjustments of the precipitation patterns in the eastern Pacific, while MJO-related impacts are linked to more regional north–south adjustments and potential predictability increases in the core monsoon region when both ENSO and the MJO are

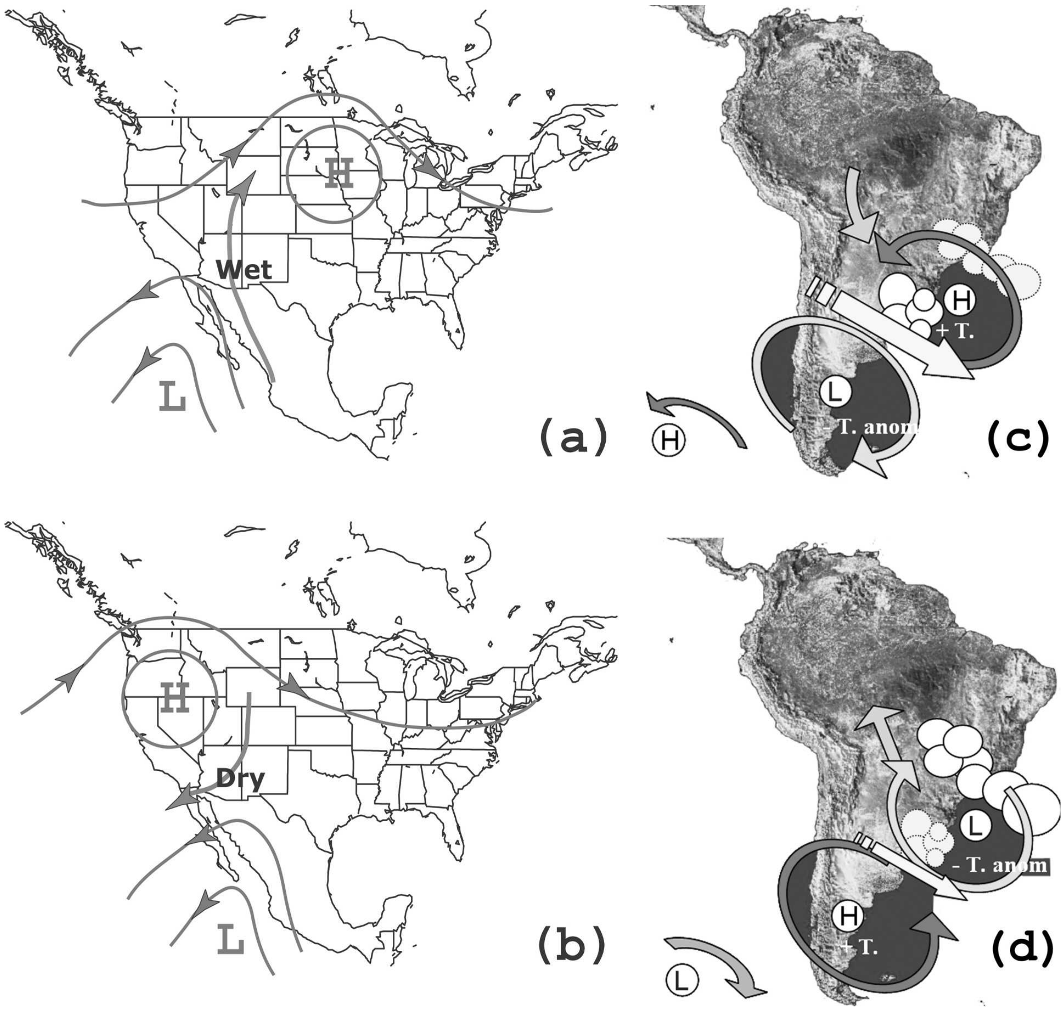

FIG. 10. Schematic of the typical 700-hPa circulation features of the NAMS (heights and winds) that accompany (a) wet and (b) dry surges keyed to Yuma, AZ. Regional circulation anomalies characterizing periods of (c) enhanced and (d) suppressed convective cloudiness over SESA during austral spring and summer. Letters H and L and the associated circulation arrows represent anticyclonic and cyclonic circulation anomalies, respectively. Here +T and −T stand for positive and negative temperature anomalies. The relative strength of the subtropical jet stream is represented by arrows of different sizes. The arrow eastward from the central Andes indicates the direction of the low-level wind anomaly. (From Diaz and Aceituno 2003.)

taken into account (Higgins and Shi 2001). When the seasonal ENSO phase is known, the explained variance in summer precipitation is concentrated in a zonal band near the ITCZ. However, if the seasonal MJO activity is also known, then there is a plume of higher explained variance that extends northward along the west coast of Mexico and into the southwestern United States. Recently Cazes-Boezio et al. (2003) found that ENSO teleconnection patterns observed over the South Pacific to South America in austral spring are due to changes in the frequency of occurrence and amplitude of intraseasonal circulation regimes.

There is some evidence that MJO activity might increase predictability over equatorial South America (Wheeler and Weickmann 2001; Mo 2001). Intraseasonal variations in the 10–90-day band may impact the skill of numerical weather forecasts over South America (Nogués-Paegle et al.1998) and the forecast skill seems to increase (decrease) during periods of strong (weak) convective activity associated with the MJO (Jones and Schemm 2000).

d. Interannual variability

Interannual variability of the American monsoon systems exhibits strong regional characteristics and within each monsoon domain there are coherent regional patterns in summer season rainfall anomalies (Comrie and Glenn 1998; Gutzler 2004). Out-of-phase relationships exist between precipitation in southwestern North America and that in the Great Plains of the United States (e.g., Higgins et al. 1999) and between precipitation over the east-central Amazon and northeast Brazil, and that over south Brazil, northeast Argentina, and Uruguay (Ropelewski and Halpert 1986; Marengo 2002; Nogués-Paegle et al. 2002; Gan et al. 2004; Grimm 2003).

Previous studies have considered five factors for the interannual variability of American monsoon precipitation: SST anomalies, land surface conditions, the position and strength of the tropical convergence zones, water vapor transport, and large-scale circulations. The last two factors are largely controlled by internal atmospheric dynamics. However seasonal anomalies in the large-scale circulation may be predictable to the extent that they are related to the boundary conditions. The fraction of total interannual variance explained by any one of the factors above is small, suggesting that predictability may be modest or that all of these factors (which actually are not independent) must be considered together, though their relative influences on predictability must be understood (Koster et al. 2000; Marengo et al. 2003).

1) Role of SST anomalies

SST effects on the monsoons can be both local [e.g., Gulf of California (GC), Gulf of Mexico (GM), Peruvian coast, and the southwestern Atlantic] and/or remote (e.g., tropical Pacific or tropical Atlantic), and in many cases the influences are modest. Carleton et al. (1990) showed that monsoon precipitation in the southwest United States is negatively correlated with SST anomalies along the northern Baja coast, while Huang and Lai (1998) found positive correlations with SST anomalies over the Gulf of Mexico. Roughly 80% of the rainfall in the Arizona/New Mexico region occurs after SST values in the northern Gulf of California exceed 28.5°C, which seems to be important for the timing, intensity, and extent of the monsoon (Mitchell et al. 2002). SST anomalies near the Peruvian coast may affect rainfall over South America by changing the land–sea temperature contrast (Yu and Mechoso 1999).

Remote SST anomalies in the equatorial Pacific associated with El Niño (La Niña) tend to be associated with anomalously dry (wet) North American summer monsoons in the core monsoon region over western and southern Mexico (e.g., Gutzler and Preston 1997; Higgins et al.1999; Hu and Feng 2002). Rainfall over the east-central Amazon and northeast Brazil (southeastern South America and central Chile) tends to be below (above) normal during the warm phase of ENSO (e.g., Moura and Shukla 1981; Pisciottano et al. 1994; Grimm et al.1998).

Tropical Atlantic SST anomalies have a strong influence on precipitation in tropical South America (e.g., Mechoso et al. 1990; Giannini et al. 2001, and references therein) and in particular the gradient of SST anomalies between the tropical North and South Atlantic appears to be the key element associated with rainfall anomalies over northeast Brazil and north-central Amazonia (Marengo 1992; Hastenrath and Greischar 1993; Uvo et al. 1998). Summer precipitation anomalies are correlated with SST anomalies in the western subtropical South Atlantic (WSSA; e.g., Diaz et al. 1998). Positive (negative) SST anomalies in the WSSA region are associated with moisture transport from tropical latitudes southeastward (eastward), and with positive rainfall anomalies over northeastern Argentina and southern Brazil (SACZ region; Doyle and Barros 2002; Robertson and Mechoso 2003).

2) Role of land surface conditions

Gutzler and Preston (1997) and Gutzler (2000) found that heavy snow in the southern Rockies tends to be associated with deficient summer rainfall in portions of the southwest United States and vice versa in analogy with the results of Yang et al. (1996) for the Asian monsoon. Higgins et al. (1998) argued that the effects of SST anomalies on cold season precipitation indirectly affects warm season rainfall, since they play a role in determining the initial springtime soil moisture conditions and vegetative cover, which in turn can feed back upon the climate during the warm months.

Modeling studies have demonstrated 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 (Goddard et al. 2003; Koster et al. 2000; Marengo et al. 2003). The SAMS onset is characterized by the transition of a polluted atmosphere (due to winter subsidence conditions and biomass burning) to a clear one during the rainy season. Biogenic aerosol and aerosol produced by biomass burning have a direct influence on the surface and tropospheric energy budget because of their capacity to scatter and absorb solar radiation (Artaxo et al. 1990). Recent modeling and observational results have indicated that the aerosol plume produced by biomass burning at the end of the dry season is transported to the south, where it may interact with frontal systems and possibly impact the precipitation regime (Freitas et al. 2005) through the radiative forcing of cloud microphysical processes (Silva Dias et al. 2002).

3) Role of tropical convergence zones

ENSO-related impacts on precipitation in the NAMS are strongly linked to meridional adjustments of the ITCZ (e.g., Yu and Wallace 2000; Higgins and Shi 2001). During El Niño the ITCZ shifts southward, resulting in an anomalously strong local meridional (Hadley) circulation and a reduction in seasonal mean rainfall over much of Mexico and the Caribbean (Higgins and Shi 2001; Hu and Feng 2002). During La Niña the ITCZ shifts northward, with attendant weakening of the local Hadley circulation and an increase in NAMS precipitation.

During austral summer, SST anomalies in the tropical Atlantic modulate the location of the ITCZ and thus precipitation over northeast Brazil (e.g., Nogués-Paegle and Mo 2002). Over the southwestern Atlantic, a dipolar structure of the SST anomalies accompanies interannual variations of the SACZ (Robertson and Mechoso 2000). Coupled atmosphere–ocean GCM simulations indicate that warm SST anomalies in the South Atlantic lead to a northward displacement and intensification of the SACZ. An intensified SACZ tends to induce underlying cooler SST anomalies (Chavez and Nobre 2004). Enhanced convective activity in the SACZ may lead to intensification of the Eurasian teleconnection pattern (Grimm and Silva Dias 1995).

4) Role of moisture transport

The Gulf of California and Great Plains low-level jets contribute to the summer precipitation and hydrology in the southwestern United States and Mexico and central United States, respectively (e.g., Bosilovich and Schubert 2002). In South America, the SALLJ east of the Andes shows a weak tendency to be stronger and more frequent when SST anomalies are positive in the tropical Pacific (Marengo et al. 2004). Also, numerical simulations and diagnostic studies have shown that the El Niño of 1998 featured more frequent and intense summer SALLJ episodes than the La Niña of 1999 (Douglas et al. 1999; Saulo et al. 2000). Nevertheless, there is considerable interannual variability of these LLJs and water vapor transport that is quite independent of ENSO (e.g., Hu and Feng 2001; Berbery and Barros 2002).

5) Role of the large-scale circulation

Interannual variations in large-scale teleconnection patterns, such as the Pacific–North American (PNA) pattern, are related to interannual variations in key components of the North American monsoon (Carleton et al. 1990), including the upper-tropospheric monsoon anticyclone over southwestern North America (Higgins et al. 1998). While some of this variability is directly linked to ENSO, at least a portion of it is related to the leading patterns of extratropical climate variability (e.g., the Arctic Oscillation), which are largely independent of ENSO (e.g., Higgins et al. 2000).

During warm (cold) ENSO phases, Rossby wave trains emanating from either the western or central Pacific toward SH high latitudes [Pacific–South America (PSA) patterns] maintain a distinctive cyclonic (anticyclonic) circulation anomaly in the vicinity of South America that enhances (suppresses) precipitation over SESA (e.g., Grimm et al. 2000). Despite the fact that the ENSO signal projects strongly on the PSA patterns, numerical simulations show that they can also be generated by internal atmospheric dynamics (Cai and Watterson 2002). Recently Vera et al. (2004) showed that differences in the teleconnection patterns in the SH among different ENSO warm events are related to changes in the intensity and location of the SPCZ as well as to differences in the activity of the three leading modes of circulation variability in the SH, independent of ENSO. Silvestri and Vera (2003) have shown that the Antarctic Oscillation, which is largely independent of ENSO, influences interannual precipitation changes over subtropical South America (especially during austral winter and spring).

e. Variability on decadal (and longer) time scales

The summer monsoon in the southwest United States seems to be modulated by longer-term (decade scale) fluctuations in the North Pacific SSTs associated with the Pacific decadal oscillation (PDO; Higgins and Shi 2000). The link between the North Pacific wintertime SST pattern and the summer monsoon appears to be established via the impact of variations in the Pacific jet on west coast precipitation regimes during the preceding winter. This affects local land-based sources of memory in the southwestern United States, which in turn influence the subsequent timing and intensity of the summer monsoon (e.g., Castro et al. 2001).

A number of studies have reported the existence of decadal and longer time-scale variability in South American rainfall, related to ocean surface changes on those time scales in both the Pacific and the Atlantic Oceans (Robertson and Mechoso 2000; Zhou and Lau 2001). Interdecadal variability of around a 15-yr period has been observed in SACZ activity, in southwest Atlantic SSTs and river flows of the La Plata Basin (Robertson and Mechoso 2000). Positive correlations have been found between SST trends in the tropical South Atlantic (Fig. 11c) and precipitation anomalies in northeast Brazil (Wagner 1996), Amazonia (Marengo 2004)) and SESA (Fig. 11a; Liebmann et al. 2004b). Marengo et al. (2004) showed positive trends in SALLJ activity (Fig. 12a) consistent with trends in the North Atlantic Oscillation (NAO; Fig. 12d) and the PDO (Fig. 12e). In particular, the climate shift that occurred in the Pacific around the late 1970s (Mantua et al. 1997) is significantly related with an increase of SALLJ activity (Marengo 2004) as well as with an increase in precipitation over SESA.

4. Land surface variations in the American monsoon regions

Land use/land cover changes are occurring in the core region of the both NAMS and SAMS regions. Overgrazing within the steep, semiarid terrain of the SMO has resulted in dramatic changes in surface vegetation and soil characteristics, which, in turn, produce several important hydrological responses that have not yet been assessed (Viramontes and Descroix 2003). There also exists, along the western slope of the SMO, large expanses of deciduous tropical forests, which experience a dramatic green-up cycle immediately following the onset of the monsoon. It has been suggested that following this green-up these large expanses of forest may serve as critical pathways for the recycling of monsoon precipitation back to the atmosphere, although the effects of such activities on the regional climate are still open to question.

The Amazon basin is the world's most extensive tropical rain forest and more than one-third of the world's woody biomass is located in South America, with 27% in Brazil alone. However, the rate of deforestation in Central and South America is among the highest in the world at an annual average rate of 0.48%. From the last 30 yr to present, the conversion of forest land to other uses (such as agricultural activities) has been causing the deforestation and degradation of the forest ecosystem. Since the mid-1980s, atmospheric

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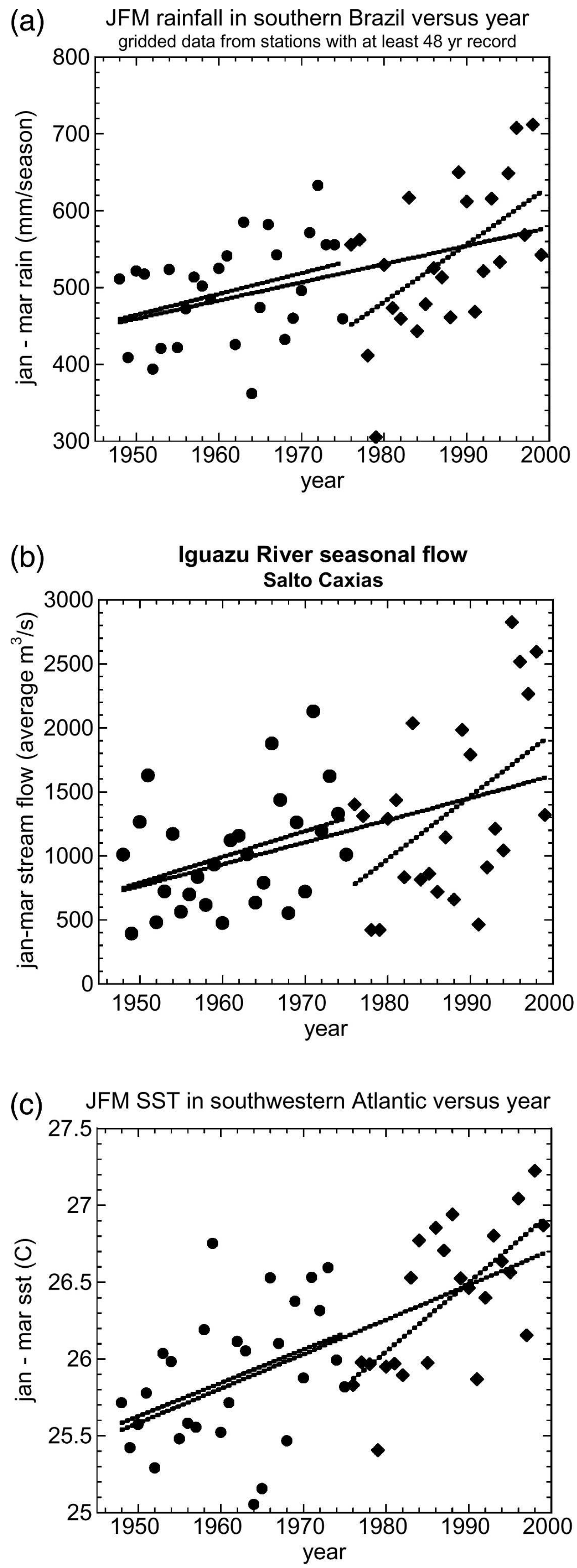

Fig. 11. JFM rainfall (mm season^{-1}) over southern Brazil from station data vs year. Filled circles represent years 1948–75. Filled squares represent years 1976–99. Dashed curves represent linear, least squares fits for segments 1948–75, 1976–99, and 1948–99. (b) Same as in (a) except ordinate is average JFM discharge at Salto Caxias, Brazil, on the Iguazú River, near the confluence with the Paraná. (c) Same as in (a) except that ordinate is an index of average SST (21.9°–27.6°S on a Gaussian grid, 28.2°–41.2°W; from Liebmann et al. 2004b).

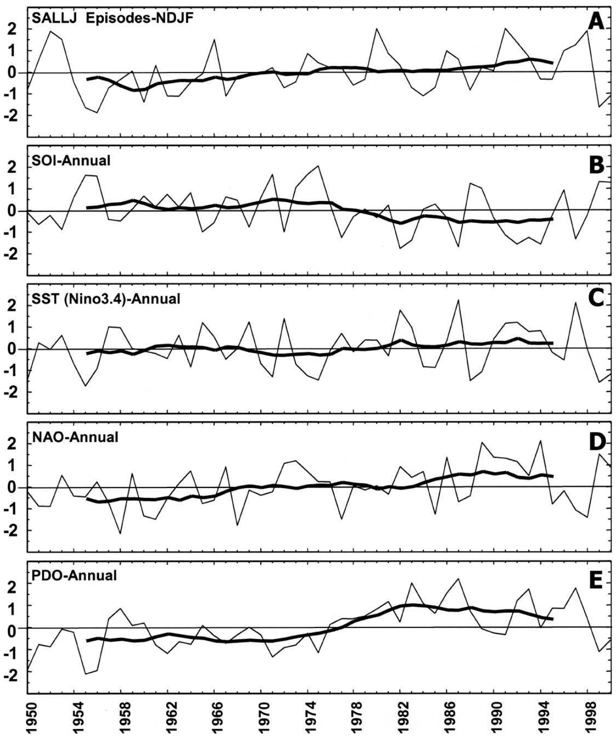

FIG. 12. Normalized annual departures of (a) SALLJ event annual counts, (b) Southern Oscillation index (SOI), (c) El Niño-3.4 SSTs, (d) NAO, and (e) PDO indexes. Thick lines represent the 11-yr moving averages.

GCM simulations have suggested that deforestation in tropical South America can cause a significant decrease in precipitation and evapotranspiration and a decrease in moisture convergence, especially in central and northern Amazonia (e.g., Marengo and Nobre 2001). New developments in dynamic vegetation schemes and coupled climate–carbon models (Cox et al. 2000; Betts et al. 2004; Huntingford et al. 2004) have shown that the physiological forcing of stomatal closure can contribute 20% to the rainfall reduction in the Amazon associated with rising atmospheric CO_2 levels. Nevertheless, the influence of land surface variations in explaining the slightly positive rainfall trends documented in southern Amazonia since the mid-1970s (Marengo 2004) and the significant positive trends detected in rainfall and streamflow over the subtropical portion of the La Plata Basin (Figs. 11b,c; Genta et al. 1998; Liebmann et al. 2004b) is still unclear. It has been suggested that an increase in southern Amazonia rainfall could be due to mesoscale circulations triggered by the deforestation (Silva Dias et al. 2002).

5. Monsoon hydrology

From the standpoint of land surface hydrology, monsoon regions and systems have several characteristics relevant for the land surface hydrology. A comparison between the mean monthly runoff as a fraction of the annual total precipitation (Fig. 13) and the mean monthly runoff as a fraction of average annual total precipitation (Fig. 14) shows that runoff is a more important process in wet regions. The average annual runoff ratio for the Little Colorado River (NAMS) is only about 0.1, while that for the Mahaweli Ganga (India region) is around 0.5. Furthermore, the relatively modest winter (nonmonsoon) precipitation in the NAMS leads to considerably higher runoff, on average, than does the monsoon precipitation, primarily because of reduced evapotranspiration in winter, and the contribution within small portions of the NAMS domain (e.g., the Little Colorado River) of winter snow that contributes disproportionately to winter and spring runoff. The figures also show that there are strong differences in the seasonality of runoff and streamflow across the regions as well. In the Sahel, precipitation and runoff are very low in months outside the monsoon period, whereas in India, there is significant runoff outside the monsoon period.

In the SAMS the seasonal signature of runoff is extremely damped and is only modestly higher in the months following the precipitation peak. The rivers in the southern Amazon basin, extended over the northern part of the SAMS core region, exhibit their maximum discharges around 2–3 months after the rainfall peaks, while those in the northern Amazon basin are more sensitive to the interannual variability linked to strong ENSO events (Marengo 2004). On the other hand, the upper Paraná basin located over the central and southern portion of the SAMS region exhibits maximum discharges during February and March associated with copious summer rainfall in the SACZ (Camilloni and Barros 2000).

Therefore, there are important similarities in the land surface hydrology of most of the monsoon regions (strongly seasonal runoff, but with peak runoff ratios much lower than in a strongly seasonal nonmonsoon environment, primarily due to high evaporative demand during the maximum precipitation season). However, there are major differences as well. In the arid to semiarid monsoon environments (NAMS and Sahel) runoff ratios are much lower than in the humid SAMS and India regions, which also have more uniform runoff over the year. Again, while these differences are attributable in considerable part to the characteristics of the land surface forcing (especially precipitation), some of the differences, especially in the seasonality and temporal variability of runoff, are due to catchment-specific topographic, geological, and vegetation characteristics.

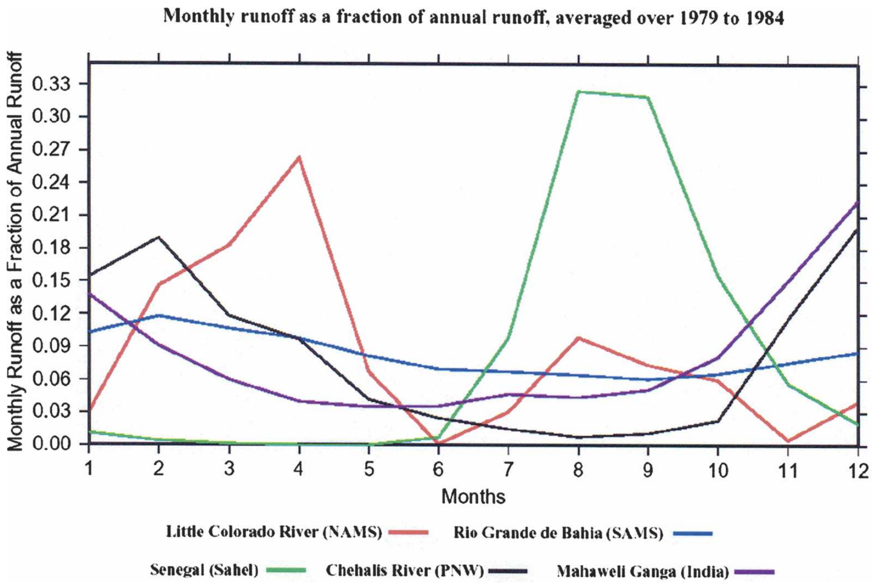

FIG. 13. Normalized (by annual average) mean monthly runoff for selected rivers in each basin. NAMS [river (Little Colorado near Cameron, AZ), drainage area (68 634 km^2), (lat, lon) = (35.92°N, 111.57°W)]; SAMS [river (Rio Grande de Bahia near Boqueirao, Brazil), drainage area (65 900 km^2), (lat, lon) = (11.33°S, 43.80°W)]; India [river (Mahaweli Ganga near Manampitiya, Sri Lanka), drainage area (7343 km^2), (lat, lon) = (7.92°N, 81.08°E)]; Sahel [river (Senegal River near Bakel, Senegal), drainage area (218 000 km^2), (lat, lon) = (14.90°N, 12.45°W)]; PNW [river (Chehalis River at Porter, WA), drainage area (3351 km^2), (lat, lon) = (46.94°N, 123.31°W)].

6. Future challenges

Despite significant advances in our understanding of the American monsoon systems, many challenges remain. At local scales, additional attention to the low-level circulation and precipitation patterns is needed. Some principal research questions in common to both the NAMS and the SAMS include the following: 1) What are the relationships between local low-level circulation features (e.g., the low-level jets; mountain–valley circulation) and the diurnal cycle of moisture and convection? 2) What are the dominant sources of precipitable moisture for monsoon precipitation? 3) What are the relative roles of local variations in sea surface temperature and land surface parameters (topography, soil moisture, and vegetation cover) in modulating precipitation?

The diurnal cycle (and related processes and feedbacks) is poorly represented in models not only over the core monsoon regions, but also at most locations over the Americas during the warm season. The various atmospheric pathways for moisture transport into the Americas should be identified and understood. An improved characterization of deep convection will play a critical role in accurate depiction of moisture transport over the continents. The relative contributions of local oceanic forcing, soil moisture, and vegetation to warm season precipitation variability in the monsoon regions are not known. At present, we know relatively little about land surface conditions and fluxes over most of the NAMS and SAMS regions, and characterizing the surface fluxes is critical for understanding and simulating the hydrologic cycle. The role of the transport of aerosols produced by biomass burning in the Amazon region on the SAMS variability should also be further explored and basic questions that need further investigation are as follows: What is the comparative impact of remote climate forcing and of the regional effects of biomass burning generated aerosol in the beginning of the rainy season? Is the local effect of aerosol more significant through a radiative impact or through cloud microphysics?

On regional scales, an improved description and understanding of intraseasonal aspects of the monsoon, especially those related to the MJO is needed. Some general questions include the following: 1) What is the nature of the relationship between the MJO and key components of the monsoon systems (e.g., low-level jets, gulf surges, etc.)? 2) What are the relationships between the MJO and extreme weather events in the Americas? 3) What are the relative influences of the MJO and ENSO on monsoon precipitation?

Of practical significance is the fact that there is a

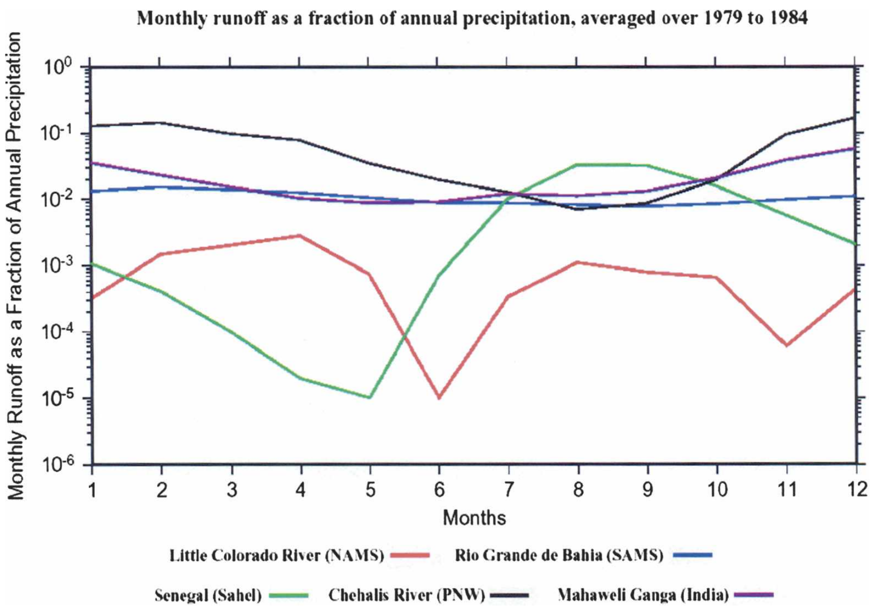

FIG. 14. Monthly runoff ratios (monthly average streamflow as a fraction of average of annual precipitation) for five river basins listed in the caption of Fig. 13.

coherent relationship between the phase of the MJO and the precipitation pattern around the global Tropics. Additional studies are needed to determine how the phase of the MJO relates to the frequency and intensity of hurricanes and tropical storms, what fraction of the summer rainfall over the Americas is due to tropical storms and hurricanes, and whether interannual-to-interdecadal variability of the MJO contributes to tropical cyclone frequency and intensity. While ENSO-related impacts on the monsoons are reasonably well documented, MJO-related impacts, as well as the relative influences of the MJO and ENSO, are not well understood.

At continental scales, the focus should be on improved description and understanding of the spatial and temporal linkages between warm season precipitation, the large-scale circulation patterns, and the dominant boundary forcing parameters (both land and ocean). Some general scientific questions are as follows: 1) How is the evolution of monsoonal precipitation related to the seasonal evolution of the oceanic and continental boundary conditions? 2) What are the relationships between interannual variations in the boundary conditions, the atmospheric circulation, and the continental hydrologic regime? 3) What is the correlation between the anomaly sustaining atmospheric circulation and the land and ocean surface boundary conditions that characterize precipitation and temperature anomalies during the summer?

Studies that enhance our dynamical understanding of the seasonal march of the monsoon and its variability are critical. The relative importance of the land and ocean influences on monsoon precipitation changes with the seasons as does the potential predictability of circulation and precipitation anomalies. The remote influences of other low-frequency modes of variability besides ENSO (e.g., PDO) on the warm season precipitation regimes over the Americas may be important. There is evidence that potentially predictable anomalies of soil moisture, snow cover, and vegetation play a role in the seasonal variability of precipitation patterns, and it has been suggested that there are important feedbacks (of either sign) between the atmosphere and land surface. Prediction of the detailed distribution of continental precipitation is a challenging task since it requires the skillful modeling of the subtle interplay between land surface and oceanic influences such as the complicating influences of terrain and coastal geometry.

The availability of comprehensive datasets that document key features of the American monsoons is crucial. Depiction of the diurnal cycle of precipitation and atmospheric circulation over the NAMS and SAMS regions and long-term monitoring of oceanic and land surface conditions are but two examples of the types of datasets that are needed.

MESA and NAME

The WCRP/CLIVAR/VAMOS Program is implementing two international monsoon programs in the

Americas: the Monsoon Experiment South America (MESA) and the North American Monsoon Experiment (NAME), aimed at improving warm season precipitation forecasts over the Americas on seasonal-to-interannual time scales. To help MESA and NAME achieve this goal, CLIVAR and the Global Energy and Water Cycle Experiment (GEWEX) are implementing empirical and modeling studies plus dataset development and enhanced monitoring activities, and several field campaigns [e.g., SALLJEX, NAME 2004, and La Plata Basin Program (PLATIN)]. As a consequence, these programs will deliver improved infrastructure to monitor and predict the American monsoons, more comprehensive understanding of climate variability and predictability, strengthened multinational scientific collaboration across Pan America, and improved climate models that predict monsoon variability months to seasons in advance.

To meet these challenges, the VAMOS Program is developing a modeling strategy that incorporates both MESA and NAME activities. The strategy includes baseline seasonal simulations that correspond to major field campaigns and multiyear simulations focused on key physical processes (e.g., the diurnal cycle of convection). Preliminary SALLJEX modeling coordinated activities, which included participation by groups from South America and the United States, have provided useful hints to understanding and improving the poor model performances in simulating precipitation associated with the MCSs over central South America (Paegle et al. 2004). The North American Monsoon Assessment Project (NAMAP; Gutzler et al. 2004) proved to be an excellent way to benchmark and assess current global and regional model simulations of the North American monsoon. In addition, joint experimental prediction (e.g., sensitivity to soil moisture and SST) and product development (e.g., hazards assessments and drought monitors) activities are in the planning stages. Additional information about MESA and NAME is available on the VAMOS Web site (http://www.clivar.org/organization/vamos/vamos.php; VAMOS datasets are available at http://www.joss.ucar.edu/vamos/data/).

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