

The Central American Midsummer Drought: Regional Aspects and Large-Scale Forcing*

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ABSTRACT

The midsummer drought (MSD) is a diminution in rainfall experienced during the middle of the rainy season in southern Mexico and Central America, as well as in the adjacent Caribbean, Gulf of Mexico, and eastern Pacific seas. The aim of this paper is to describe the regional characteristics of the MSD and to propose some possible forcing mechanisms. Satellite and in situ data are used to form a composite of the evolution of a typical MSD, which highlights its coincidence with a low-level anticyclone centered over the Gulf of Mexico and associated easterly flow across Central America. The diurnal cycle of precipitation over the region is reduced in amplitude during midsummer. The MSD is also coincident with heavy precipitation over the Sierra Madre Occidental (part of the North American monsoon). Reanalysis data are used to show that the divergence of the anomalous low-level flow during the MSD is the main factor governing the variations in precipitation. A linear baroclinic model is used to show that the seasonal progression of the Pacific intertropical convergence zone (ITCZ), which moves northward following warm sea surface temperature (SST) during the early summer, and of the Atlantic subtropical high, which moves westward, are the most important remote factors that contribute toward the low-level easterly flow and divergence during the MSD. The circulation associated with the MSD precipitation deficit helps to maintain the deficit by reinforcing the low-level anticyclonic flow over the Gulf of Mexico. Surface heating over land also plays a role: a large thermal low over the northern United States in early summer is accompanied by enhanced subsidence over the North Atlantic. This thermal low is seen to decrease considerably in midsummer, allowing the high pressure anomalies in the Atlantic and Pacific Oceans to extend into the Gulf of Mexico. These anomalies are maintained until late summer, when an increase in rainfall from the surge in Atlantic tropical depressions induces anomalous surface cyclonic flow with westerlies fluxing moisture from the Pacific ITCZ toward Central America.

1. Introduction

The Central Americas and surrounding oceans have a pronounced annual cycle in precipitation with a rainy season typically from May to October (Fig. 1). How-

ever, as pointed out by Magana et al. (1999) (see also references therein), the rainy season is punctuated by a relatively dry period in July and August, termed the Central American midsummer drought (MSD). The magnitude of this event varies from year to year [Fig. 1, see also Magana et al. (1999)] but is robust enough to appear in the climatological average (Fig. 2). Although the MSD is not a true drought in the sense of a reduction to near-zero rainfall over an extended period, it does represent a significant diminution in rainfall of up to 40%. The timing and magnitude of the MSD is important to farmers in this region where agriculture is an important part of the economy and many rely on subsistence farming.

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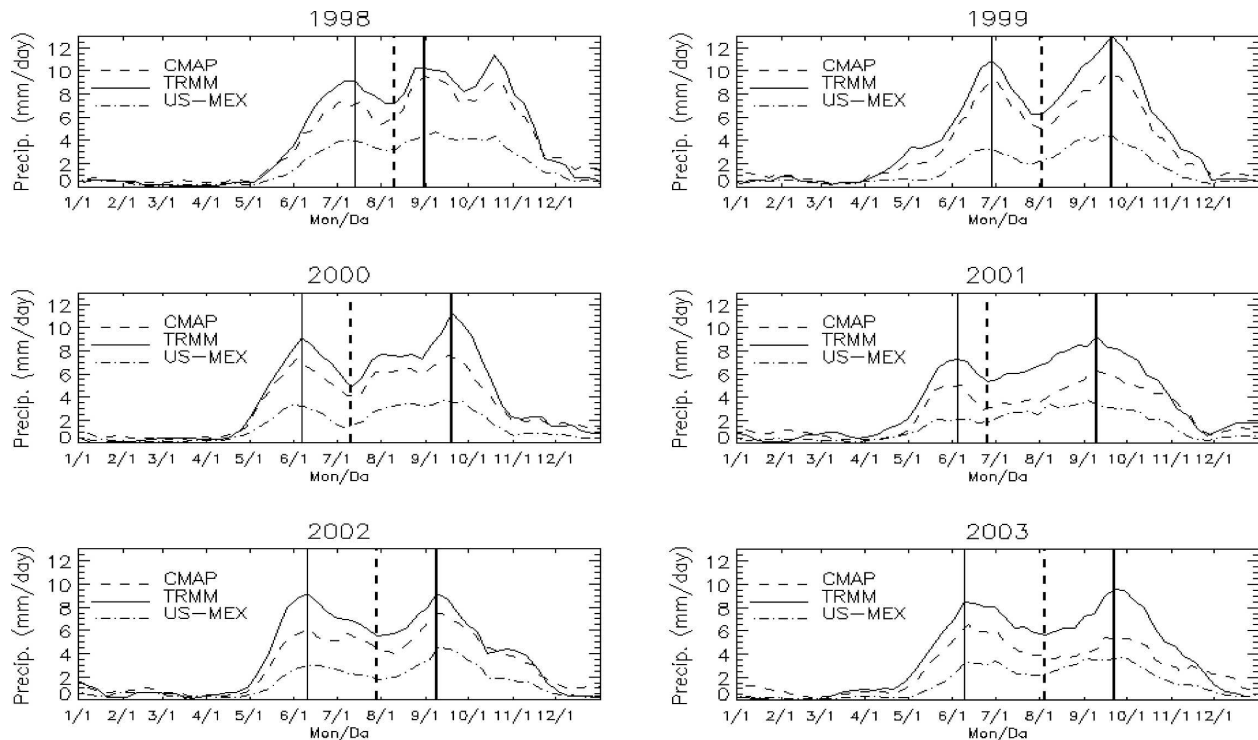


FIG. 1. Precipitation (mm day^{-1}) area averaged over the box 10° to 20°N , 100° to 85°W for the years 1998–2003 and for three datasets: CMAP, TRMM 3B42, and US-MEX gridded data. Thin, dashed, and thick vertical lines mark the early season peak, midsummer minimum, and late season peak in rainfall, respectively, based on TRMM 3B42 data. Data have been smoothed over 35 days.

a. Characteristics of the midsummer drought and North American monsoon

An estimate of the amplitude and spatial extent of the MSD may be obtained by subtracting the climatological rainfall averaged over June and September from that averaged over July and August. Similar patterns were seen in the climatology of two datasets of differing

lengths: 6 years of National Aeronautics and Space Administration (NASA) Tropical Rainfall Measuring Mission (TRMM) 3B43 data (shown in Fig. 3a) and 25 years of National Centers for Environmental Prediction (NCEP) Climate Prediction Center Merged Analysis of Precipitation (CMAP) data, giving confidence in the robustness of the spatial pattern (details of the datasets are given in section 2a). Reduced precipitation is found

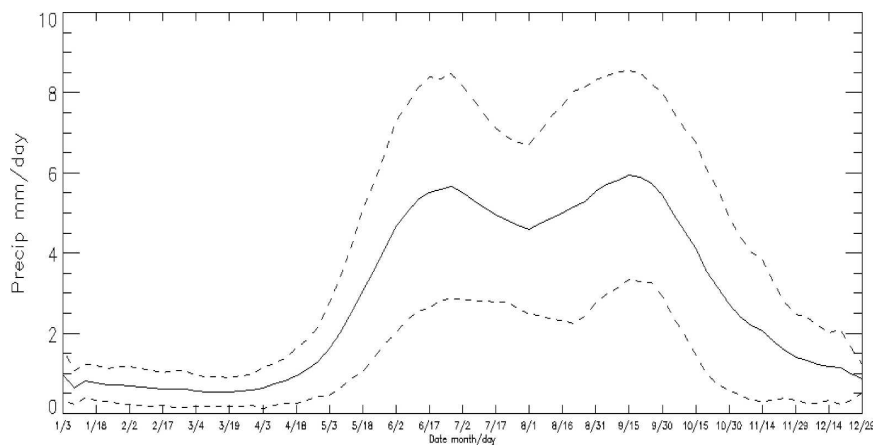


FIG. 2. Climatology of precipitation derived from CMAP pentad data (1979–2003) from 10° to 20°N , 100° to 85°W . Solid line shows the mean and the dashed lines are mean ± 1 std dev. The mean and std dev are smoothed over 35 days.

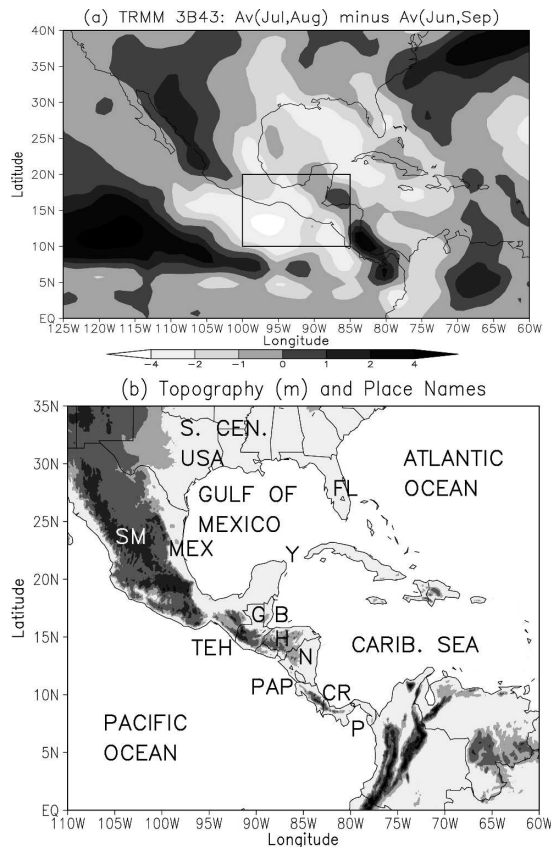


FIG. 3. (a) The average of July and August rainfall minus the average of June and September rainfall from climatological data. The domain used for composite analysis is marked by a box. Data are from TRMM 3B43. (b) Mountain topography and place names are MEX: Mexico; S. CEN. USA: south-central United States; FL: Florida; TEH: Gulf of Tehuantepec; PAP: Gulf of Papagayo; SM: Sierra Madre Occidental. The countries of Central America are G: Guatemala; B: Belize; H: Honduras; N: Nicaragua; CR: Costa Rica; P: Panama, Y: Yucatan Channel, and CARIB. SEA: Caribbean Sea.

on the west side of Central America, in the Pacific Ocean between 10° and 20°N east of 110°W, in south and east Mexico, the states of southern North America that lie east of the Sierra Madre Occidental (see Fig. 3b for locations mentioned in text), and the Gulf of Mexico and Caribbean Sea. The difference in precipitation between the dry months and the rainy months can reach up to 4 mm day^{-1} in TRMM 3B43 data (Fig. 3a) and at least 2 mm day^{-1} in CMAP (not shown).

The spatial map of Fig. 3a suggests that the MSD is coincident with enhanced rainfall over the western slopes of the Sierra Madre Occidental in southwestern United States and Mexico, part of the North American monsoon (NAM; Adams and Comrie 1997; Higgins et al. 1999; Johnson et al. 2007). The relationship between MSD and NAM will be explored further in the numerical model experiments below.

b. Proposed mechanisms driving the midsummer drought

Magana et al. (1999) proposed that feedbacks between local shortwave radiation, SST, and convection governed the evolution of the MSD. According to this view, the seasonal northward movement of the ITCZ first leads to intense convection over and west of Central America in May and June. Extensive cloudiness associated with the convection leads to reduced solar insolation over the ocean and so reduces the SST. Lower SST then leads to less convection during July and August, initiating the midsummer drought. The consequent reduction in cloudiness in July and August allows more insolation to the ocean and an increase of SST. This leads to a return to rainy season conditions toward the end of summer. Magana et al. (1999) also suggested that low-level cyclonic/anticyclonic circulation anomalies were induced by the increases/decreases in convection, which modified the strength of the easterly trade winds. Magana and Caetano (2005) found a bimodal distribution of SST at 10°N, 95°W and 12°N, 95°W but did not find observational support for the SST–convection–insolation feedback hypothesis.

Another possible factor of importance is local land–sea heating contrast between the North American continent and the adjacent seas. This was studied by Mapes et al. (2005) in an atmospheric general circulation model (AGCM) simulation with an annual and daily period artificially increased by a factor of 5, which has the main effect of increasing the temperature over land and thus the land–sea heating contrast. In their simulation, the MSD was found to be more prominent than in a normal year integration. Their study showed that the MSD was associated with an anomalous thermal low over North America and relatively high pressure over the oceans, including low-level anticyclonic circulation over the Gulf of Mexico. Brian Mapes (2006, personal communication; <http://www.rsmas.miami.edu/users/bmapes/pagestuff/pubs.html>) has further linked the observed midsummer drought to the westward extension of the subtropical Atlantic high with its enhanced levels of subsidence.

Interannual variability of the MSD may give some hints as to what are the major driving factors for MSD. Curtis (2002) and Higgins et al. (1999) present some analysis of interannual variability in the MSD and NAM regions, respectively. They find that in the summer before El Niño events the MSD was strong, particularly offshore in the eastern Pacific, and the NAM was weak in southern Mexico, suggesting a region-wide reduction in precipitation in midsummer during these years, whereas during La Niña the precipitation is in-

creased (strengthening the NAM but weakening the MSD). Curtis (2002) suggested that the influence on the MSD may be related to the changes in the ITCZ location and strength, affecting the circulation in the MSD region. Higgins et al. (1999) suggested that changes in the SST in the warm pool adjacent to the Mexican Pacific coast modified the land–sea heat temperature contrast and hence the NAM (e.g., with warmer SST during El Niño reducing the contrast in temperature and the consequent sea-breeze effect).

The previous studies listed above provide some suggestions for contributing factors to the MSD but present no conclusive mechanism. The main aim of this paper is to quantify the impact of large-scale remote factors, such as changes in the Pacific ITCZ and in the Atlantic subtropical high, on the MSD. To do this we first characterize the MSD using new satellite data combined with reanalysis products and then investigate how circulation changes due to observed diabatic heating anomalies impact the MSD, using a numerical model.

The paper is organized as follows. Section 2 describes the data and model to be used. To set the context, section 3 discusses the summer climatology and seasonal evolution of the MSD region and the surrounding area. The detailed composites from data showing the evolution of the MSD are presented in section 4. The effect of diabatic heating anomalies on the evolution of the MSD is described in section 5. Then section 6 investigates the relative importance of remote versus local diabatic forcing to the MSD. This is followed in section 7 by a discussion of how the present results relate to previous studies, and section 8 presents the conclusions.

2. Data and model

a. Observations and reanalysis

1) PRECIPITATION

Measurements of precipitation vary greatly between different types of instruments (see Adler et al. 2001; Kidd et al. 2003 for detailed discussion). For this reason it is generally recommended to use data from several sources for comparison. Land data for the United States and Mexico were taken from a 1° gridded daily dataset derived from land station data (Higgins et al. 1996), obtained from the NCEP Climate Prediction Center (CPC) Web site (see <http://www.cpc.ncep.noaa.gov/products/precip/realtime/GIS/USMEX/analysis.shtml>) and averaged into weekly periods. Two merged satellite products were also analyzed, both of which use microwave data to calibrate more frequent but less accurate infrared measurements, which may be merged

with land station data. CMAP monthly and pentad data (Xie and Arkin 1997) were used to take advantage of its long record from 1979 and is available on a 2.5° grid. CMAP data were accessed from the National Oceanic and Atmospheric Administration (NOAA) Earth System Research Laboratory (available online at <http://www.cdc.noaa.gov>). Higher-resolution data are gathered from the TRMM $\frac{1}{4}^\circ$ merged product, TRMM 3B42 (Huffman et al. 1995). For the standard analysis, the daily TRMM 3B42 (version 5) data, available from December 1998 to March 2004, were averaged over weekly periods. For the diurnal cycle analysis, the new version 6 TRMM 3B42 product, at 3-hourly resolution, is used. For Fig. 3a, the monthly TRMM 3B43 is used, which, like CMAP, also incorporates Global Precipitation Climatology Centre (GPCC) land station data. (For access to TRMM 3B42 and 3B43 see http://daac.gsfc.nasa.gov/precipitation/TRMM_README for details).

2) NEAR-SURFACE WINDS, WATER VAPOR, AND SST

Multiyear near-surface vector wind measurements are available since July 1999 from the Quick Scatterometer satellite (QuikSCAT). We use weekly averaged 10-m equivalent neutral wind data (Wentz and Smith 1999). SST and column-integrated water vapor from December 1997 is obtained from the TRMM Microwave Imager (TMI). TMI data are not affected by clouds except under heavy precipitation (Wentz et al. 2000) and hence has a significant advantage over infrared radiometers in regions of large cloud cover. All of the aforementioned data are obtained from Remote Sensing Systems (available online at www.ssmi.com), processed on a 0.25° grid in weekly averages.

3) REANALYSIS DATA

For investigation of variables that are not sensed directly by satellite, we use here the NCEP–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996), version 3 products (released in May 2002). The reanalysis provides data four times daily from 1948, and the output grid spacing is 2.5° . For this study pentad-mean data are compiled. NCEP–NCAR daily data were accessed from the NOAA Earth System Research Laboratory (www.cdc.noaa.gov).

b. Linear baroclinic model

The linear response to diabatic heating anomalies is studied using the linear baroclinic model (LBM) of Watanabe and Kimoto (2000). The primitive equations used by the Center for Climate System Research

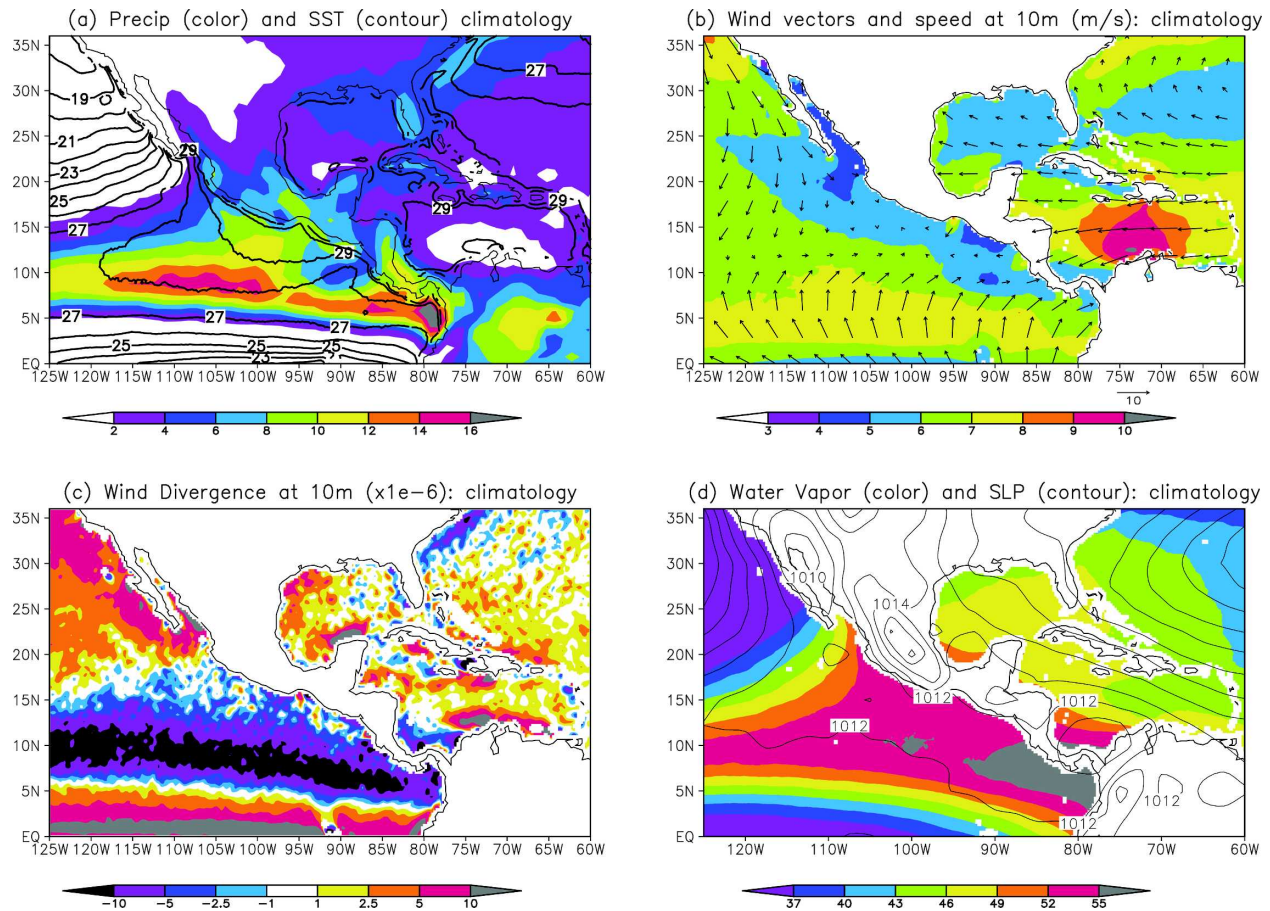


FIG. 4. The mean summer (June to September) climatology. (a) TRMM-merged rainfall product (mm day^{-1} , color) and TMI SST ($^{\circ}\text{C}$, contour). (b) QuikSCAT mean scalar wind speed (m s^{-1} , color) and mean velocity vectors (m s^{-1} , arrows, see scale arrow at bottom). (c) QuikSCAT 10-m wind divergence (10^{-6} s^{-1} , color). (d) TRMM TMI column-integrated water vapor (mm, color) and NCEP-NCAR reanalysis sea level pressure (hPa, contours). TRMM data are from 1998 to 2004, QuikSCAT from July 1999 to July 2005, and NCEP-NCAR climatology is derived from a similar period (1998–2004).

(CCSR) University of Tokyo/National Institute for Environmental Studies (NIES) AGCM are linearized about a climatological summer (June to September) mean basic state. The sigma-coordinate LBM has 20 vertical levels, contains topography, and is run at T42 resolution. The model is time integrated and has Newtonian and Rayleigh damping time scales for heat and momentum ranging from 30 days in the free troposphere to 1 day at the top and bottom levels. (The LBM was also run at T21 and at different, reasonable, values of drag to test the sensitivity of the model to horizontal resolution and to drag; none of the conclusions of the paper are significantly affected by these sensitivities.) The model responds to a specified diabatic heating anomaly, which is constant in time and has a user-defined horizontal and vertical distribution. By 20 days the solution has reached a steady state and these results are shown here. In the version of the model used here, transient eddies are not represented.

3. Summer precipitation and circulation

Before discussing the midsummer drought in detail, the context is set by describing the summer mean (defined here as June to September) fields of the east Pacific and “Intra-America Seas” (Gulf of Mexico and Caribbean Sea) and the large-scale precipitation evolution through the season. Here the mean of TRMM data is taken from 1998 to 2003, QuikSCAT from July 1999 to July 2005, and NCEP-NCAR reanalysis from a similar period (1998–2004).

a. Summer mean fields

The mean precipitation pattern derived from TRMM 3B43 is dominated by intense rainfall in the ITCZ over the ocean in the eastern Pacific centered on 10°N (Fig. 4a, color). Precipitation over land is generally less than over the ocean partly due to reduced evaporation, except in the Amazon Basin rain forest. Mean precipitation significantly decreases north of 20°N over Mexico

and the Gulf of Mexico. The Atlantic and “Intra-American Seas” region receives considerably less rainfall in summer relative to the ITCZ of the Pacific Ocean. There are two local maxima in precipitation: one in the Gulf of Panama in the Pacific, related to flow imposing on the steep topography of the Andes (Mapes et al. 2003), and a long zonal band located between 8° and 12°N, 115° and 98°W (also seen in CMAP data, not shown).

The low-level circulation in the Atlantic region in summer is dominated by the subtropical high in the western Atlantic (Fig. 4d) with associated geostrophic easterlies and southeasterlies in the Gulf of Mexico (Fig. 4b). On the Pacific side there is southerly cross-equatorial flow into the low pressure of the ITCZ (Figs. 4b,d), while farther north there is northerly geostrophic flow around the eastern edge of the Pacific subtropical high. The mean 10-m wind speeds from QuikSCAT (Fig. 4b, color) show strongest winds in the Caribbean Sea easterly trade wind region (upward of 9 m s^{-1}), weak winds over the Gulf of Mexico, and weak and variable winds off the Pacific seaboard of Central America and Mexico. Local maxima in wind speeds are seen in the Gulfs of Tehuantepec and Papagayo: these are summer manifestations of the gap wind jets (Xie et al. 2005), which are much weaker than observed in winter (Chelton et al. 2000) and are discussed more in section 4b.

The SST distribution (Fig. 4a, contours) shows the eastern Pacific warm pool underneath the ITCZ where temperatures range from 27° to 30°C, being warmer toward the Central American coast. The zonal band of highest precipitation is not located exactly over the highest SST, which occurs closer to the Mexican coast but to the south and west of it. In fact, the summer mean divergence field from QuikSCAT (Fig. 4c) indicates that there is less convergence over the warmest waters (greater than 29°C) off the Pacific Central American coast relative to that farther south and west in the main ITCZ. On the Atlantic side of Central America and in the intra-American seas the SST is also greater than 27°C and warmest in the Gulf of Mexico (>29°C), but precipitation there is minimal. These features suggest that other factors than just local SST (such as land influences and subsidence associated with large-scale atmospheric circulations) are helping to determine precipitation amount in the region and season of interest.

b. Seasonal progression of ITCZ and Atlantic subtropical high

Two of the prominent features of the large-scale circulation that can impact the northern tropical Americas

are the ITCZ and the Atlantic (Bermuda) subtropical high. The Pacific ITCZ west of 110°W moves north as the summer progresses, approximately following the SST maximum (Fig. 5a). The surface pressure low associated with the ITCZ west of 110°W follows the precipitation maximum (Fig. 5d) so that the lowest pressures west of Central America (and between 10° and 20°N) are recorded in July to September. In the Atlantic basin the seasonal evolution of the ITCZ also follows the SST maximum (Fig. 5c), but the sea level pressure distribution in the summer is dominated by the intrusion of the Atlantic high in July–August (Fig. 5f). In July–August the combination of the westward extension of the Atlantic high and the background decrease in SLP over the ITCZ west of 110°W gives rise to a peak in cross-basin pressure gradient and sets up conditions for easterly flow across the Central American landmass. These strong easterlies are important factors in the regional structure of the MSD as discussed in the next section.

The Pacific ITCZ to the east of 110°W shows different characteristics to those seen to the west of that longitude (cf. Figs. 5a,b). East of 110°W (Fig. 5b) the ITCZ does not move north between June and August, and instead a local minimum in precipitation develops over the eastern Pacific and Central Americas, centered around 11°–12°N (marked with a white ellipse in Figs. 5b,e). There appears to be no corresponding dip in SST (Fig. 5b, contours). Meanwhile the surface pressure increases slightly in July–August between latitudes 10° and 20° (Fig. 5e). These features indicate the presence of the midsummer drought, which is discussed in more detail in section 4.

c. Diurnal cycle

The precipitation over the Central American land bridge and southern Mexico in summer exhibits a pronounced diurnal cycle (Curtis 2004). Focusing on the MSD boxed area shown in Fig. 3a, the diurnal cycle from TRMM 3B42 3-hourly data is displayed as a function of month, first for land points only, then for ocean points. [All times reported below are local time (LT), assumed to be UTC – 6 h.] Land precipitation commences around 1200 and peaks at 1800 as convection is forced over the heated surface, then diminishes rapidly by 0600 (Fig. 6a). Over the ocean, the surface is most unstable during the night when the overlying air rapidly cools, and a peak between 0300 and 0600 is observed in all months, with a minimum around 1800 (Fig. 6b). Over the ocean and the land, there is a local midsummer minimum in precipitation observed during July and August (Figs. 6a,b), consistent with the results of Curtis (2004). This is seen more clearly in the annual cycle

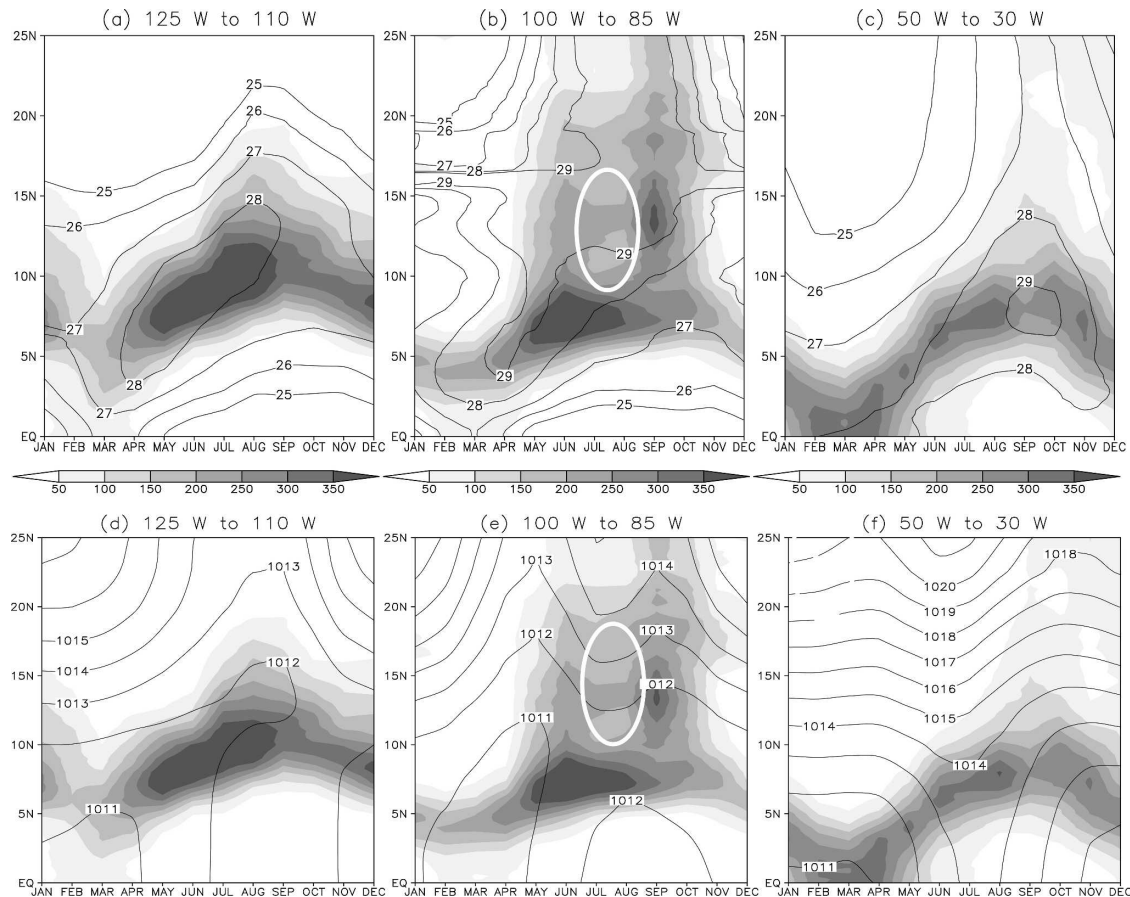


FIG. 5. Seasonal progression of pressure, SST, and rainfall rate in given longitude bands. (a)–(c) Rainfall rate from TRMM 3B43 (mm month^{-1} , shaded) vs latitude and month for three longitude bands (a) 125° – 110° W, (b) 100° – 85° W, and (c) 50° – 30° W. Overlaid are contours of SST ($^{\circ}\text{C}$) from TMI. (d)–(f) As in (a)–(c) but for pressure from NCEP–NCAR reanalysis (hPa, contours) and rain rate (mm month^{-1} , shaded). The period of the midsummer drought is marked with a white ring in (b) and (e).

computed at the time of maximum rain in the day, over land at 1800, and over the ocean at 0600 (Fig. 6c, solid and dashed lines, respectively). The magnitude of the diurnal cycle of precipitation (P) over land [represented by the difference $P(1800) - P(0600)$] likewise reduces in July and August relative to June and September (Fig. 6c, asterisks). The diurnal cycle over the ocean [represented by $P(0600) - P(1800)$] is comparable in June, July, and August but much greater in September (Fig. 6c, diamonds).

4. Composites of the midsummer drought

a. Method

To illustrate the progression of the midsummer drought, composites have been made relative to the principal maxima and minima in summer rainfall for individual years. Here the precipitation is area averaged over a box covering the peak rainfall deficit, chosen from Fig.

3a, namely, 10° – 20° N, 100° – 85° W, which includes the offshore Pacific region as well as much of southern Mexico and Central America, referred to as the MSD region.

The years chosen for initial analysis were those for which high-resolution satellite data of winds and precipitation were jointly available from QuikSCAT and TRMM at the start of this study and covered 2000 to 2003. The time series of this area-averaged precipitation is computed for each year of data and then a running average of 5 weeks is applied to the time series to remove synoptic variability but retain the midsummer drought (as shown in Fig. 1).

Despite the difference in absolute magnitude of precipitation between datasets,¹ the phasing of the rainy

¹ US–MEX, being confined to land points only, gives smaller values of rain rate. Further, the TRMM 3B42 data, derived from satellite data only, give higher values over land than CMAP, which include in situ data.

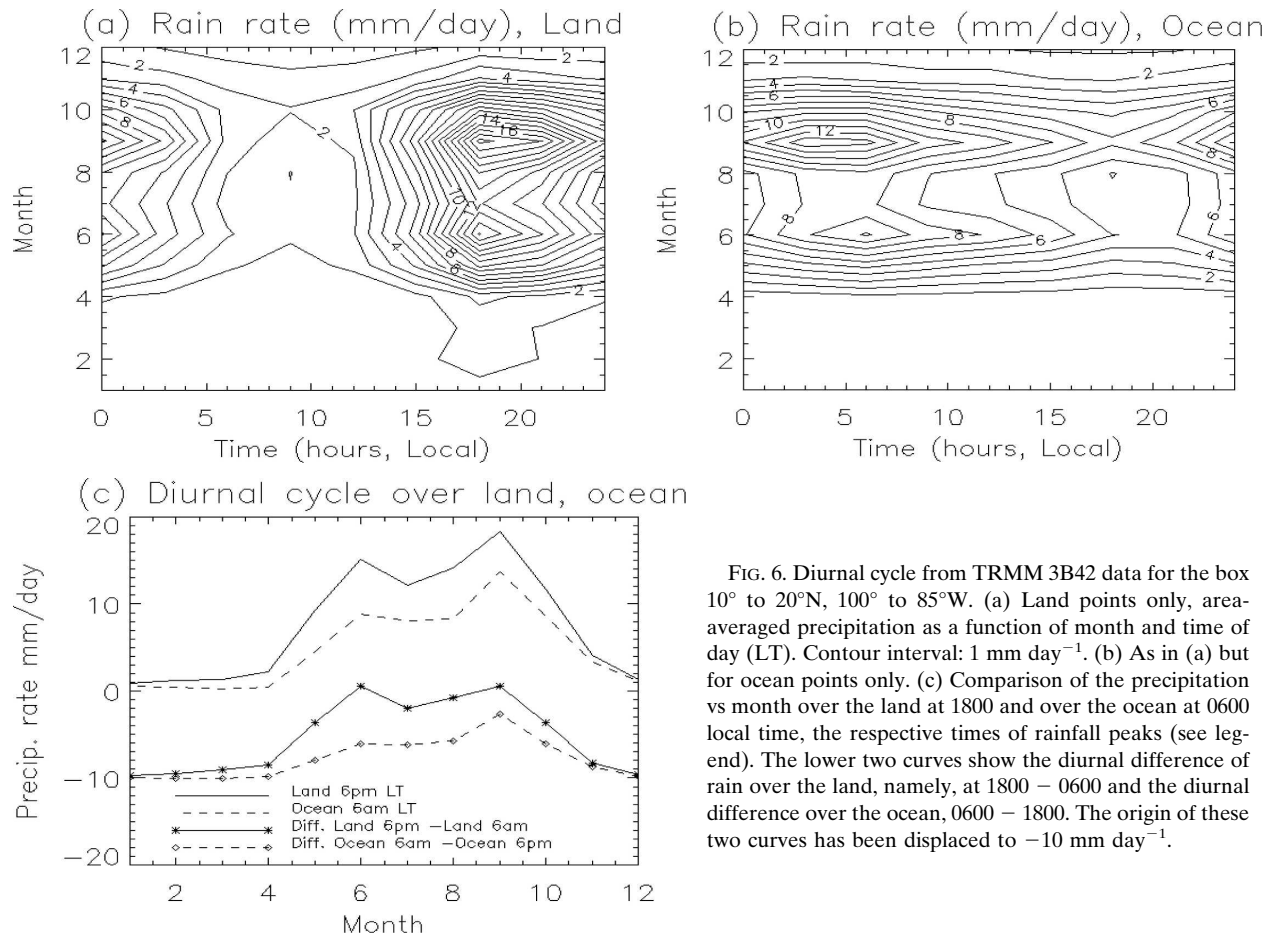


FIG. 6. Diurnal cycle from TRMM 3B42 data for the box 10° to 20°N , 100° to 85°W . (a) Land points only, area-averaged precipitation as a function of month and time of day (LT). Contour interval: 1 mm day^{-1} . (b) As in (a) but for ocean points only. (c) Comparison of the precipitation vs month over the land at 1800 and over the ocean at 0600 local time, the respective times of rainfall peaks (see legend). The lower two curves show the diurnal difference of rain over the land, namely, at $1800 - 0600$ and the diurnal difference over the ocean, $0600 - 1800$. The origin of these two curves has been displaced to -10 mm day^{-1} .

and dry periods from TRMM 3B42 agrees reasonably well with those from CMAP and the United States and Mexico (US-MEX) in Fig. 1. For this reason, and because of its good temporal and spatial resolution, the TRMM 3B42 data are used for the composites displayed below.

From Fig. 1 the early summer maximum, midsummer minimum, and late season maximum of rainfall are identified from the TRMM 3B42 data (marked with vertical lines). The timing of these events is used to define three phases of the typical summer rainfall cycle: phase 1 (early summer peak rainfall or pluvial), phase 2 (the midsummer minimum or drought), and phase 3 (late season pluvial). For each phase the corresponding spatial distributions of precipitation, winds, SST, divergence, and other variables of interest are found, also smoothed with a 5-week window and binned into 1° bins for visual clarity. Finally, composites are made by averaging each particular phase over all the years of the record and the summer mean is subtracted to give anomaly fields.

The mean and median dates of the three phases for

each year from CMAP and from TRMM 3B42 are shown in Table 1. Phase 1 is typically centered between 10 and 24 June, phase 2 between 25 July and 8 August, and phase 3 between 8 and 22 September.

b. Precipitation, circulation, and SST

The evolution of the precipitation, surface pressure, and 10-m vector wind fields is shown in Fig. 7. In phase 1 there is an increase in precipitation over most of the MSD region as the ITCZ east of 100°W extends northward toward the Central American coast (Fig. 7a). This is associated with anomalous westerlies and southwesters, which would aid moisture transport from the ITCZ. (Fig. 4d shows that the mean column-integrated water vapor is higher in the eastern Pacific than in the Caribbean Sea and Gulf of Mexico. The role of moisture advection is discussed further in section 4c.) The pressure distribution (Fig. 7a, contours) shows an anomalous low pressure covering the northern American continent and extending toward Central America, a westward extension of the Atlantic subtropical high, reaching just into the Gulf of Mexico, and lower pres-

TABLE 1. Timing of phases in the summer for the area shown as a box in Fig. 2a. The dates of the mean and median pentad number for phases 1, 2, and 3 are shown for CMAP and TRMM 3B42.

	Phase 1		Phase 2		Phase 3	
	Dates	Pentad	Dates	Pentad	Dates	Pentad
CMAP mean	20–24 Jun	35	4–8 Aug	44	8–12 Sep	51
CMAP median	20–24 Jun	35	30 Jul–3 Aug	43	13–17 Sep	52
TRMM mean	15–19 Jun	34	29–25 Jul	42	13–17 Sep	52
TRMM median	10–14 Jun	33	30 Jul–3 Aug	43	18–22 Sep	53

sure in the Pacific under the high precipitation anomalies. The westerly/southwesterly winds on the Pacific side of Central America are in an approximate geostrophic balance with the cyclonic flow around the continental low, suggesting that the enhanced precipitation in the MSD region in this phase is a result of the circulation due to continental heating, and this will be discussed further in section 5.

Phase 2 is the main midsummer drought period (Fig. 7c) during which precipitation is reduced throughout the MSD boxed region as well as in the Gulf of Mexico, Caribbean Sea, and the region of the southern United States between the Sierra Madre and the Mississippi River (referred to here as the south-central U.S. region). The surface pressure anomaly is high over most of the region illustrated, with a local maximum in the southern Gulf of Mexico (Fig. 7c, contours). There are increased trade winds in the southwest Caribbean Sea, and the winds are anomalously northerly–easterly off the Pacific coast of Central America. This includes strong winds off the Gulf of Papagayo, shown as a local maximum in wind speed there (Fig. 7d), and northerly anomalies off the Gulf of Tehuantepec. This is the summertime secondary maximum of the gap winds identified by Xie et al. (2005).

Locally enhanced precipitation is observed in phase 2 over Panama, in the Gulf of Panama, and over the Sierra Madre Occidental (Fig. 7c). The maximum east of Panama is seen more clearly in the high-resolution TRMM Precipitation Radar product, as well as a peak over the eastern part of Nicaragua (not shown), suggesting that the rainfall is on the fine spatial scale of the highlands, that is, orographic precipitation on the windward side of the steep Central American Cordillera. Some evidence for this was also found from land station data by Magana et al. (1999). The maximum in the Sierra Madre is part of the NAM discussed in section 1 and is accompanied by southerly anomalous winds in the Gulf of California, related to Gulf surges (Adams and Comrie 1997; Bordoni et al. 2004; Johnson et al. 2007).

By phase 3 precipitation is enhanced over a large region north of 10°N in the far-east Pacific and over

Central America, coincident with anomalous westerly and southwesterly winds toward Central America (Fig. 7e), and precipitation is enhanced over most of the Atlantic side. Pressure is anomalously low over both the Pacific and Atlantic, with a minimum extending from the subtropical Atlantic into the Gulf of Mexico. In contrast, precipitation decreases east of Panama, where the mean trade winds are reduced, and in the NAM region.

During phase 2 the SST is warm in the Gulf of Mexico and offshore of Baja, California, while cool anomalies exist offshore of the Gulf of Papagayo and in the southwest Caribbean Sea (Fig. 7d). The region off Papagayo includes part of the Costa Rica Dome, a permanent cyclonic eddy (Kessler 2002). Here the thermocline is shallow (<40-m depth) all year round, partly as a result of Rossby wave propagation and persistence of sea surface height (SSH) anomalies generated by strong winter gap wind events (Xie et al. 2005). Hence the SST in this region is quite sensitive to any entrainment or upwelling induced by gap wind events in summer so that a local cool patch arises.² In phases 1–3 there is a generally negative correlation between wind speed and SST (Figs. 7b,d,f), suggesting wind speed is forcing an SST response by modulating the mixing and entrainment and evaporation, and in particular the wind speeds are larger than normal in the southwest Caribbean Sea and off Papagayo (Fig. 7d).

The lack of a long data record from TRMM and QuikSCAT raises the question of whether the composites are showing robust climatological features. To test this, the results were compared with those from a longer period (1979 to 2003) of CMAP precipitation and NCEP–NCAR reanalysis wind data. The results from these longer records (not shown) are similar to those from TRMM and QuikSCAT, although CMAP recorded a much weaker rainfall signature over the NAM.

² The measurements by Magana and Caetano (2005) of a bimodal distribution of SST were found at Tropical Atmosphere Ocean moorings within or close to the influence of the cool SST patch offshore of Papagayo.

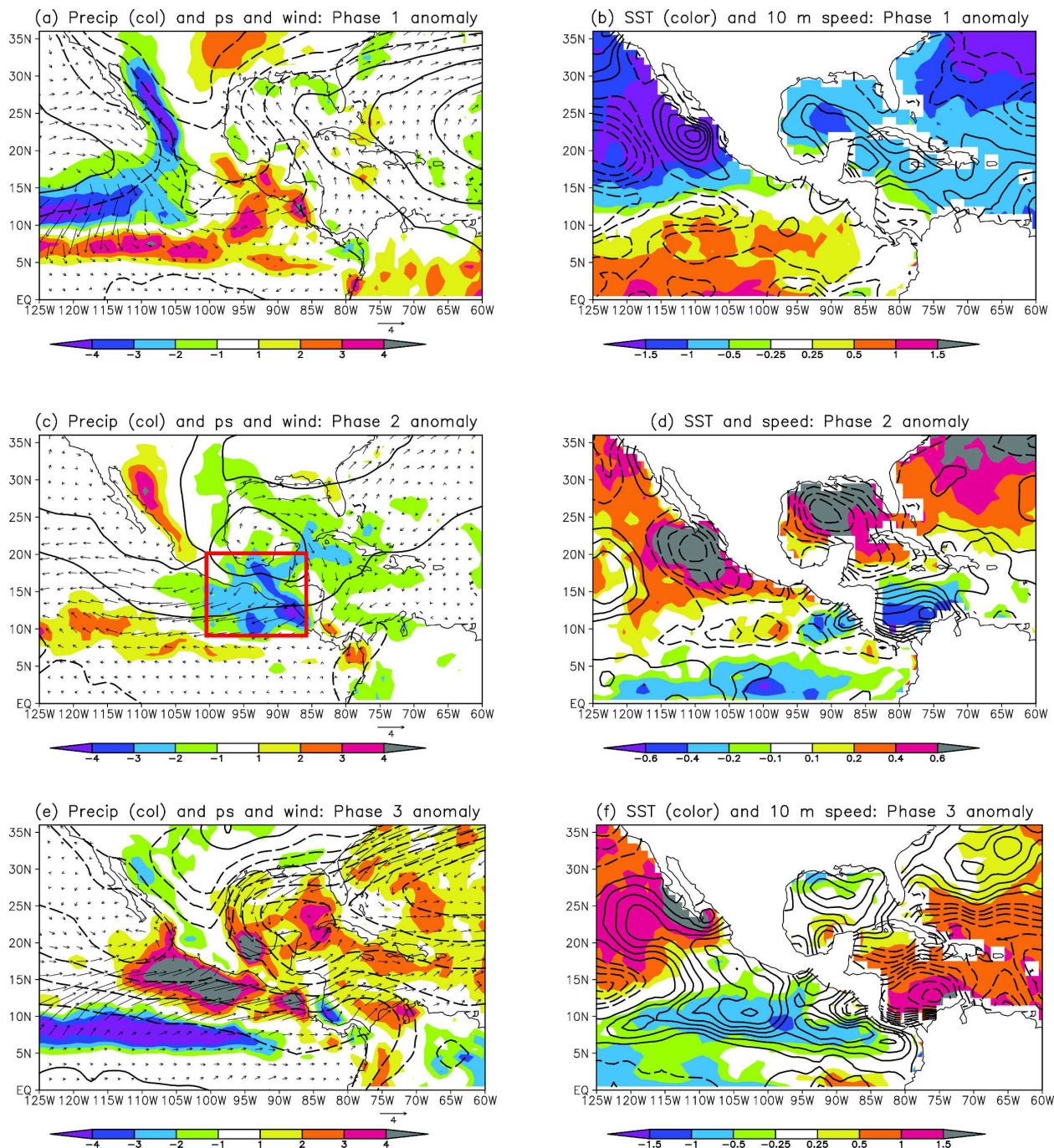


FIG. 7. (a), (c), (e) Precipitation (mm day^{-1} , color), surface pressure (hPa, 0.4-hPa intervals, positive values solid and zero and negative values dashed), and equivalent neutral 10-m winds (m s^{-1} , arrows, see scale arrow at bottom right) composited into three phases: (a) phase 1 anomaly, (c) phase 2 anomaly, and (e) phase 3 anomaly. (b), (d), (f) Corresponding SST (K, color) and 10-m wind speed (m s^{-1} , contoured at 0.2 m s^{-1} intervals, negative dashed and zero omitted) composited into the same three phases. Precipitation from TRMM 3B42, winds from QuikSCAT SST from TMI (K, color), and surface pressure from NCEP-NCAR. The area on which the composites are based is shown as a box in (c).

c. Moisture flux

In this section, the relative contributions of the evaporation and the moisture flux to the precipitation variations are quantified. As evaporation and vertically integrated moisture flux are not directly observable from space, reanalysis data from NCEP–NCAR are used to estimate these quantities.

Following Holton (1992), we assume that in the steady state,

$$P = - \int_{z=0}^{z=z_m} \nabla \cdot (\rho q \mathbf{u}) dz + E, \quad (1)$$

where P is precipitation rate, E is evaporation rate, $\mathbf{u} = (u, v)$ is the horizontal velocity, q is the specific humidity, ∇ is the horizontal gradient operator, and the limits of the integral are the ocean surface and the top of the moist layer. The moisture flux $\mathbf{Q}_u = (Q_u, Q_v)$ is computed in pressure coordinates as

$$Q_u = \frac{1}{g} \int_{300}^{ps} q u dp, \quad Q_v = \frac{1}{g} \int_{300}^{ps} q v dp, \quad (2)$$

where ps is the surface pressure and the integral is computed to the highest available NCEP–NCAR humidity data at 300 hPa. Humidity, precipitation, evaporation,

and velocity fields are extracted each day for analysis. The moisture flux vectors, precipitation, and evaporation from the reanalysis were compiled into pentads, then composited relative to the three phases of the summer (timings of which were obtained from CMAP), as in section 4a, using data from 1979 to 2003.

The moisture flux can be decomposed by separating out the time mean and anomaly parts of the water vapor and winds:

$$\begin{aligned} q &= \bar{q} + q' \\ \mathbf{u} &= \bar{\mathbf{u}} + \mathbf{u}', \end{aligned} \quad (3)$$

where an overbar denotes the climatological summer mean. Then we may write

$$\bar{\mathbf{Q}}_u = \frac{1}{g} \int_{300}^{ps} (\bar{q} \bar{\mathbf{u}} + \overline{q' \mathbf{u}'}) dp, \quad (a)$$

$$\mathbf{Q}'_u = \frac{1}{g} \int_{300}^{ps} (\overline{q' \mathbf{u}'}) + (q' \mathbf{u}' - \overline{q' \mathbf{u}'}) dp, \quad (b) \quad (4)$$

i) ii) iii)

where $\mathbf{Q}'_u$ denotes the anomalies at different phases of the MSD, and the convergence of the anomalous moisture flux $-\nabla \cdot \mathbf{Q}'_u$ is given by

$$-\nabla \cdot \mathbf{Q}'_u = - \frac{1}{g} \int_{300}^{ps} (\bar{q} \nabla \cdot \mathbf{u}' + \mathbf{u}' \cdot \nabla \bar{q}) dp - \frac{1}{g} \int_{300}^{ps} [\nabla \cdot (\text{term ii}) + \nabla \cdot (\text{term iii})] dp. \quad (5)$$

I) II)

Here term I of (5) denotes the contribution of the anomalous wind convergence times the mean moisture, and term II is the anomalous wind acting on the mean moisture gradients.

The convergence of the anomalous moisture flux is shown in Fig. 8b, and this is the greatest contributor to the precipitation rate (Fig. 8a) with a much weaker evaporation rate (an order of magnitude smaller, not shown). The largest contributor to the anomalous moisture flux convergence seen over the eastern Pacific and intra-Americas seas is term I (shown in Fig. 8c, cf. Fig. 8b). A separate computation indicated that term I is mainly due to the integral from the surface to 850 hPa. Further, the spatial distribution of term I over the ocean agrees qualitatively with the convergence of the anomalous 10-m winds from QuikSCAT (Fig. 8d).

The above results show that the MSD precipitation deficit is due mainly to the column-integrated moisture flux divergence and that this is determined primarily by

the divergence of the anomalous low-level winds. The anomalous winds may be considered to be partly a local response to the precipitation anomaly and partly a response to remote forcing. In the next two sections the relative contribution of these two forcings is estimated using a global, linear model.

5. Circulation response to changes in large-scale precipitation

In this section the observed precipitation anomalies from the three phases described in the previous section will be related to the circulation. The approach taken here is to insert the diabatic heating anomalies from each phase into the LBM linearized about the summer mean basic state, derived from NCEP–NCAR reanalysis data. In the following subsections the LBM response to condensational and sensible heating anomalies are

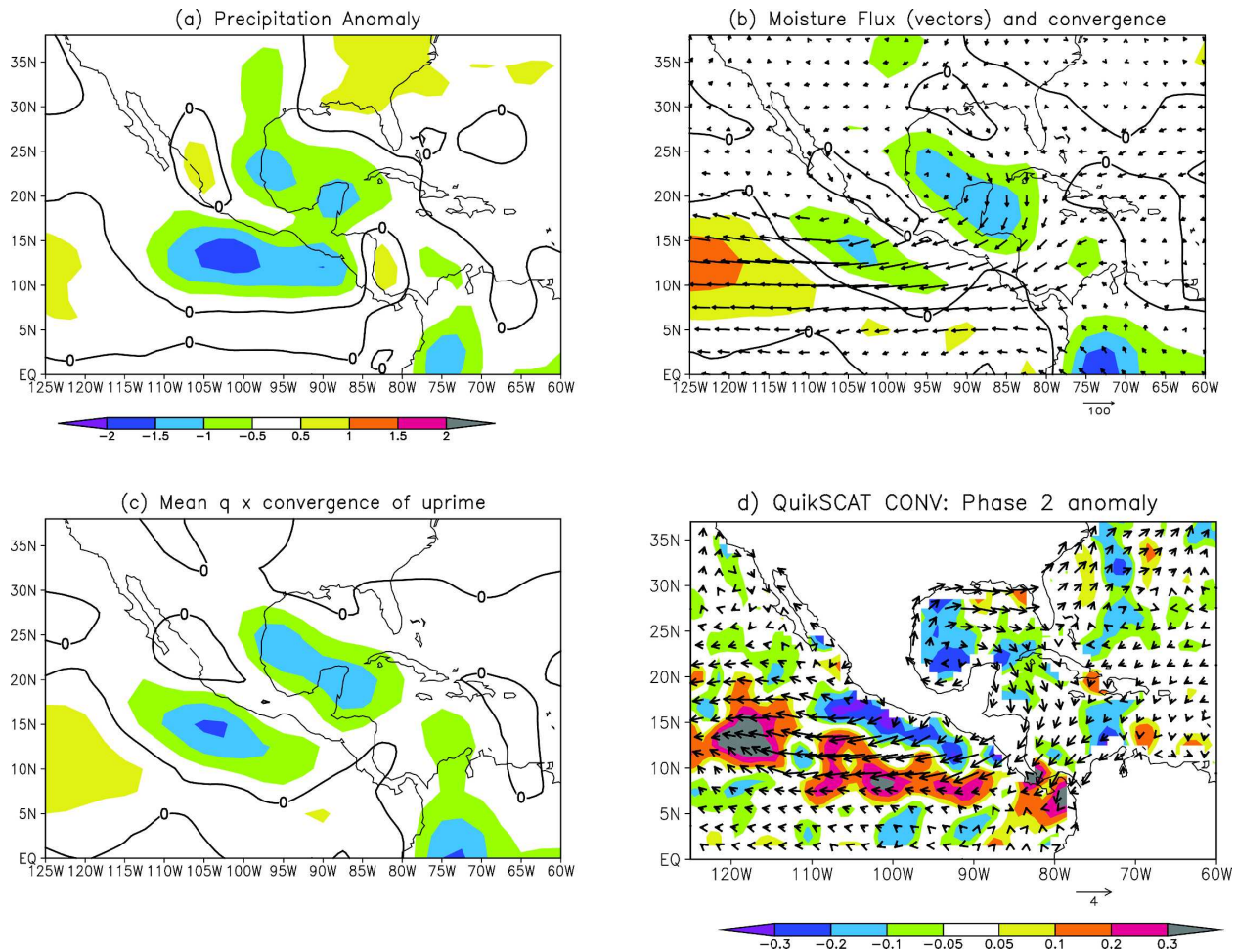


FIG. 8. Anomalies in the moisture budget during phase 2: (a) precipitation (mm day^{-1}), (b) convergence of moisture flux (mm day^{-1} , color) and moisture flux vectors (kg m s^{-1}), (c) mean moisture multiplied by convergence of anomalous winds, and (d) convergence anomaly of 10-m winds (10^{-5} s^{-1}) and wind velocity anomalies (m s^{-1}) (see text for details). The legend in (a) applies to (a)–(c) in which the zero contour is overlaid. Panels (a)–(c) are derived from NCEP–NCAR reanalysis data and (d) from QuikSCAT satellite data.

first analyzed separately and then in combination. All plots show low-level (surface or 850 hPa) quantities that are the most relevant for this study.

a. Effect of condensational heating due to precipitation

Precipitation anomalies from the Tropics and subtropics have been composited from TRMM 3B42 data following the method of section 4 for all longitudes and latitudes within 40° of the equator. An idealized vertical heating structure of a Gaussian function centered on $\sigma = 0.45$ with an e -folding vertical decay scale of $L_\sigma = 0.22$ is used (Fig. 9, solid line). Although this may not be appropriate for all environments [e.g., Thompson et al. (1979) found a more complex structure over the east Atlantic], it is a useful approximation to many observed divergence and heating profiles (see, e.g., Lin et al.

2004), including that observed over the Sierra Madre in the North American Monsoon Experiment (NAME; Johnson et al. 2007). The data are then binned onto the T42 grid of the model. The resulting maps of diabatic heating anomaly are shown in Figs. 10a,c,e (only longitudes east of the date line are shown here) and the corresponding low-level response from the LBM is given in Figs. 11a,c,e.

In phase 1, early in the summer, there is a dipole of precipitation anomalies over the Atlantic and Pacific ITCZ, which indicates that the ITCZ is displaced south of its summer mean (Fig. 10a). As discussed in regard to Fig. 5, this is mostly due to the relatively lower latitude SST maximum in early summer. As a result, the LBM produces a higher surface pressure than normal in both the eastern Pacific and the western Atlantic (Fig. 11a).

When the precipitation anomalies from phase 2 (Fig.

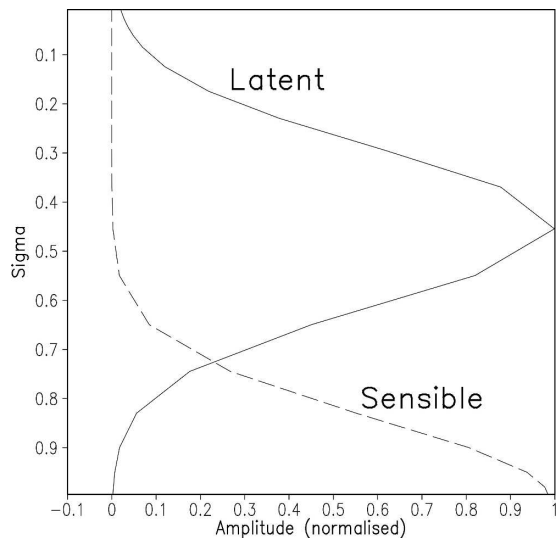


FIG. 9. Vertical structures of heating rate used in LBM. Solid line shows latent heating and dashed line shows the sensible heating. [Note that for the latent heating vertical distribution, a precipitation anomaly of 1 mm day^{-1} (30 W m^{-2}) corresponds to a tendency of 0.62 K day^{-1} at $\sigma = 0.45$ (or a vertical average of 0.24 K day^{-1}), while 1 W m^{-2} of sensible heat is equivalent to a surface temperature heating rate of 0.06 K day^{-1} .]

10c) are used to force the model, the high pressures in the Pacific and Atlantic seen in response to phase 1 merge into one large system straddling the landmass and having a maximum over the Gulf of Mexico of amplitude 0.7 hPa (Fig. 11c). The anomalous winds at 850 hPa are easterly across Central America with magnitudes between 0.5 and 1 m s^{-1} .

The precipitation distribution in phase 3 is marked by a broad stretch of positive anomalies over the Atlantic (from 10° to 40°N), the intra-America seas, and off Mexico in the Pacific (Fig. 10e). The LBM response to phase 3 (Fig. 11e) shows negative pressure anomalies everywhere in the eastern Pacific and western Atlantic domain. The Atlantic low is a baroclinic response to the increased Atlantic precipitation and the Pacific low is linked to a dipole of precipitation anomalies in the eastern Pacific, indicating a northward-displaced ITCZ off Mexico.

b. Effect of local sensible heating

In addition to the forcing of circulation by latent heat release, there may also be an effect due to sensible heating at the land surface. In the absence of direct measurements of sensible heating from satellite, we use NCEP–NCAR reanalysis sensible heat patterns for the western hemisphere from 60°S to 60°N , 180° to 30°W (thus focusing on the effects of the American landmass and adjacent seas). The sensible heat is converted to

heating rate by assuming a Gaussian vertical structure that has a maximum at the surface and has an e -folding scale of $\sigma = 0.16$, thus approximating boundary layer heating (Fig. 9, dashed line).

Phase 1 typically occurs within a week or two of the time of the summer solstice (22 June) when insolation is maximum in the Northern Hemisphere. At this time the intensity of the heat source of the high topography of the Sierra Madre Occidental reaches a maximum, as seen in Fig. 10b, and this results in a thermal low anomaly (relative to the summer mean) in the LBM response (Fig. 11b). This low pressure is accompanied by northwesterly winds along the California coast and westerly flow from the Pacific into southern Mexico. This would lead to an advection of relatively dry north-east subtropical Pacific air (see Fig. 4d) toward northern Mexico and moist Pacific ITCZ air toward southern Mexico. This may partly explain why the precipitation anomalies in phase 1 (Fig. 10a) are low over northern Mexico and the Sierra Madre but are high over the Gulf of Tehuantepec in southern Mexico. The thermal low over northern America is accompanied by a weak surface anticyclone over the subtropical Atlantic (Fig. 11b) and subsidence (not shown). This contributes to the reduction in precipitation over the subtropical Atlantic (Fig. 10a) during that phase. Meanwhile cool land surfaces over South America (Fig. 10b) give rise to a continental thermal high whose influence also extends into the northern Atlantic (Fig. 11b).

The spatial distribution of the forcing at the surface for phase 2 (Fig. 10d) shows weaker anomalies over the Sierra Madre than in phase 1, and the resulting LBM circulation anomalies (Fig. 11d) show weaker thermal low effects over the North American landmass and a correspondingly smaller area of higher pressure in the subtropical Atlantic than in phase 1. By phase 3 the relatively cool land surface in late summer provides negative heating anomalies over north and western America (Fig. 10f), which induce a thermal high in the LBM response (Fig. 11f). The ridge of the high pressure extends down through Mexico. Associated with the continental high is lower pressure over the Atlantic Ocean, helping to sustain conditions for the enhanced rainfall seen in Fig. 10e.

c. Comparison of LBM results with NCEP–NCAR reanalysis

The sum of the LBM responses to latent heating and surface sensible heating are compared to the NCEP–NCAR reanalysis composites in Fig. 12. The LBM simulations and the reanalysis show a similar general trend between the three phases of the summer; a westward extension of the Atlantic high and eastward ex-

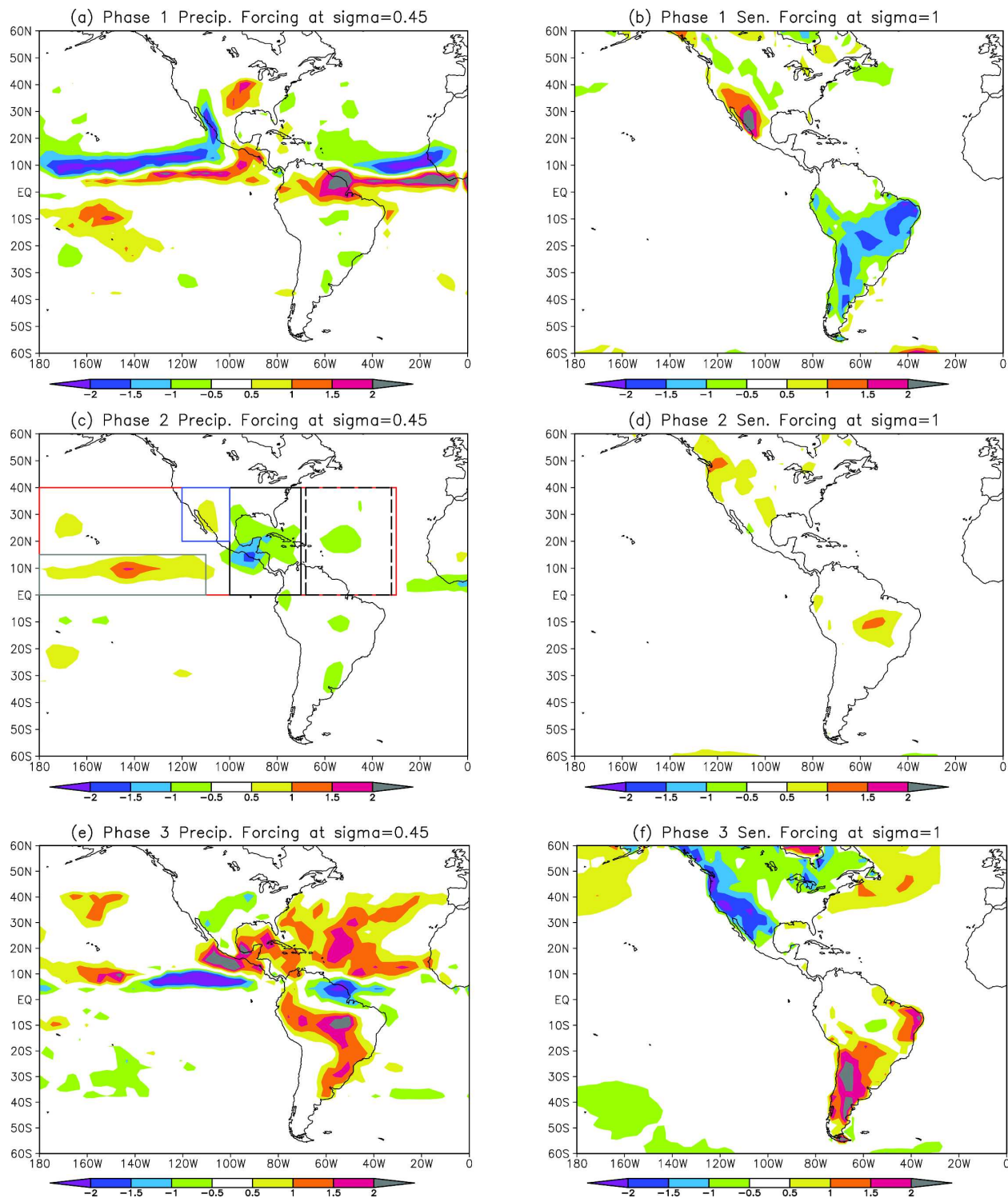


FIG. 10. Equivalent heating rate anomalies (K day^{-1}) derived from (left) TRMM 3B42 precipitation and (right) NCEP sensible heat for (a), (b) phase 1, (c), (d) phase 2, and (e), (f) phase 3. The heating rate is shown at its maximum level in the vertical, i.e., at $\sigma = 0.45$ for condensation heating and $\sigma = 1$ for sensible heating. Regional areas for analysis are shown in (c) (see text). Note that the precipitation data only cover from 40°S to 40°N and that sensible heating was not computed east of 30°W .

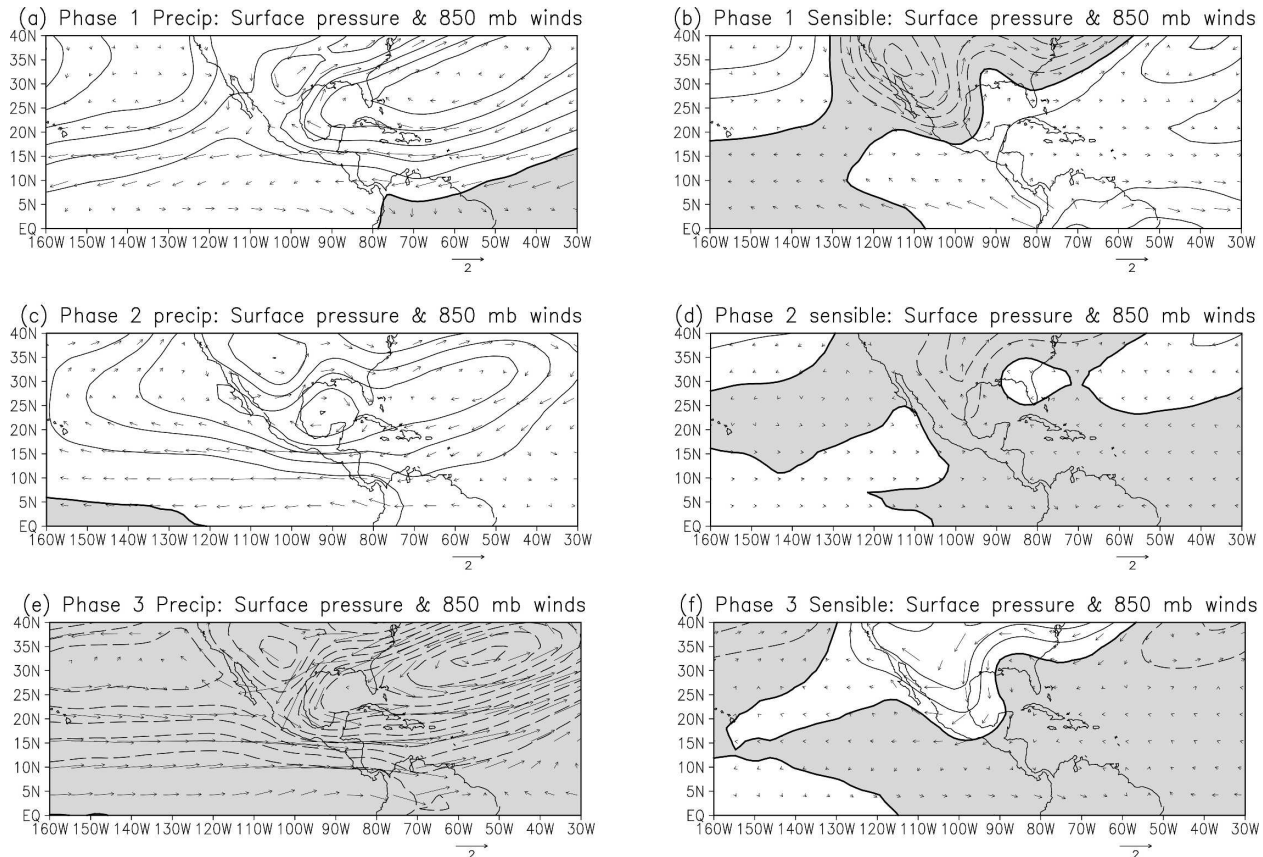


FIG. 11. LBM response to diabatic heating anomalies due to (left) precipitation and (right) sensible heating: (a), (b) phase 1, (c), (d) phase 2, and (e), (f) phase 3. The contours show surface pressure with intervals of 0.2 hPa in (a)–(d) and 0.4 hPa in (e), (f); negative anomalies are dashed contours on dark shading. The vectors are 850-hPa wind; the 2 m s^{-1} scale arrow is shown below the plot.

tension of high pressure in the subtropical Pacific, separated by a continental low in phase 1 (Figs. 12a,b); the merging of the highs in phase 2 with a maximum over the Gulf of Mexico and easterly flow across Central America (Figs. 12c,d); and large cyclonic low pressure circulations in the Atlantic and Pacific and high pressure over northern America in phase 3, with westerly flow across Central America (Figs. 12e,f).

On closer inspection there are several quantitative differences between LBM and reanalysis, as expected from comparing a simple process model with a full physics model with data assimilation. For example, the simulated extension of the Atlantic high in phase 1 reaches all the way into the Gulf of Mexico and across Central America (Fig. 12a), whereas the reanalysis high only extends as far west as 90°W (Fig. 12b). This results in the too early appearance of easterly wind anomalies across Central America in the model. The reason for this may be that the thermal low over the continent is too weak in the model (Figs. 11b and 12a) and insufficient to counter the high pressure response to the Atlantic precipitation deficit, or that the pressure re-

sponse to the Atlantic deficit in phase 1 (Figs. 11a and 12a) is too strong.

6. Relative effects of regional precipitation anomalies on the midsummer drought

In this section the remote influence of precipitation anomalies on the MSD during phase 2 will be classified by region using the LBM model and then compared to effects of local forcing. The model response is illustrated by showing the surface pressure, 850-hPa winds, and the low-level divergence fields vertically averaged between 1000 and 700 hPa.

a. Impact of individual regions

Experiments were performed to estimate the response separately to the forcing from four main heating anomalies seen in Fig. 10c, which include the local precipitation deficit in the MSD region (black box), as well as remote forcings such as the east Pacific ITCZ (gray box), the Sierra Madre (blue box), and the subtropical

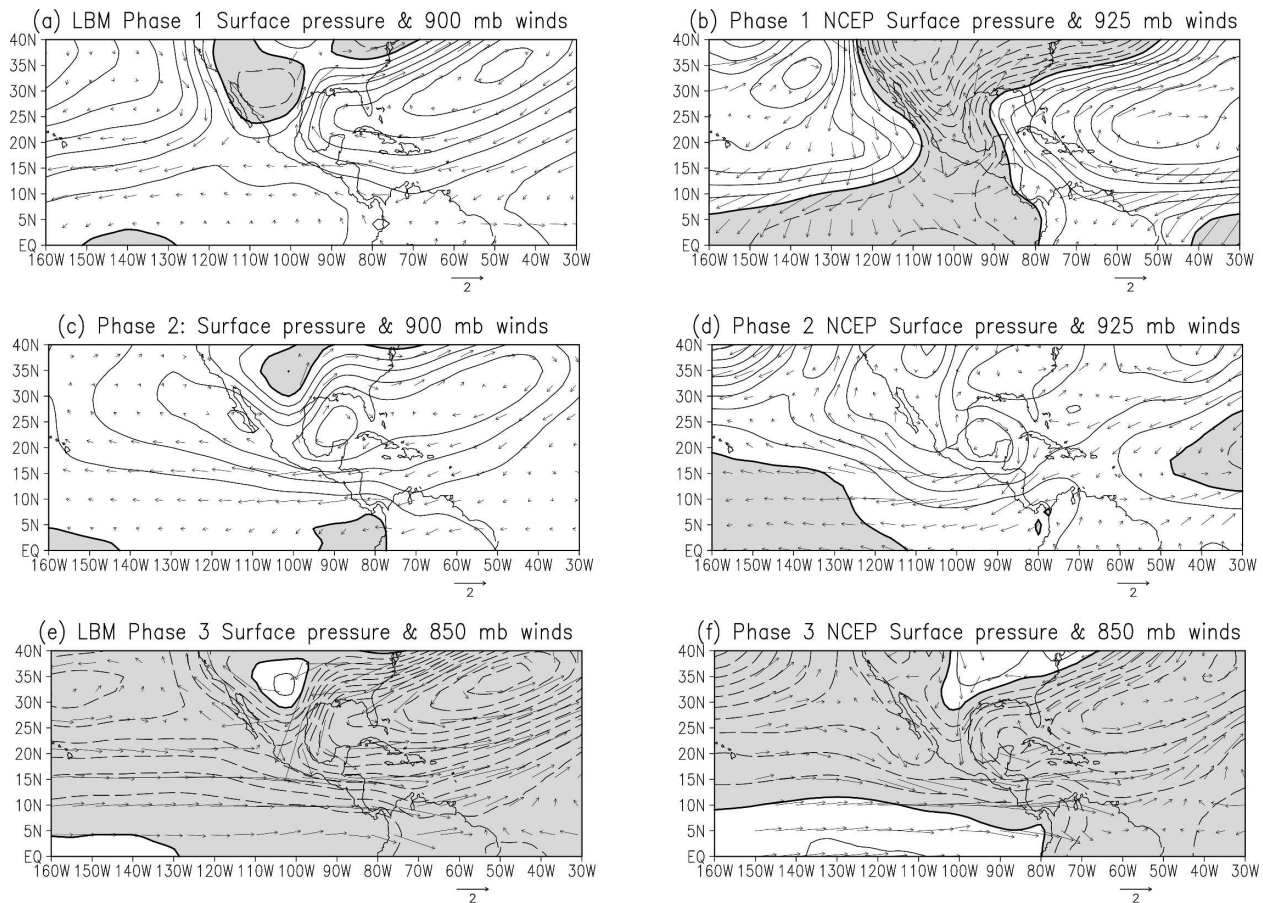


FIG. 12. Lower-troposphere circulation patterns in the three phases of summer. Response to a combination of diabatic heating anomalies from precipitation and sensible heating from (left) the LBM compared to (right) composites of NCEP–NCAR reanalysis: (a), (b) phase 1, (c), (d) phase 2, and (e), (f) phase 3. The contours show surface pressure with intervals of 0.2 hPa in (a)–(d) and 0.4 hPa in (e), (f); negative values are dashed contours on dark shading. The vectors are 850-hPa wind; the scale arrow of 2 m s^{-1} is shown below the plot.

Atlantic (black dashed box). In preliminary experiments (not shown), it was found that the significant part of the response in phase 2 was due to forcing east of the date line and west of 20°W , marked as the red box in Fig. 10c.

First, when the model is forced with the deficit of heating in the MSD region, one would expect descent over the heating deficit region and low-level divergence, together with a Rossby wave response including a low-level anticyclone to the northwest of the center of the anomaly (following Gill 1980), as seen in Figs. 13a and 14a. The strong southward flow in the Gulf of Mexico (which is part of the anticyclone) is due to a Sverdrup balance relating low-level equatorward flow to a region of descent (Rodwell and Hoskins 2001). The local low-level anticyclonic flow that is observed over the Gulf of Mexico (Fig. 7c) is consistent with the modeled response (Fig. 13a).

The Pacific ITCZ anomaly (Fig. 10c) is somewhat

closer to the equator and sets up a significant equatorial Kelvin wave response that influences the MSD region, with easterly flow across the Central American isthmus at 850 hPa toward low pressure located over the heating anomaly (Fig. 13c). To the north and east of the ascent region, including over the North American landmass and subtropical Atlantic, there is high pressure, with a weak maximum over the Gulf of Mexico (Fig. 13c) associated with descent and low-level divergence over much of the MSD region (Fig. 14c), partly due to the forced anomalous Walker cell circulation (Gill 1980).

The Sierra Madre forcing results in localized ascent and, to the east, descent and divergence over the south-central U.S. region and Gulf of Mexico (Fig. 14b). It induces a broad surface low (Fig. 13b), westerly 850-hPa winds across Mexico, and southwesterly winds in the Gulf of Mexico and off the Atlantic seaboard, but little wind response over Central America. There is poleward flow over the southern United States, possi-

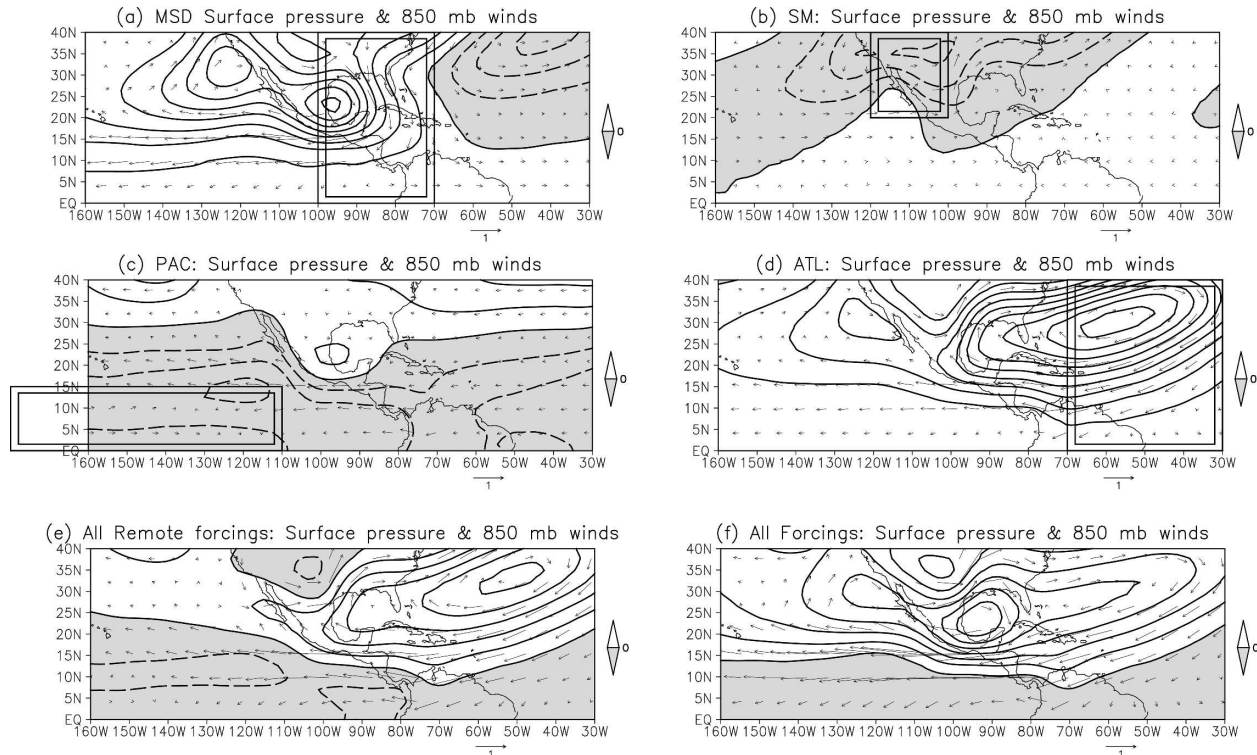


FIG. 13. Circulation patterns from the linear baroclinic model forced by TRMM 3B42 diabatic heating anomalies. Surface pressure [hPa; (a)–(d) 0.1-hPa intervals and (e), (f) 0.2-hPa intervals; low pressure shown as dashed contours and shading] and 850-hPa winds (m s^{-1} , see 1 m s^{-1} scale arrow at bottom right). (a) MSD region, (b) Sierra Madre (SM) region, (c) Pacific ITCZ region (PAC), (d) Atlantic subtropical high (ATL), (e) combination of remote forcings (SM+PAC+ATL), and (f) all forcings (MSD+SM+PAC+ATL). The location of the forcing regions is marked by double-lined boxes in (a)–(d) [the Pacific region in (c) extends outside the plot to the date line].

bly related to the influence of NAM forcing on the Great Plains jet (as discussed by Rodwell and Hoskins 2001).

The deficit of rainfall in the subtropical Atlantic sets up a low-level anticyclone of about 0.4 hPa with associated easterlies and southeasterlies across Central America (Fig. 13d). Note that the amplitude of Atlantic precipitation deficit is much smaller than over the MSD region (Fig. 10c), but the pressure response is comparable, suggesting the Atlantic heating acts better as a Rossby wave source in the local background flow. The high pressure cell extends much farther west than the extent of the forcing region, indicative of westward Rossby wave propagation. Low-level divergence is seen under the heating deficit and to the west, over Central America (Fig. 14d). By comparing Figs. 13a and 13d it can be seen that the westward development of the Atlantic high in phase 2 is due both to local MSD and to remote forcing from the Atlantic and to a lesser extent from the Pacific. Further comparison of the 850-hPa wind responses in Figs. 13c,d shows that the cross-Central America winds are roughly due in equal part to Atlantic and Pacific forcing.

b. Impact of combined remote forcings versus local forcing

When all three remote forcings are summed, the response is a surface high pressure in the Atlantic extending into the Gulf of Mexico (mainly due to the Atlantic deficit), low pressure over the American continent (due to the Sierra Madre precipitation), and easterly winds across Central America at 850 hPa (due to the combination of the push from the Atlantic high and the pull from the Pacific ITCZ low) (Fig. 13e). Convergence and ascending motion is seen both over the Pacific ITCZ and the Sierra Madre and divergence and descending motion over the subtropical Atlantic and over south-central U.S. region and Central America (Fig. 14e). Comparing this with the circulation anomalies from remote and local (MSD region) forcing combined (Figs. 13f and 14f), it can be seen that the easterly winds in the Caribbean Sea, the low-level divergence over Central America, and the subsidence over south-central United States have a significant component due to the remote forcing. (Note that here the relatively small contribution from sensible heating in phase 2 has not

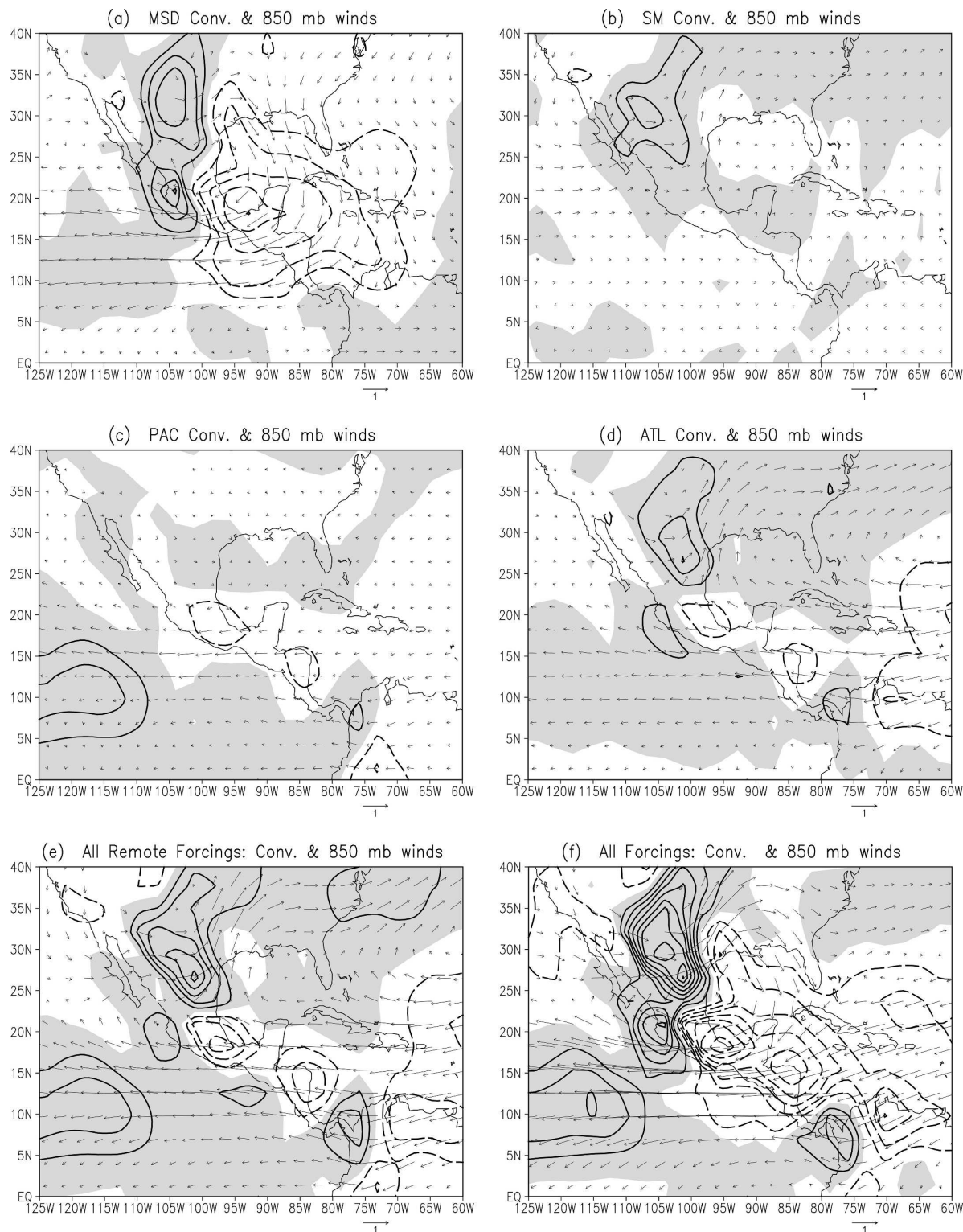


FIG. 14. Results from the linear baroclinic model forced by TRMM 3B42 diabatic heating anomalies. Wind convergence vertically averaged between 1000 and 700 hPa (s^{-1} , $1 \times 10^{-7} s^{-1}$ intervals; convergence is shaded with solid contours, divergence has dashed contours) and 850-hPa winds ($m s^{-1}$, see $1 m s^{-1}$ scale arrow at bottom right). (a) MSD region, (b) Sierra Madre region, (c) Pacific region, (d) Atlantic region, (e) sum of response to all remote forcings (b)–(d), and (f) sum of response to all forcings (a)–(d).

been included, giving rise to the small difference in surface pressure and winds between Fig. 13f and that shown in Fig. 12c.)

These results suggest that the remote forcing sets up conditions suitable to reduce precipitation in the MSD region by increasing the low-level divergence over the MSD region and inducing easterly anomalies that would flux less moist air from the Caribbean Sea toward Central America (see section 4c). Then the local forcing of the circulation (by the MSD diabatic heating anomalies) reinforces these conditions.

7. Discussion

This paper has investigated the importance of seasonal changes in the Pacific ITCZ and of the Atlantic subtropical high to the development and decay of the Central American MSD. These processes are large-scale, and remote, but not necessarily exclusive of previously suggested, more local, mechanisms.

Under the hypothesis of Magana et al. (1999) (see section 1b), locally cooler SST initiates the MSD by supporting less convection and causing a low-level anticyclonic anomaly and divergence. However, it is notable that, according to the composites of section 4b, only in a relatively small area of the domain, namely, offshore of the Gulf of Papagayo and in the southwest Caribbean Sea, does cool SST coincide with reduced precipitation during the MSD (Figs. 7c,d). Of these two areas, the precipitation signal in the southwest Caribbean Sea is fairly weak and changes sign near the Panama coast (Fig. 7c). If the reduced SST does lead to reduced convection,³ then one would expect a much larger signal in the Caribbean Sea relative to that found in the Gulf of Mexico and the Yucatan Channel, where SST is anomalously warm. This is not the case (Figs. 7c,d).

It is possible that the compositing method may smooth out some subtle temporal relationships between rain, SST, and wind. An inspection of the 35-day smoothed time series of precipitation and SST in the MSD box for 8 years (1998–2005) of TRMM data revealed that in most years SST anomalies did lead rain anomalies of the same sign by 1–3 weeks. In addition, reasonably high correlations were found between the rain rate and the zonal wind, at zero lag, such that westerly (easterly) winds coincided with high (low) rain rate, as found in the composites. These results suggest

that both the local SST variability and the easterly wind anomalies (which arise through the remote and local mechanisms proposed in sections 5 and 6) may have an impact on the MSD. The precise explanation for the local SST temporal evolution, and whether it is essential for the MSD, needs further investigation.

The influence of land–sea temperature contrast on MSD was pointed out by Mapes et al. (2005). In the present study, the relatively high temperatures over land in early and mid summer lead to the formation of a thermal low in phase 1 in both observations and model close to the time of the summer solstice, which is associated with descent over the adjacent subtropical oceans and may precondition the westward movement of the subtropical high. Interestingly, we find that the MSD appears when the thermal low starts to withdraw, possibly allowing the oceanic high pressure influence to intrude farther west into the MSD region. Once a precipitation deficit is set up in the subtropical North Atlantic, it forces a low-level anticyclonic circulation anomaly that extends over and to the west of the source, thus providing positive feedback and possible westward propagation.

Displacement of the Pacific ITCZ during the warm and cold phases of ENSO has been suggested as a possible cause of strengthening or weakening of the MSD (Magana et al. 1999; Curtis 2002). In this paper the seasonal migration of the Pacific ITCZ is also found to be important to the MSD. Future work will study the interannual variability of MSD and its dependence on ITCZ location using longer observational records and idealized model simulations.

The analysis in this paper focuses on the climatological nature of the MSD and its forcing factors to provide an overall description of the phenomenon. In reality the MSD likely includes a suppression of the transient rain-bearing systems that normally impact the region. Indeed, a slight drop in hurricane activity in midsummer has been noted by Magana et al. (1999) and Curtis (2002) in the eastern Pacific and by Inoue et al. (2002) in the Caribbean Sea. By contrast, the timing of the peak in frequency of hurricanes in the tropical Atlantic (around 10 September; Landsea 1993) is close to the time of phase 3, when the tropical SST is warmest in the northern Atlantic and Caribbean Sea (Fig. 7f), one of the favorable conditions for tropical depression activity. As tropical depressions are significant contributors to the broad positive precipitation band observed over the Atlantic in phase 3 (Fig. 10e), they may affect the termination of MSD by inducing the type of broad cyclonic response seen in Fig. 11e.

Maloney and Hartmann (2000) showed that low-level easterlies (westerly) anomalies across the Central

³ Inoue et al. (2002) noted a midsummer minimum in hurricanes in the southwest Caribbean Sea and suggested that cooler SST in that period was one of the contributing factors.

Americas in the Madden–Julian oscillation (MJO) were coincident with reduced (increased) tropical depression activity and precipitation in the eastern Pacific and in the Gulf of Mexico. These results suggest that the MJO may modulate the MSD. The relationship between the MSD and the somewhat shorter period MJO (30–60 day; Madden and Julian 1994) is currently being investigated by the authors.

Experiments with the International Pacific Research Center (IPRC) regional ocean–atmosphere model (IROAM; Xie et al. 2007) confirm the Atlantic influence on the Pacific and Central America and further predict a small effect of local SST variability on the MSD. The control run of the coupled model and of its atmosphere-only equivalent give a fairly realistic representation of the eastern Pacific mean state and annual cycle (see Xie et al. 2007 for discussion), as well as a MSD with many similar characteristics to those observed. In a sensitivity experiment where the atmospheric component of this regional model is forced with observed SST filtered to remove intraseasonal variability (on MSD and MJO time scales), the MSD was not significantly different from the control, supporting the hypothesis that SST variability is not essential for the MSD. In another experiment, the response to an imposed SST cooling over an extensive region of the tropical North Atlantic, constant in time, shows a precipitation decrease and intensification of the subtropical high in the tropical Atlantic (Xie et al. 2007). The positive sea level pressure anomalies penetrate into the eastern Pacific, triggering a precipitation anomaly pattern in summer similar to the MSD. This occurs despite small or even positive SST anomalies in the eastern Pacific.

8. Conclusions

The presence of a midsummer minimum of precipitation in Central America and the adjacent seas has been studied using satellite observations, reanalysis data, and a linear baroclinic model. The following conclusions have been drawn.

Before the MSD is established, in late June, solar heating is at a maximum in the Northern Hemisphere and a large thermal low forms over the North American continent. Associated geostrophic winds around this low pressure anomaly flux moisture from the Pacific ITCZ toward the Gulf of Tehuantepec and Central America, which experience an early season peak in rainfall.

By midsummer (July–August) the ITCZ has moved north in the eastern Pacific following the seasonal march of SST. The combination of the movement of the

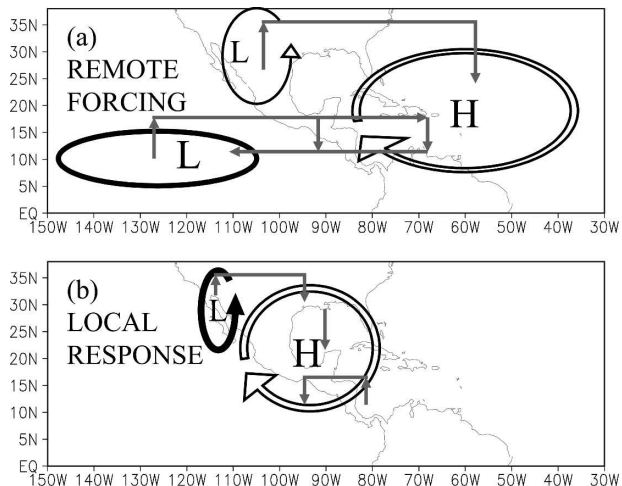


FIG. 15. Schematic showing the main processes described in the text that govern the midsummer drought. Here thick gray lines and arrows denote circulation in the vertical plane (such as Walker cells), thick solid ellipses denote enhanced rain and low pressure, parallel-lined ellipses denote reduced rain and high pressure, and (a) a thin solid ellipse denotes a thermal low. The sense of cyclonic or anticyclonic circulation is marked by arrows. (a) Remote forcing from the Atlantic and Pacific Oceans and North American landmass and (b) local forcing from precipitation deficit over MSD region and enhanced rainfall over Sierra Madre.

Pacific ITCZ and westward extension of the Atlantic subtropical high sets up easterly anomalies across the Caribbean Sea to the eastern Pacific, as shown schematically in Fig. 15a. These easterly winds are associated with subsidence and divergence over the Central American region, as shown in a numerical model, giving rise to the MSD. The MSD is self-sustaining in that the circulation anomaly induced by the MSD precipitation deficit induces northeasterlies across Central America and induces low-level divergence and subsidence drying (schematically shown in Fig. 15b).

The seasonal rains return in September, when the SST in the Atlantic and Caribbean Sea reaches its peak, ITCZs reach their northernmost point in the eastern Pacific and western Atlantic, and tropical depression activity also peaks. Associated with the consequent enhanced precipitation over a broad range of the tropical and subtropical Atlantic (and in the eastern Pacific off Mexico), surface low pressure anomalies exist in the subtropical Atlantic and Pacific, and there is westerly flow from the Pacific toward the Atlantic over Central America. The low pressure conditions give rise to low-level convergence and Central America experiences its second pluvial of the summer.

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