

Impact of Tropical Easterly Waves on the North American Monsoon

JENNIFER L. ADAMS

Cooperative Institute for Mesoscale Meteorological Studies, University of Oklahoma, and NOAA/National Severe Storms Laboratory, Norman, Oklahoma

DAVID J. STENSRUD

NOAA/National Severe Storms Laboratory, Norman, Oklahoma

(Manuscript received 24 August 2005, in final form 5 June 2006)

ABSTRACT

The North American monsoon (NAM) is a prominent summertime feature over northwestern Mexico and the southwestern United States. It is characterized by a distinct shift in midlevel winds from westerly to easterly as well as a sharp, marked increase in rainfall. This maximum in rainfall accounts for 60%–80% of the annual precipitation in northwestern Mexico and nearly 40% of the yearly rainfall over the southwestern United States. Gulf surges, or coastally trapped disturbances that occur over the Gulf of California, are important mechanisms in supplying the necessary moisture for the monsoon and are hypothesized in previous studies to be initiated by the passage of a tropical easterly wave (TEW). Since the actual number of TEWs varies from year to year, it is possible that TEWs are responsible for producing some of the interannual variability in the moisture flux and rainfall seen in the NAM.

To explore the impact of TEWs on the NAM, four 1-month periods are chosen for study that represent a reasonable variability in TEW activity. Two continuous month-long simulations are produced for each of the selected months using the Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model. One simulation is a control run that uses the complete boundary condition data, whereas a harmonic analysis is used to remove TEWs with periods of approximately 3.5 to 7.5 days from the model boundary conditions in the second simulation. These simulations with and without TEWs in the boundary conditions are compared to determine the impact of the waves on the NAM. Fields such as meridional moisture flux, rainfall totals, and surge occurrences are examined to define similarities and differences between the model runs. Results suggest that the removal of TEWs not only reduces the strength of gulf surges, but also rearranges rainfall over the monsoon region. Results further suggest that TEWs influence rainfall over the Southern Plains of the United States, with TEWs leading to less rainfall in this region. While these results are only suggestive, since rainfall is the most difficult model forecast parameter, it may be that TEWs alone can explain part of the inverse relationship between NAM and Southern Plains rainfall.

1. Introduction

During the warm season, there is a distinct maximum in rainfall over the southwestern United States and northwestern Mexico due to the North American monsoon (NAM). The NAM accounts for 60%–80% of the annual rainfall in northwestern Mexico and nearly 40% of the annual rainfall in Arizona (Douglas et al. 1993). Several areas in Mexico receive greater than 900 mm of rain during the NAM, and along the slopes of the Sierra

Madre Occidental (SMO) the ground cover drastically changes from desert-like to tropical in a matter of weeks (Douglas et al. 1993).

The NAM has a distinctive life cycle that includes development, mature, and decay phases (Higgins et al. 1997). The NAM starts in May and June over southern Mexico. Precipitation amounts amplify over southern Mexico and then quickly move northward along the western slopes of the SMO in June and early July. The mature phase of the NAM begins in July and continues through August. Surges of tropical air toward the north supply ample moisture for increased rainfall near and to the south of the monsoon anticyclone (Higgins et al. 1997). In association with the development of the mature phase of the NAM, there is an observed decrease

Corresponding author address: David J. Stensrud, National Severe Storms Laboratory, National Weather Center, Room 4368, 120 David L. Boren Blvd., Norman, OK 73072.
E-mail: David.Stensrud@noaa.gov

in Great Plains rainfall (Tang and Reiter 1984; Douglas et al. 1993; Mock 1996; Higgins et al. 1997). Analyses of precipitation data indicate that the rainfall in the Southern Great Plains begins to decrease a few days prior to NAM precipitation onset in Arizona and New Mexico (Higgins et al. 1997). Modeling results suggest that the decrease of Southern Great Plains rainfall is associated with a weakening of the low-level jet and its associated moisture transport near the Texas coast (Mo and Berbery 2004).

One atmospheric phenomenon that transports NAM low-level moisture northward is the gulf surge, a coastally trapped wave that develops over the Gulf of California and propagates northward. The gulf surge transport mechanism has grown to become widely accepted in recent years (Hales 1972, 1974; Brenner 1974; Douglas et al. 1993, 1998; Douglas 1995; McCollum et al. 1995; Stensrud et al. 1995, 1997; Schmitz and Mullen 1996; Fuller and Stensrud 2000; Berbery 2001; Berbery and Fox-Rabinovitz 2003; Douglas and Leal 2003; Saleeby and Cotton 2004). Generally, gulf surges appear to be initiated by the passage of a TEW across the southern end of the Gulf of California (Hales 1972; Brenner 1974; Stensrud et al. 1997) and vary in both intensity (Hales 1972; Stensrud et al. 1997) and initiation region (Stensrud et al. 1997). Surge events are characterized by a net transport of cool, moist air northward using the Gulf of California as a natural channel, bounded by Baja California to the west and the SMO to the east.

The relationship between gulf surges and tropical easterly waves (TEWs) is investigated by Fuller and Stensrud (2000) using the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis data. They find that nearly three-fourths of gulf surges are associated with TEWs. Possibly related to fluctuating surge frequency and intensity, the NAM also has a great deal of interannual variability. Higgins et al. (1998) term these differences “wet” and “dry” monsoons. Wet monsoons have a much longer period of heavy rainfall compared to dry monsoons, with the variability linked to the large-scale circulation. Surges also are found to change the precipitation pattern over Mexico and the United States (Higgins et al. 2004). Northwestern Mexico and the Southern Great Plains regions are anomalously dry prior to surge passage into Arizona. At the time of surge passage, northwestern Mexico becomes anomalously wet, while Arizona becomes anomalously wet after surge passage into Arizona. The Southern Great Plains region typically remains anomalously dry after surge passage (Higgins et al. 2004).

Mesoscale model simulations have shown success in

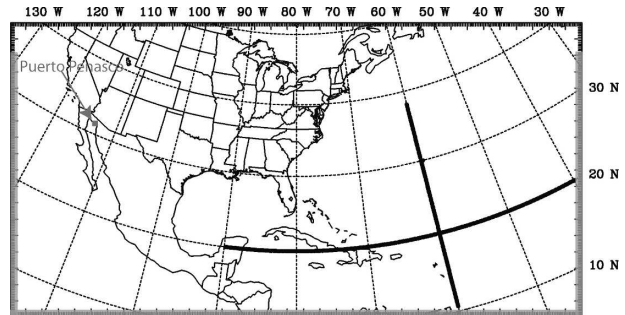


FIG. 1. The model domain using 25-km grid spacing. The location of Puerto Penasco, Mexico, also is shown at the northern end of the Gulf of California. The gray zone along the western, eastern, and southern boundaries indicates the grid points for which a harmonic analysis is conducted to remove TEWs from the model boundary conditions for the no-TEW simulation. The darker line along 20°N shows the region along which Hövmoller diagrams are created (see Fig. 3), while the darker line at 60°W indicates the longitude selected for display of data in Fig. 2.

simulating many of the important NAM features, including NAM precipitation patterns, gulf surges, Gulf of California low-level jets, and the diurnal cycle of precipitation, when provided with good large-scale boundary conditions (Stensrud et al. 1995, 1997; Gochis et al. 2002; Saleeby and Cotton 2004; Mo and Berbery 2004; Gutzler et al. 2005). Thus, to further explore the influence of TEWs on gulf surges and the NAM, four month-long control runs of a mesoscale model are compared to similar month-long simulations where TEWs are by and large removed from the model boundary conditions. Through careful analysis of several different meteorological fields, many similarities and differences are noted between the paired simulations and discussed. It is hoped that these comparisons lead to an improved understanding of how TEWs influence the NAM. Section 2 describes the model used for the month-long simulations, while the experimental methodology is discussed in section 3. Results from the model simulations of gulf surges are presented in section 4, followed by an examination of model rainfall totals in section 5. A final discussion is found in section 6.

2. Mesoscale model description

The model used to produce continuous month-long simulations is the nonhydrostatic fifth-generation Pennsylvania State University–National Center for Atmospheric Research (PSU–NCAR) Mesoscale Model (MM5; Dudhia 1993). A one-way grid domain of $350 \times 180 \times 23$ (x, y, σ) is created with a horizontal grid spacing of 25 km (Fig. 1). Ten of the 23-sigma levels in this study are contained below 700 hPa to better cap-

ture gulf surge events. The model configuration includes a five-layer soil model, the Kain–Fritsch convective parameterization scheme (Kain and Fritsch 1990), the Hong–Pan nonlocal planetary boundary layer scheme (Hong and Pan 1996), a simple water and ice explicit microphysics scheme (Dudhia 1989), the Rapid Radiative Transfer Model longwave radiation scheme (Mlawer et al. 1997), and the Dudhia (1989) shortwave radiation scheme. Each of the continuous month-long simulations begins at 0000 UTC on the first day of the month and ends at 1800 UTC on the last day of the month.

Model initial and boundary conditions are specified using the National Centers for Environmental Prediction (NCEP)–NCAR reanalysis data produced by a T62 (210-km resolution) model (Kalnay et al. 1996). These data are available at 6-h increments with a horizontal grid spacing of approximately $2.5^\circ \times 2.5^\circ$. The data are interpolated onto the 25-km mesoscale model grid at the start time to provide the model initial conditions and also at each subsequent 6-h interval to provide boundary conditions throughout the month-long simulation. The Davies and Turner (1977) relaxation method is used on the model boundaries, in which the model variables on the outermost four grid points are nudged toward the reanalysis data. Reanalysis data do not include sea surface temperatures (SSTs), therefore SST analyses are obtained from NCEP operational analyses. Unfortunately, the SST analyses do not capture the significant warming of the Gulf of California during the summer months (Stensrud et al. 1995), making it necessary to further modify the water temperatures over this region. A constant value of 29.0°C is assumed over the Gulf of California for all simulations.

Error growth in limited-area models is strongly limited by the imposed model boundary conditions (Anthes et al. 1985; Paegle et al. 1997). Simulations from limited-area models can reproduce well many of the observed features if the boundary conditions are good but cannot overcome errors in the global data that drive the boundary conditions (Mo et al. 2005). Thus, it is not surprising that a number of studies indicate that NAM circulations are reasonably well simulated by limited-area mesoscale models that use observations or reanalysis data as boundary conditions (Stensrud et al. 1995; Anderson et al. 2000; Gochis et al. 2002; Saleeby and Cotton 2004; Mo and Berbery 2004; Gutzler et al. 2005). In particular, Gochis et al. (2002) produce a 2.5-month continuous simulation with MM5 during the 1999 NAM season using reanalysis data as boundary conditions. Comparisons between simulated and observed soundings and rainfall totals during July are made and indicate that the MM5 with the Kain–Fritsch

convective scheme does a reasonable job in simulating both NAM rainfall and monthly mean soundings over Mexico. Thus, there is reason to expect that the MM5 can be used to examine the influence of TEWs on the NAM.

3. Methodology

To explore the influence of TEWs on the NAM, it is important to sample months with different numbers of TEWs crossing western Mexico. In addition, it is helpful if these months have different total monthly rainfall distributions over the NAM region. Thus, TEW activity and total monthly rainfall totals from 1985 through 1993 are examined during the mature phase of the NAM life cycle. The number of TEWs crossing Mexico each month is determined by examining Hövmoller diagrams of the 850-hPa meridional wind component at 20°N [as in Fuller and Stensrud (2000)] from the ECMWF reanalysis data. While 700 hPa is the most desirable atmospheric layer to use in determining TEW passage (Reed et al. 1977), 850 hPa is the closest layer available in the locally available reanalysis data, and this level still provides a clear indication of the presence of TEWs. Rainfall patterns over the NAM region are obtained from the Climate Prediction Center (CPC) Unified Precipitation Data Set over the United States and gauge stations in Mexico at a resolution of one degree longitude by one degree latitude. There are 300 gauge stations included in the Mexican rainfall data prior to 1990 and approximately 600 thereafter (Higgins and Shi 2000). Four 1-month periods then are chosen subjectively that provide a reasonable variation in the number of TEWs that pass over western Mexico during each month and also have different observed monthly rainfall distributions. After examining the Hövmoller diagrams and the monthly rainfall totals, the months selected for further study are July 1990, July 1992, August 1986, and August 1988.

The month of July 1990 is chosen because it is well known through both observational and modeling studies to have produced a strong NAM (Stensrud et al. 1995, 1997; Douglas 1995). The ECMWF reanalysis data indicate the presence of four TEWs during this month while the CPC precipitation dataset shows NAM precipitation reaching into Arizona. August 1988 also is an active month, with five TEWs crossing Mexico and the northern extent of the NAM extending into both Arizona and New Mexico. July 1992 and August 1986 are both less active, with only one or two TEWs revealed in the Hövmoller diagrams. The relative TEW inactivity of these two months is associated with a shift in the northern extent of the NAM rainfall toward the south.

Continuous month-long simulations (control runs) of each of the four months selected are produced and the model output is saved every 6 h. To remove the effects of the TEWs, a harmonic analysis is conducted on the 6-h data used to create the model boundary conditions (discussed below). The end result of the harmonic analysis is that TEWs are removed from the model boundary conditions from the sea surface to the model domain top for latitudes less than 30°N. This allows for the generation of separate month-long simulations in which the effects of TEWs are largely removed and the simulations with and without TEWs can be compared. The studies of Anthes et al. (1985) and Paegle et al. (1997) provide evidence that removing the effects of TEWs from the model boundary conditions is a reasonable approach to studying their role in the NAM. Both of these studies clearly show that the boundary conditions imposed upon limited-area models strongly limit error growth and have significant control over the evolution of features within the model domain. Thus, by altering the model boundary conditions to remove TEWs, we can evaluate how the model atmosphere evolves without their presence and understand better the role of TEWs on the NAM.

Harmonic analysis on boundary condition data

A harmonic analysis is performed to remove TEWs from the boundary conditions provided to the meso-scale model in order to explore the link between gulf surges, NAM rainfall, and TEWs. Reed et al. (1977) examine eight different TEWs and, using a compositing method, determine that these waves have an average wavelength of 2500 km, a period of approximately 3.5 days, and an average easterly propagation speed of 8 m s⁻¹. In addition, TEWs are most pronounced in the horizontal wind field at 700 hPa, where there is a cyclonic circulation with a distinct northeast–southwest tilt. Burpee (1972, 1974) also examines TEWs and finds wave periods between 3.5 and 7.5 days. Thus, at most eight TEWs can pass across Mexico during a 1-month period, although months with no TEW passages also are observed.

The model boundary conditions are defined from the NCEP–NCAR reanalysis data and applied using the Davies and Turner (1977) relaxation method. The relaxation method uses the outermost four grid points along all four outer edges of the model domain to nudge the model variables toward their known boundary values. Thus, to prevent TEWs from influencing the model simulations, month-long time series of each of the model variables (temperature, water vapor mixing ratio, geopotential height, *u*-wind component, *v*-wind component, and sea level pressure) are constructed at

each grid point on the four outermost rows of grid points on the southern, eastern, and western model boundaries and at each of the 23 vertical model levels. This leads to the creation of over 65 000 month-long time series for each three-dimensional model variable. These time series are then passed to a harmonic analysis, which removes the waves with periods typically associated with TEWs. The modified time series as reconstructed without TEWs are used to create new boundary conditions.

Since the model data used to create the boundary conditions are both equally spaced in time (every 6 h) and contain no missing values, the model data can be represented exactly given a series of *n* points in time by summing a series of *n*/2 harmonic functions (Wilks 1995)

$$y_t = \bar{y} + \sum_{k=1}^{n/2} \left[A_k \cos\left(\frac{2\pi kt}{n}\right) + B_k \sin\left(\frac{2\pi kt}{n}\right) \right], \quad (1)$$

where

$$A_k = \frac{2}{n} \sum_{t=1}^n y_t \cos\left(\frac{2\pi kt}{n}\right), \quad (2)$$

$$B_k = \frac{2}{n} \sum_{t=1}^n y_t \sin\left(\frac{2\pi kt}{n}\right). \quad (3)$$

Please see Wilks (1995) for additional details, especially how to handle the coefficients when $k = n/2$. Based on the characteristics of TEWs documented by both Reed et al. (1977) and Burpee (1972, 1974), waves with periods between 3.5 to 7.5 days are removed from the time series by identifying and replacing the amplitudes A_k and B_k of waves representing TEWs with a value of zero south of 30°N and then linearly increasing the amplitudes A_k and B_k back to their original values between 30° and 35°N (see Fig. 1). The modified boundary condition data with TEWs removed from the surface to model top along the southern, eastern, and western boundaries are used as input to the MM5 in the same manner as the control run and month-long simulations produced for each of the four selected months. These simulations with TEWs removed from the boundary conditions are called no-TEW runs.

Graphs of the various harmonic amplitudes and Hövmoller diagrams of the meridional wind component at 700 hPa both verify that TEWs are damped if not entirely removed in the no-TEW runs south of 30°N (Figs. 2, 3). In general, TEW wave amplitudes greater than 3 m s⁻¹ contained in the control run (Fig. 2a) are decreased to less than 1.0 m s⁻¹ in the no-TEW simulation

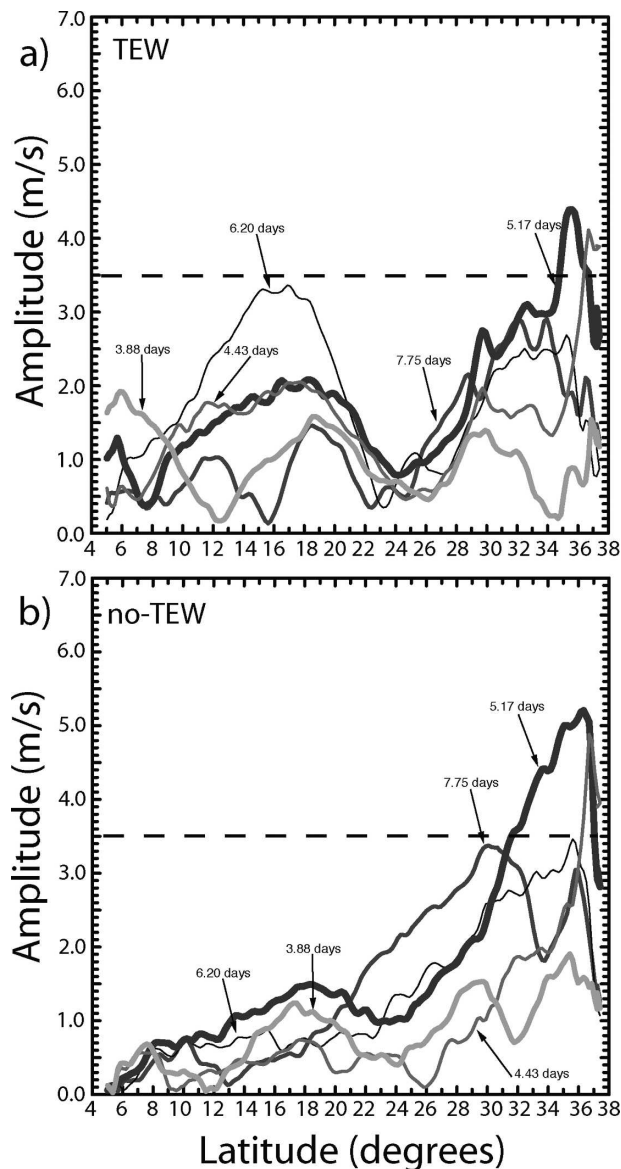


FIG. 2. Amplitudes (m s^{-1}) of the meridional wind component vs latitude ($^{\circ}\text{N}$) for waves with periods between 7.75 and 3.88 days along 60°W longitude (location indicated in Fig. 1) from the August 1988 (a) control and (b) no-TEW runs. Note how the waves with periods between 7.5 and 3.5 days are decreased in (b) compared to (a) below 30°N , especially the dominant 6.2-day wave period.

for periods between 3.5 and 7.5 days and for latitudes less than 30°N (Fig. 2b). In particular, the dominant TEW wave period is 6.2 days during August 1988 and waves with this period are largely absent from the no-TEW run (cf. Figs. 2a,b). The TEW wave amplitudes are not expected to be exactly zero in the no-TEW runs away from the model boundaries, since models are able to generate waves at scales not present in the initial or

boundary conditions (Tribbia and Baumhefner 2004). In addition, troughs that move around the midlatitude high over the United States and into the Tropics will not be removed by this harmonic analysis on the model boundaries. The wave amplitudes for latitudes greater than 30°N and for periods greater than 7.5 days sometimes are enhanced in the no-TEW runs, suggesting that the model response to the removal of TEWs is complex and deserves further study. Plots of model data every 6 h suggest that the harmonic analysis creates no problems along the boundaries of the model domain, and midlatitude features are clearly seen in both runs.

Hövmoller diagrams (Fig. 3) depict TEWs as features that move from the upper right to lower left, in zones where the wind shifts from northerly to southerly. The difference field (Fig. 3c) clearly shows that many of the TEWs in the control run are absent from the no-TEW run. However, since harmonic analysis is done only on the boundary conditions, a TEW contained in the initial conditions cannot be removed and is contained within both model simulations. Comparisons with the ECMWF analyses indicate that the same number of TEWs are contained in the MM5 control run as in the ECMWF data except during July 1992 (Table 1).

4. Simulated gulf surges

Since TEWs have been shown to be important to the development of gulf surges, we begin our exploration of the model simulations by examining the gulf surges produced in the two runs. To get an initial idea of the number and frequency of simulated surge occurrences, time series of various meteorological parameters are generated using MM5 output data at Puerto Penasco, Mexico (Fig. 1). Puerto Penasco is located at the far northern end of the Gulf of California, making it a prime location to capture most gulf surge events. The fields that are examined include lowest sigma level temperature and dewpoint, wind direction, and wind speed. Potential surges are identified when the time series indicate a southerly wind shift, a maximum daily dewpoint exceeding 65°F (18.3°C) for at least two days, peak wind speeds greater than 5 m s^{-1} , and a decrease in maximum temperature of at least 5°F (2.8°C) from the previous day. After examining the time series at Puerto Penasco, plots of horizontal wind vectors and water vapor mixing ratio contours are examined at the lowest sigma level to verify whether an actual surge is occurring or if any are missed by the time series plots.

Over the four months examined, 18 surges are observed in the control runs of the model simulations. The surge events exhibit varying degrees of strength and

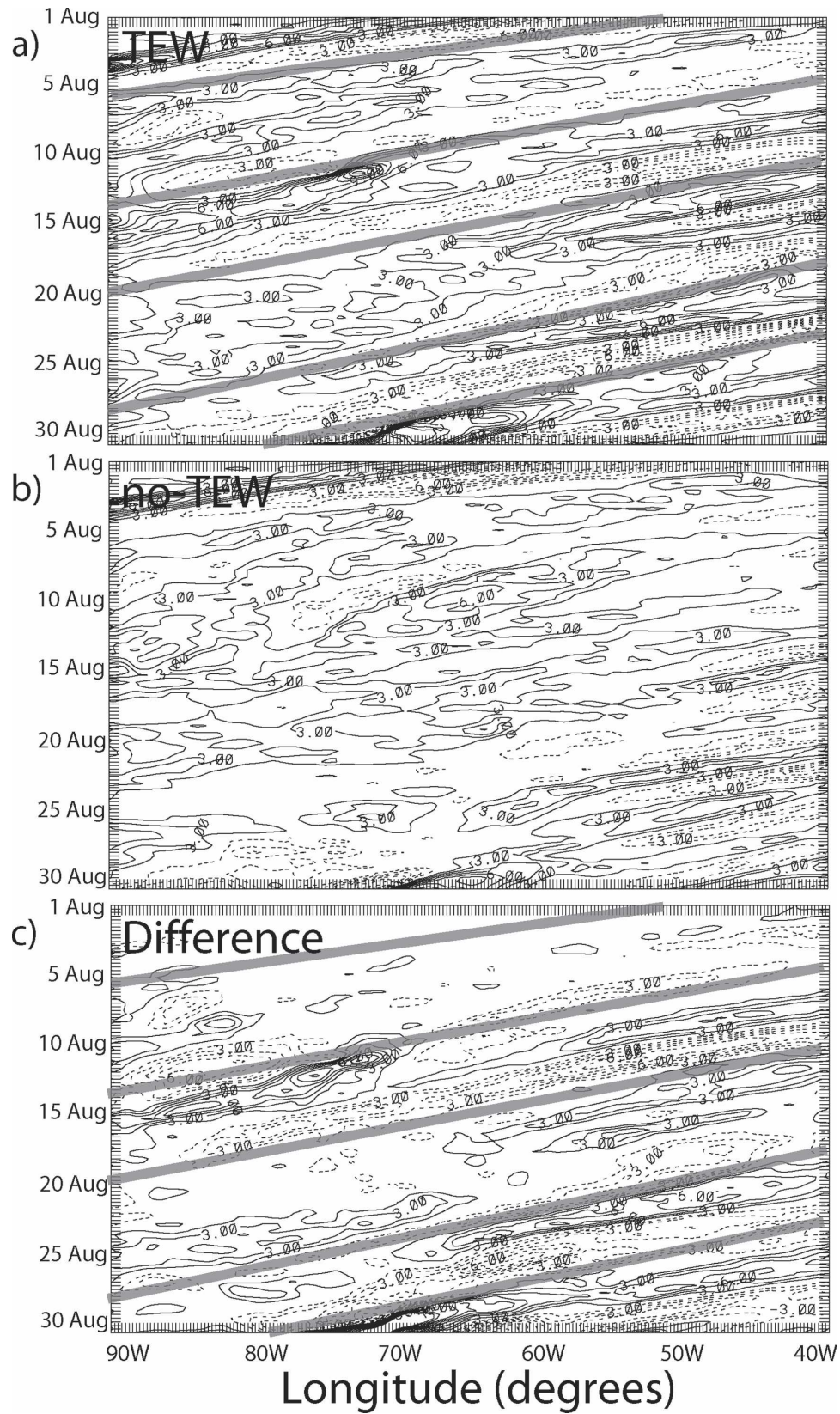


TABLE 1. Number of TEWs in the ECMWF reanalysis data compared to the MM5 control runs.

	ECMWF reanalysis data	MM5 control run
August 1986	3	3
August 1988	5	5
July 1990	4	4
July 1992	1	2

frequency with 17 of the surges associated with either a TEW or a tropical storm. In several of the cases, surges also are present in the model run with waves removed from the boundary conditions; however, their timing and strength are altered. Two surges that represent 17 of the 18 surges seen in this dataset are described to convey the two major types of surges (full and partial) seen in the model simulations and to compare surge evolution in the control and no-TEW runs. A full gulf surge occurs when the entire Gulf of California region has southeasterly low-level winds and cool, moist tropical air is advected along the entire gulf region. A partial gulf surge occurs when the low-level winds in only the northern gulf region become southeasterly.

a. Partial gulf surge

Of the five surges present in the July 1990 control run, all are detected by the Puerto Penasco time series data (Fig. 4). One surge is not associated with a TEW during this month, making it the only such surge in the entire study. All surges affecting Yuma, Arizona, in the control run also are seen in the observations at Yuma.¹ A partial gulf surge occurs in conjunction with a TEW exiting the western coast of Mexico at 0000 UTC 12 July (Fig. 5a). This TEW originally enters the model domain through the eastern model boundary over the Atlantic Ocean near 40°W many days prior to the surge. As this TEW moves over the northern end of the Gulf of California, winds shift to southerly over the northern gulf at the same time as the wave passage (Fig. 6b) in association with a surface trough that precedes

¹ Yuma, Arizona, is the closest National Weather Service surface observation station to the Gulf of California, and data from Yuma are often used to study gulf surges.

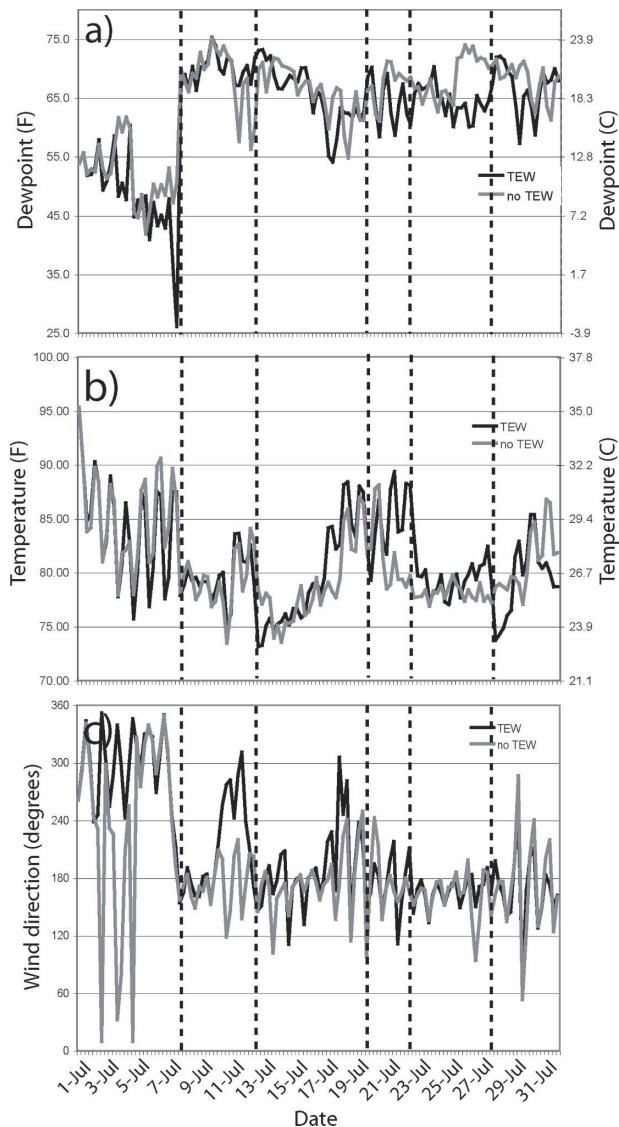


FIG. 4. Time series plots of model-simulated lowest model level (a) dewpoint temperature, (b) temperature, and (c) wind direction at Puerto Penasco, Mexico, during July 1990 from the control run. Black lines are from the control run with TEWs, while the gray lines are from the no-TEW runs. Vertical dashed lines indicate the diagnosed onset of surge events from the control run.

the main wave at 700 hPa. While the remainder of the gulf itself has difficulty shifting to southerly winds, the winds are southerly over the western slopes of the SMO (Fig. 6c). Moisture immediately streams northward into

FIG. 3. Hövmøller diagrams of the meridional wind component (m s^{-1}) at 20°N for the August 1988 (a) control and (b) no-TEW model runs, and (c) their difference (control – no-TEW). Dashed (solid) contours indicate northerly (southerly) winds. Westward-moving waves are indicated by coherent wind zones moving from the upper right to the lower left in the diagram. Gray solid lines indicate the location of wave troughs in (a) and (c). Location of the cross section along 20°N shown in Fig. 1. Isolines every 1.5 m s^{-1} with the zero line omitted. Note the waves indicated in (c) with northerly winds prior to trough passage and southerly winds after trough passage, indicating that the harmonic analysis is largely removing TEWs.

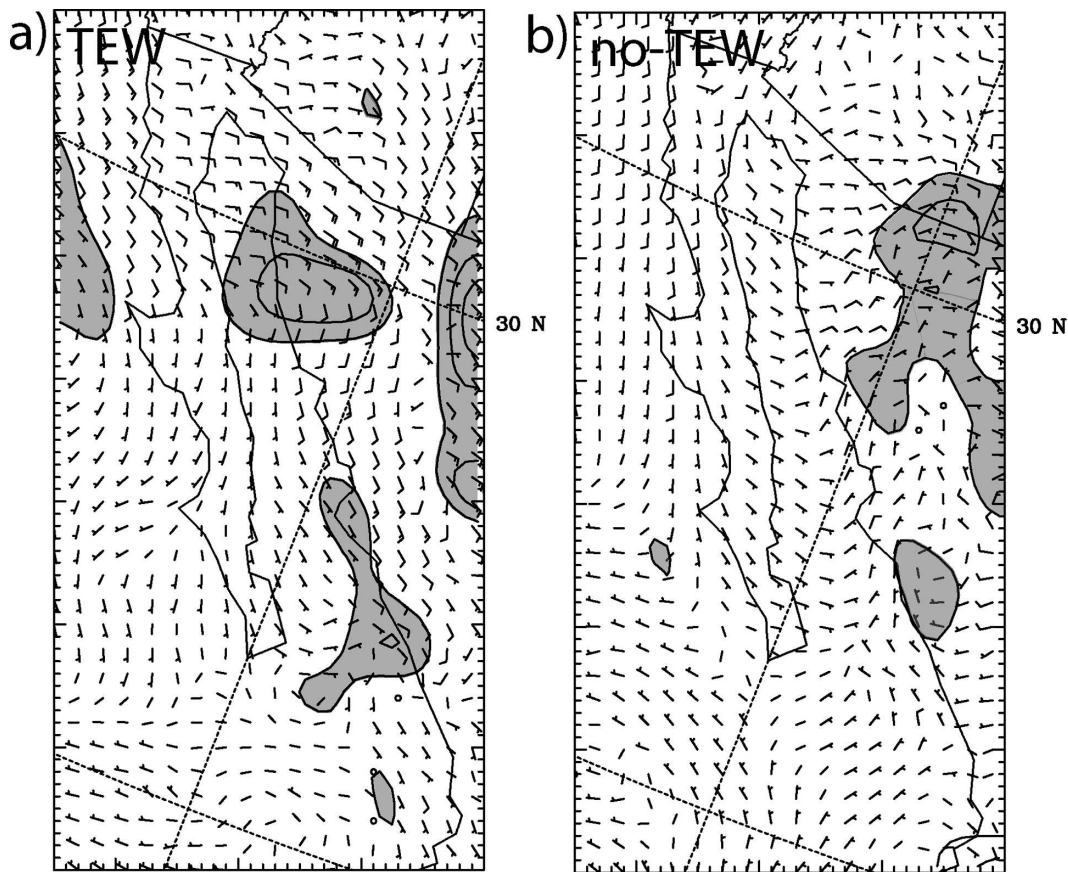


FIG. 5. Relative vorticity isolines and wind barbs at 700 hPa in the (a) control and (b) no-TEW runs at 0000 UTC 12 July 1990. Relative vorticity isolines every $2 \times 10^{-5} \text{ s}^{-1}$ starting from $2 \times 10^{-5} \text{ s}^{-1}$, and full wind barbs are equivalent to 5 m s^{-1} .

Arizona and southern California as the winds shift. The elevated levels of water vapor mixing ratio near the surface persist for nearly 42 h, through approximately 1800 UTC 13 July (Fig. 6d) when the effects of the surge across the desert southwest become ill defined. This partial gulf surge pattern is present in 7 of the 18 surges identified in this study, making it the second most common type of surge event in the simulations.

In the no-TEW run, the TEW is absent at 0000 UTC 12 July (Fig. 5b). This essentially eliminates all evidence of a gulf surge in this run until 0000 UTC 13 July. At this time, a wave-like feature that originates over Mexico moves across the northern end of the Gulf of California. A northward moisture surge is induced over northwestern Mexico and the southwestern United States approximately 24-h later than in the control run (Figs. 7a–d). However, the winds have difficulty establishing a strong southerly direction as seen in the control run, and the moisture is unable to propagate as far to the west into Arizona. This run illustrates that features that are not TEWs can induce a gulf surge, albeit a weak partial surge in this case.

The other six partial gulf surges seen in the control runs are weaker in the corresponding no-TEW runs and often occur at slightly different times. Results further indicate that the partial gulf surges in the no-TEW runs often are linked to waves produced by the model over the Caribbean Sea and Mexico instead of being linked to TEWs as in the control run. It is uncertain if differences in wave structure, size, and magnitude have any influence on the strength of the resulting gulf surge, or if the gulf surge strength is determined mainly by the local environment over the Gulf of California and the location of the large-scale monsoon anticyclone (Higgins et al. 2004).

b. Full gulf surge

August 1986 is characterized by an NAM system that is most active over northwestern Mexico into Arizona, as indicated by the composite frequency of cloud-top temperatures below -38°C (not shown). Six surges occur in the control run during this month, five of which are induced by a TEW while one is initiated by a tropical storm in the model domain. Every surge is repre-

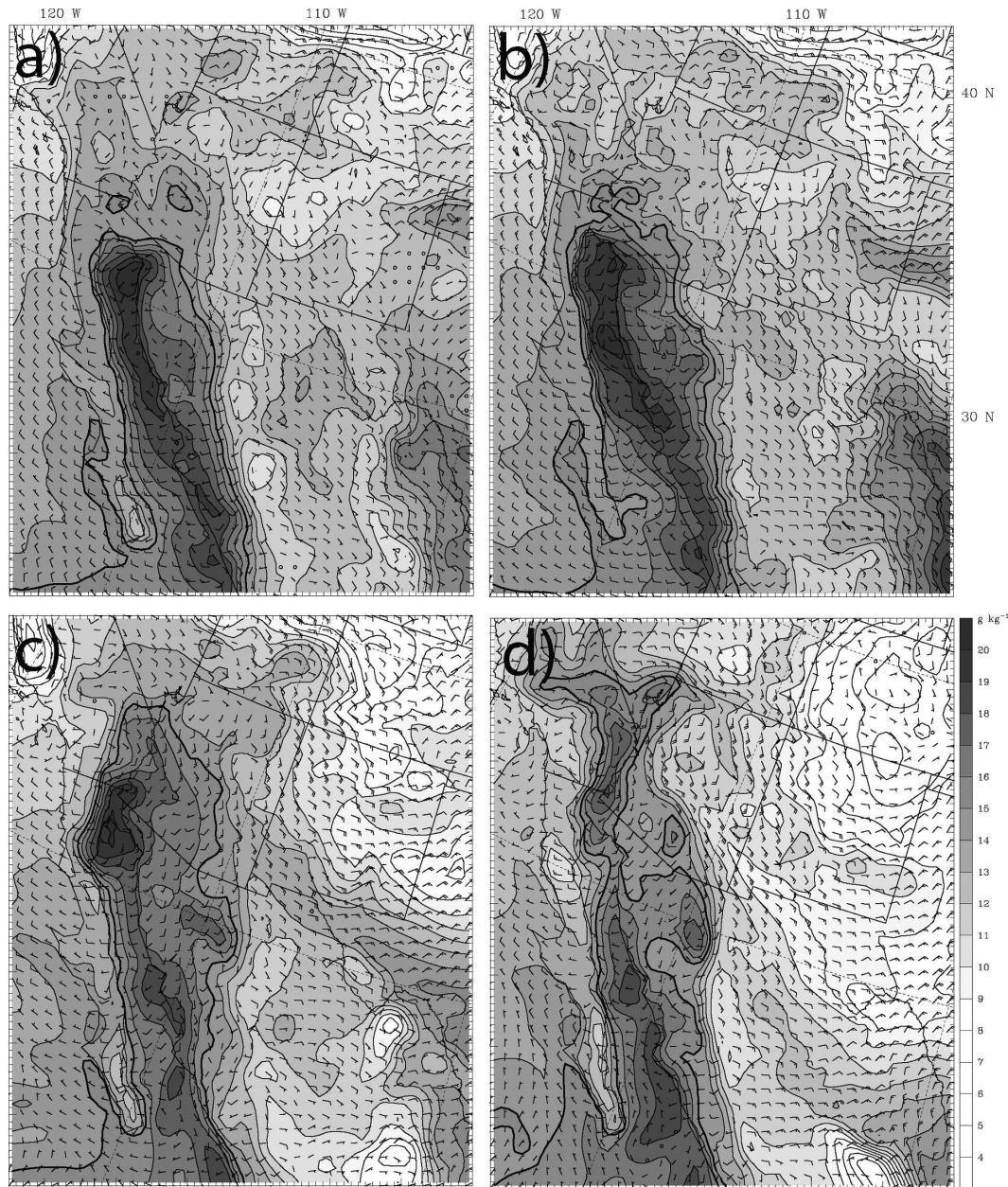


FIG. 6. Isolines of water vapor mixing ratio and wind barbs at $\sigma = 0.995$ from the July 1990 control run, showing (a) presurge (1800 UTC 11 July), (b) surge onset (0000 UTC 12 July), (c) during the surge (1800 UTC 12 July), and (d) postsurge (1800 UTC 13 July) conditions. Isolines of water vapor mixing ratio every 1 g kg^{-1} with the 15 g kg^{-1} isoline shown in bold.

sented in the model time series data at Puerto Penasco (Fig. 8). Five of the six surges are apparent in the observational data from Yuma, although only three of the surges are evident at Yuma in the model data. This comparison emphasizes that the location of Yuma is not optimal for diagnosing gulf surges, even though it is the best site available in the observational dataset.

The final surge of August 1986, which is not present

in the observational data at Yuma, is initiated by the passage of a series of waves across the mouth of the gulf starting on 1200 UTC 26 August (Fig. 9). The surge begins on 0600 UTC 27 August at the southern end of the Gulf of California (Fig. 10b), and by 1200 UTC 27 August the winds over the entire Gulf of California are strong south-southeasterly and a sharp moisture gradient has developed over the northern portion of the gulf

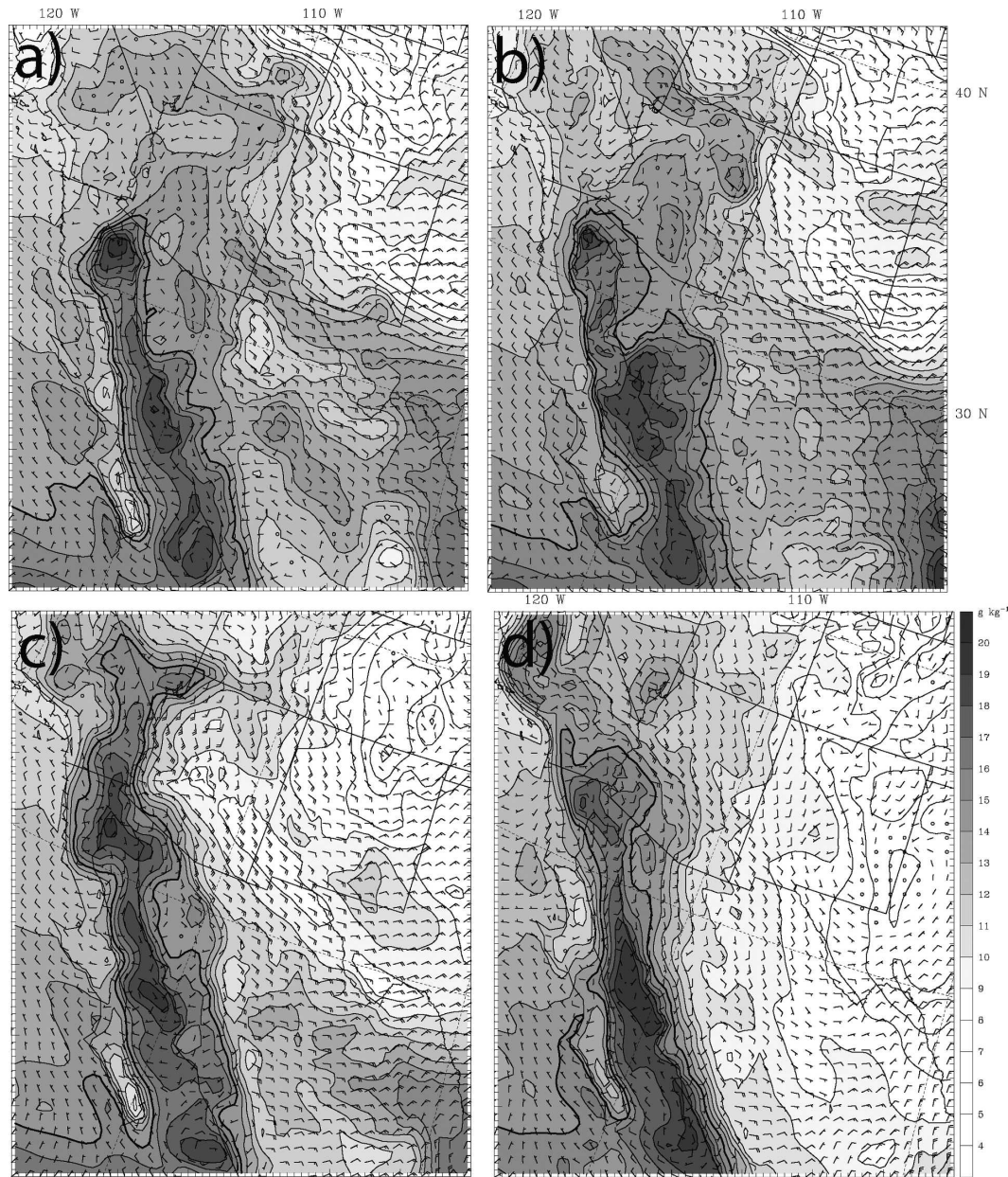


FIG. 7. Same as in Fig. 6, but for the no-TEW run, showing (a) presurge (1800 UTC 12 July), (b) surge onset (0000 UTC 13 July), (c) during the surge (1800 UTC 13 July), and (d) postsurge (1800 UTC 14 July) conditions.

(Fig. 10c). Extremely high values of water vapor mixing ratio are pushed into the desert southwest by strong southerly winds through approximately 1200 UTC 29 August, when the winds finally subside and lose some of their southerly component (Fig. 10d). It is particularly impressive how the region of highest mixing ratios moves northward cohesively during this surge. Full gulf surges are present in 10 of the 18 surge cases, making them the most common in the control run dataset.

A weak wave develops over Mexico and moves off

the west coast of Mexico at 1200 UTC 26 August in the no-TEW run (not shown). However, the winds over the Gulf of California are never able to achieve a southerly direction over the water for an extended period of time. The highest moisture levels at the surface remain confined to the central and southern Gulf of California, thus resulting in the absence of a gulf surge in the simulation without waves in the boundary conditions (Figs. 11a–d). This is the only full gulf surge event where there is no evidence of its presence in the no-TEW run. The

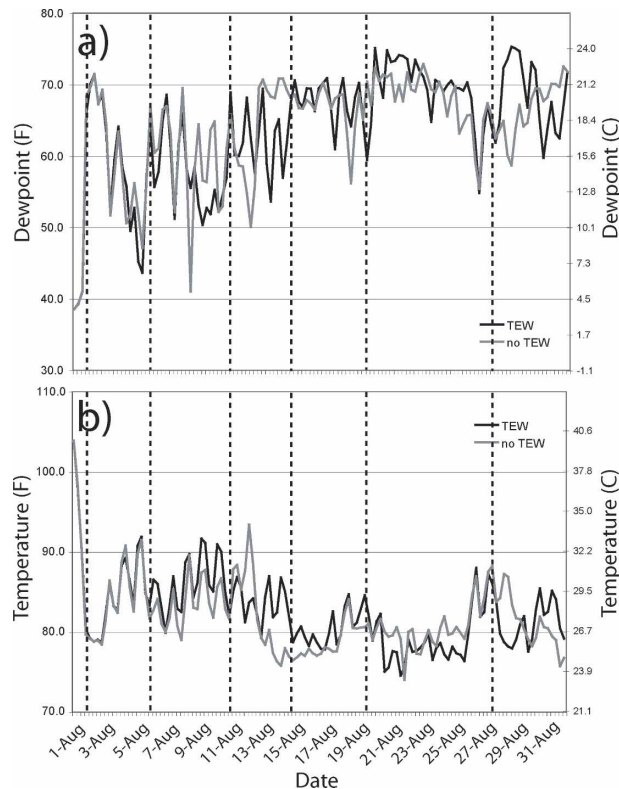


FIG. 8. Same as in Fig. 4, but for (a) dewpoint temperature and (b) temperature during August 1986.

other full gulf surge events seen in the control run are evident to some extent in the no-TEW run, but all are weaker than their control run counterparts. Again, it appears that most of the waves that induce gulf surges in the no-TEW runs develop over the Caribbean Sea and eastern Mexico instead of entering through the eastern boundary of the model domain as a TEW.

5. Simulated monthly rainfall totals

Another important aspect of the NAM that has strong societal and economic consequences is rainfall. Unfortunately, rainfall is one of the most difficult parameters for models to reproduce correctly, and thus the results must be viewed with caution. In general, the control run captures the main features of the NAM precipitation distribution when compared to the $1^\circ \times 1^\circ$ CPC dataset during August 1988 (Fig. 12). The areas of maximum rainfall over Arizona, New Mexico, Colorado, and Mexico are generally in the same location in both the control run and observations, with the simulated rainfall totals roughly twice those from the CPC dataset. Observed monthly rainfall totals exceeding 500 mm are not unusual along the slopes of the SMO (see

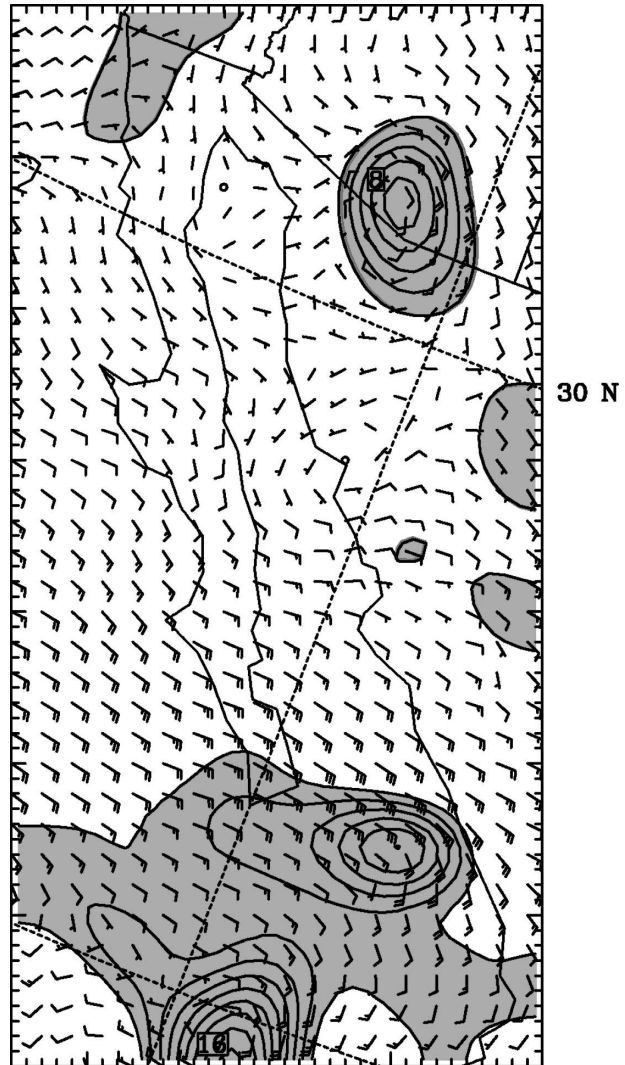


FIG. 9. Isolines of relative vorticity and wind barbs at 700 hPa in the control run at 1800 UTC 26 Aug 1986. Relative vorticity isolines every $2 \times 10^{-5} \text{ s}^{-1}$ starting from $2 \times 10^{-5} \text{ s}^{-1}$, and full wind barbs equivalent to 5 m s^{-1} .

Stensrud et al. 1995; Gochis et al. 2002), so the larger simulated rainfall totals are believed to be reasonable.

The largest discrepancies between the control run and the CPC precipitation analysis are evident over southern Nevada and California, and over eastern Mexico just south of the Texas border. The control run produces precipitation in excess of 200 mm in both these regions that is virtually absent from both the CPC dataset (Fig. 12a) and the no-TEW run (Fig. 12c). These differences may be partially attributed to the fact that the rain gauge network over these areas is extremely sparse, thereby potentially missing many localized heavy precipitation events over the region. A cloud-top temperature composite from this month

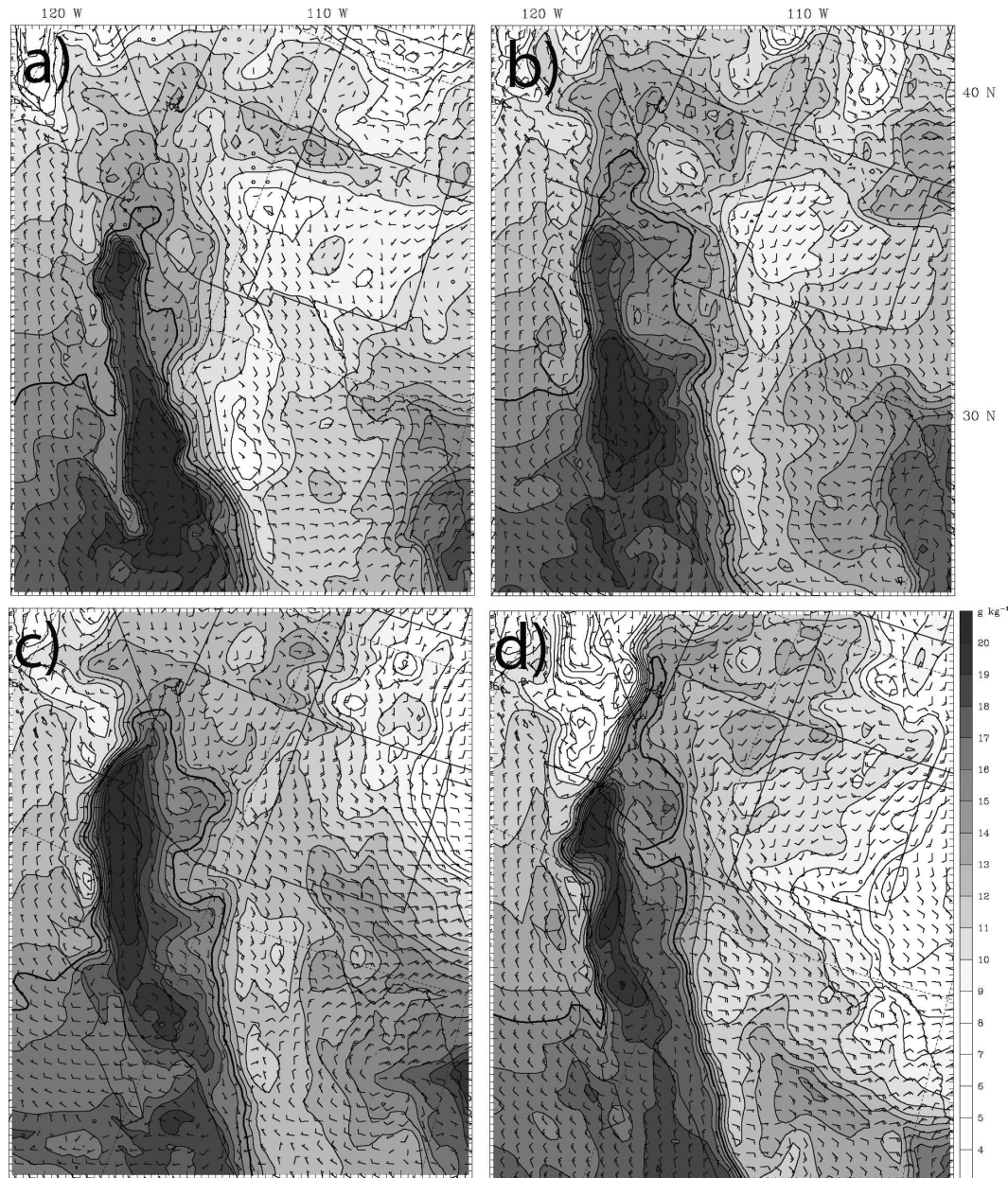


FIG. 10. Same as in Fig. 6, but for the August 1986 control run, showing (a) presurge (1800 UTC 26 August), (b) surge onset (0600 UTC 27 August), (c) during the surge (1200 UTC 28 August), and (d) postsurge (1200 UTC 29 August) conditions.

shows slightly elevated frequencies of temperatures below -38°C over California and Nevada (not shown), and there are numerous reports of flash flooding and heavy thunderstorm rainfall in this area (NOAA 1988), suggesting that convection and rainfall are more common over this region than seen in the CPC analysis. Over eastern Mexico, the cloud-top temperature composite shows several local maxima in the region just south of the Texas border, although widespread rainfall

totals exceeding 200 mm appear unlikely. Thus, the control run clearly produces too much rainfall in eastern Mexico compared to observations.

Monthly simulated rainfall totals over the central plains of the United States (Fig. 13) are reproduced less well than seen over the NAM region. While the control run replicates the southwest-to-northeast zone of heavier rainfall observed in the CPC precipitation analysis stretching from New Mexico to Minnesota, it misses the

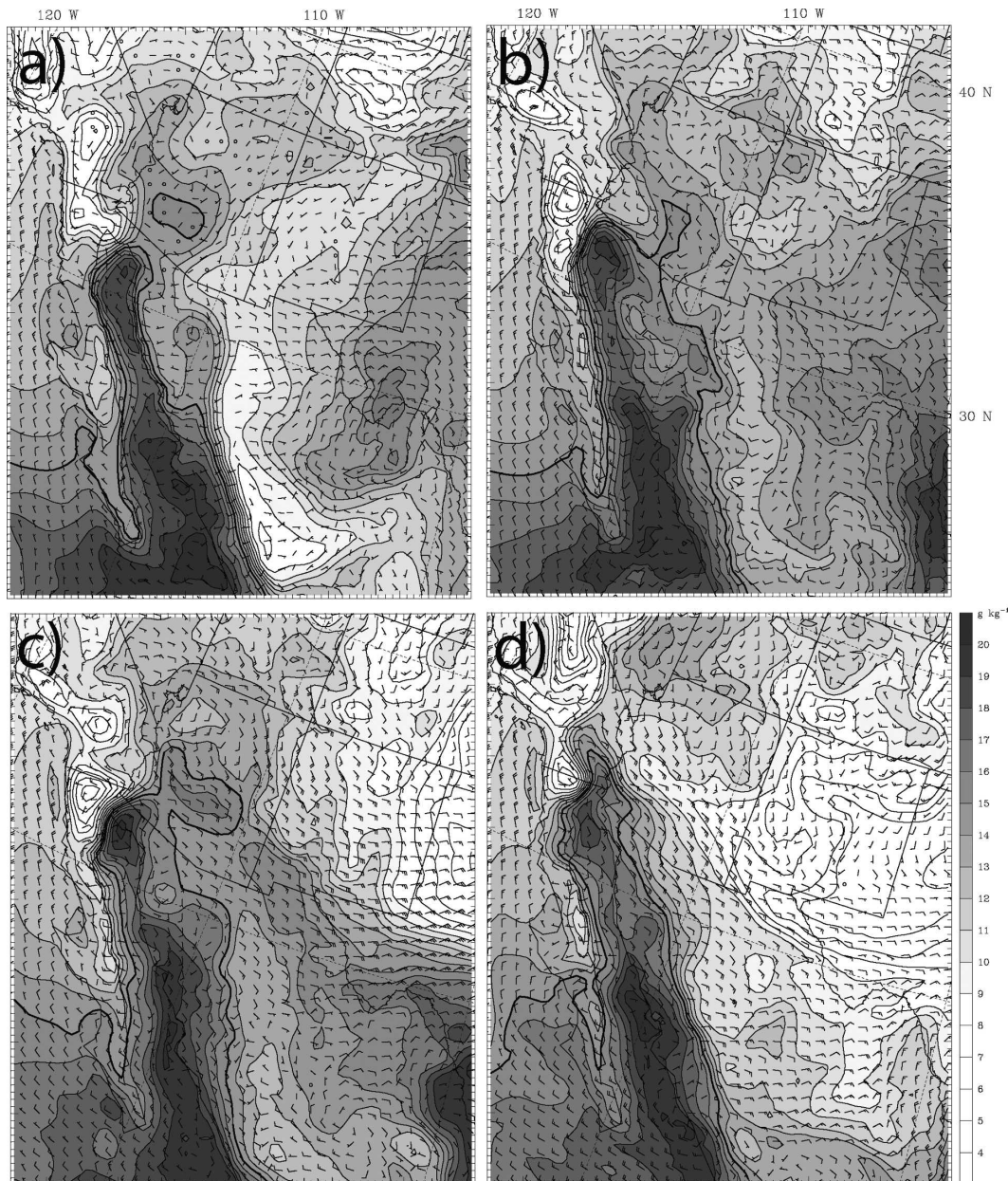


FIG. 11. Same as in Fig. 6, but for the August 1986 no-TEW run, showing (a) presurge (1800 UTC 26 August), (b) surge onset (0600 UTC 27 August), (c) during the surge (1200 UTC 28 August), and (d) postsurge (1200 UTC 29 August) conditions.

south-to-north zone of heavier rainfall stretching from Louisiana to Iowa. However, the comparison between simulated and observed rainfall is even less favorable in the no-TEW run, where the corridor of elevated precipitation amounts in the simulation is oriented more east-west across Kansas (Fig. 13c). The other three month-long control runs generate monthly total precipitation patterns that compare similarly to the CPC rainfall gauge analyses over the NAM and central

plains regions (not shown). In general, many of the observed larger-scale features of the rainfall analyses are captured reasonably well by the model simulations, if one allows for some shifting in location. However, some observed rainfall features admittedly are missed by the model simulations. Thus, while the model simulations reproduce many features that mimic the observed monthly rainfall distributions, the month-long simulations also have errors.

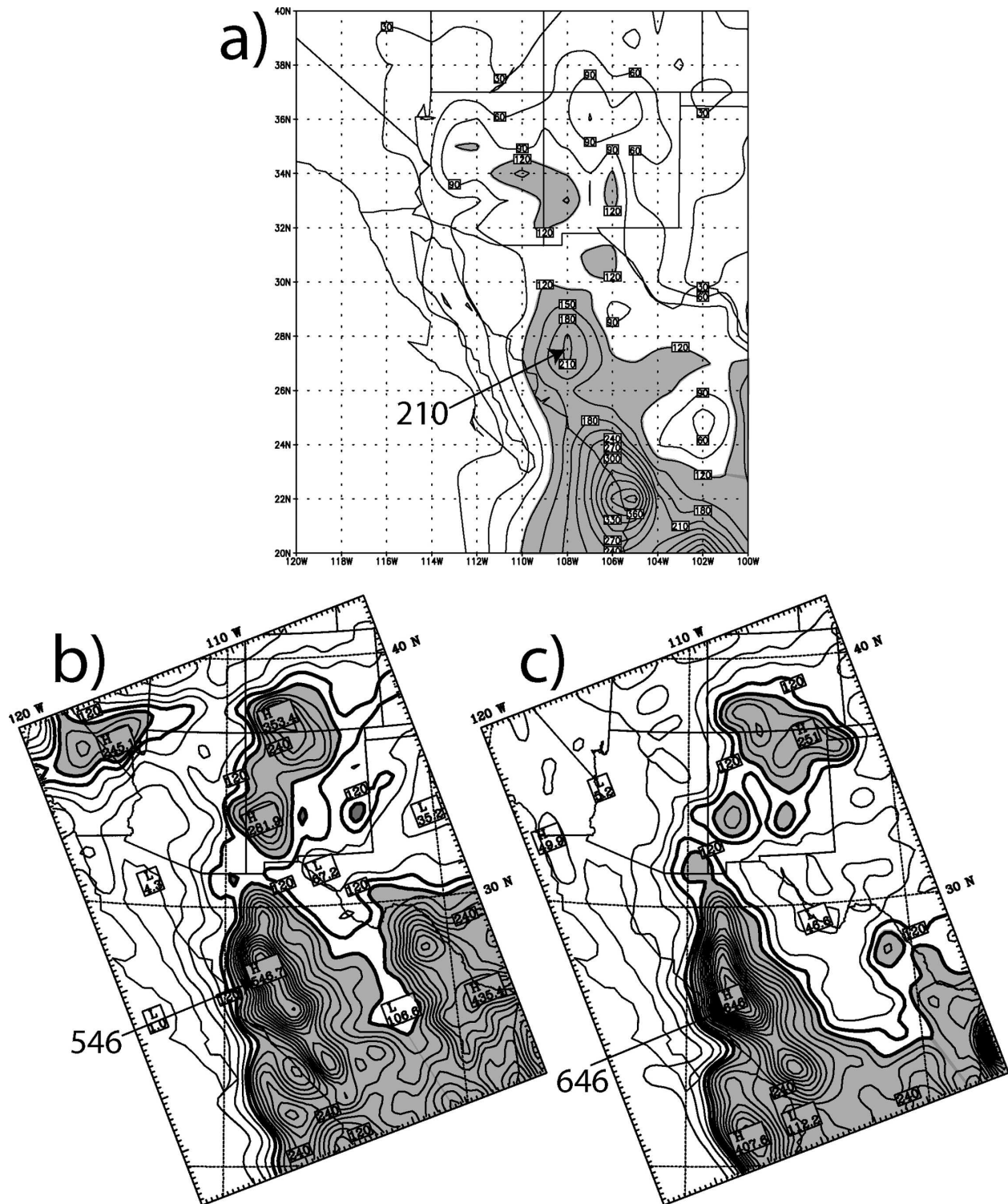


FIG. 12. Rainfall totals from August 1986 (mm) over the NAM region from (a) the CPC Unified Precipitation Data Set, (b) the control run, and (c) the no-TEW run. Isolines at 20-mm intervals. Values greater than 120 mm shaded in (a), while values greater than 140 mm shaded in (b) and (c). Bold line is 120 mm in (b) and (c) for comparison.

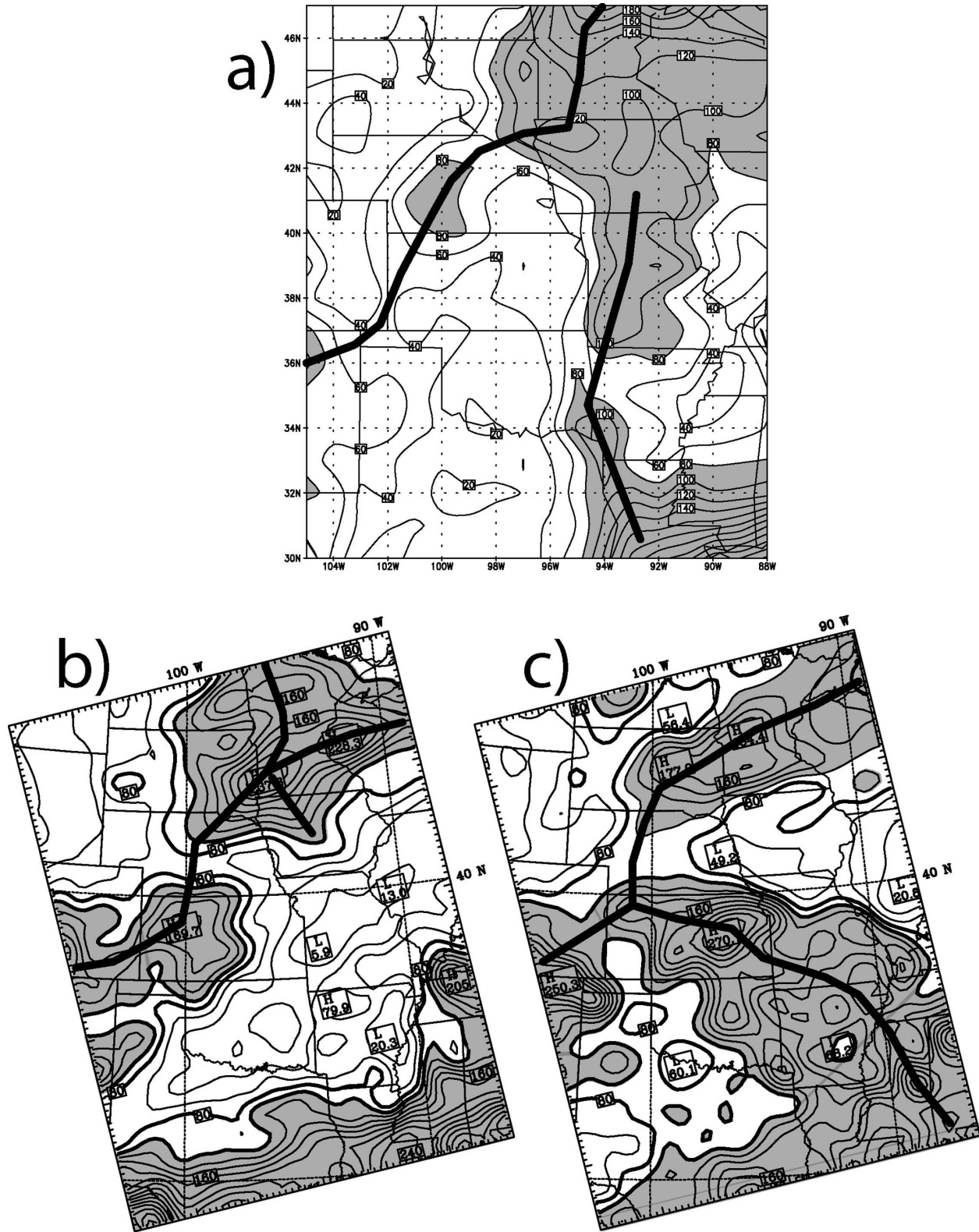


FIG. 13. Same as in Fig. 12, but over the central United States. Values greater than 80 mm shaded in (a), while values greater than 100 mm shaded in (b) and (c). Bold line is 80 mm in (b) and (c) for comparison. Thick bold lines indicate axes of higher rainfall totals.

TABLE 2. Maximum of monthly rainfall totals (mm) over western Mexico north of 25°N from the CPC analysis, the control run, and the no-TEW run.

	CPC analysis	Control (TEW)	no-TEW
August 1986	210	530	546
August 1988	210	546	646
July 1990	330	716	638
July 1992	140	477	633

Another approach to evaluate the ability of the model to simulate rainfall is to compare the maximum rainfall totals in the NAM region from both the model and the CPC analyses to determine if the model can reproduce the year-to-year variability in monthly rainfall amounts. The location of the maximum NAM rainfall is relatively consistent in both the model runs and observations, since it is tied to the sloping terrain of the SMO. Results indicate that the control run captures the variation in maximum rainfall more accurately than the no-TEW run, with July 1990 having the largest observed rainfall and July 1992 having the smallest rainfall (Table 2). In contrast, the no-TEW run has a reduced variation in maximum monthly rainfall totals and these smaller variations correlate poorly with the variations seen in the CPC precipitation data.

The ability of the control run to capture the year-to-year variability in monthly rainfall totals, combined with its ability to capture many (if not all) of the observed features in the monthly rainfall distribution over the NAM and central plains regions, yields some hope that we can explore the effects of TEWs on monthly rainfall totals in these two regions. While these results are only suggestive, owing to the admitted difficulties in simulating monthly precipitation amounts correctly, they are perhaps worthy of consideration.

Simulation rainfall differences

The absence of TEWs from the model simulations not only has an impact on NAM surge events, as shown earlier, but also on rainfall amounts over the NAM region. To capture better the model signal, the differences between the control and no-TEW rainfall totals are averaged for the four months. Results show that TEWs move rainfall out of the region along the SMO and over the Gulf of California (Fig. 14). In particular, rainfall is enhanced because of TEWs near the mouth of the gulf, where full gulf surges often start, and in far northwestern Mexico. The increase in rainfall over northwestern Mexico is consistent with the results of Douglas and Leal (2003) and Higgins et al. (2004), who show positive precipitation anomalies in this same re-

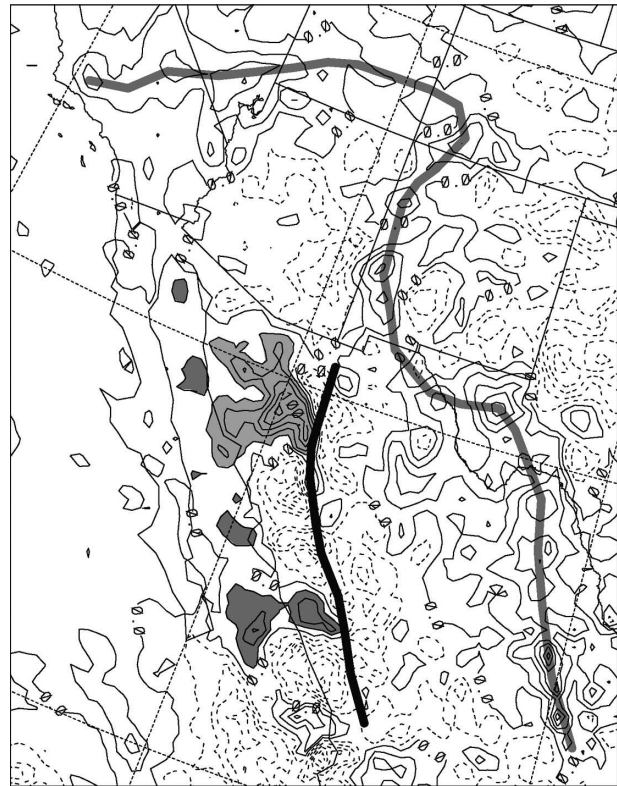


FIG. 14. Total monthly rainfall (mm) differences (control – no-TEW runs) averaged from all four months simulated over the NAM region. Isolines every 25 mm, with values greater than 25 mm shaded over the Gulf of California region. Solid lines are positive differences, and dashed lines are negative differences. Thick black line represents the axis of heaviest monthly rainfall in the control and no-TEW runs along the western slopes of the SMO. Thick gray line connects regions of positive rainfall differences along the northern and eastern peripheries of the NAM.

gion of northwestern Mexico in association with both strong and weak surges. The presence of TEWs also appears to produce increased precipitation along the northern and eastern periphery of the NAM region, stretching from California to southern Utah, eastern New Mexico, and on across southwest Texas and eastern Mexico (Fig. 14). Curiously, the average rainfall over the entire region shown in Fig. 14 differs by only 0.1% between the control and no-TEW runs. This suggests that while TEWs may help to expand the region of NAM rainfall, they have little influence on the total NAM-related rainfall.

The model simulations also indicate that the effects of TEWs on rainfall are not confined to northwestern Mexico and the southwestern United States. The Southern Plains experiences a substantial decrease in precipitation when TEWs are present (Fig. 15). The average monthly rainfall over the entire region in Fig. 15 decreases by 25.3 mm (23% of the average monthly

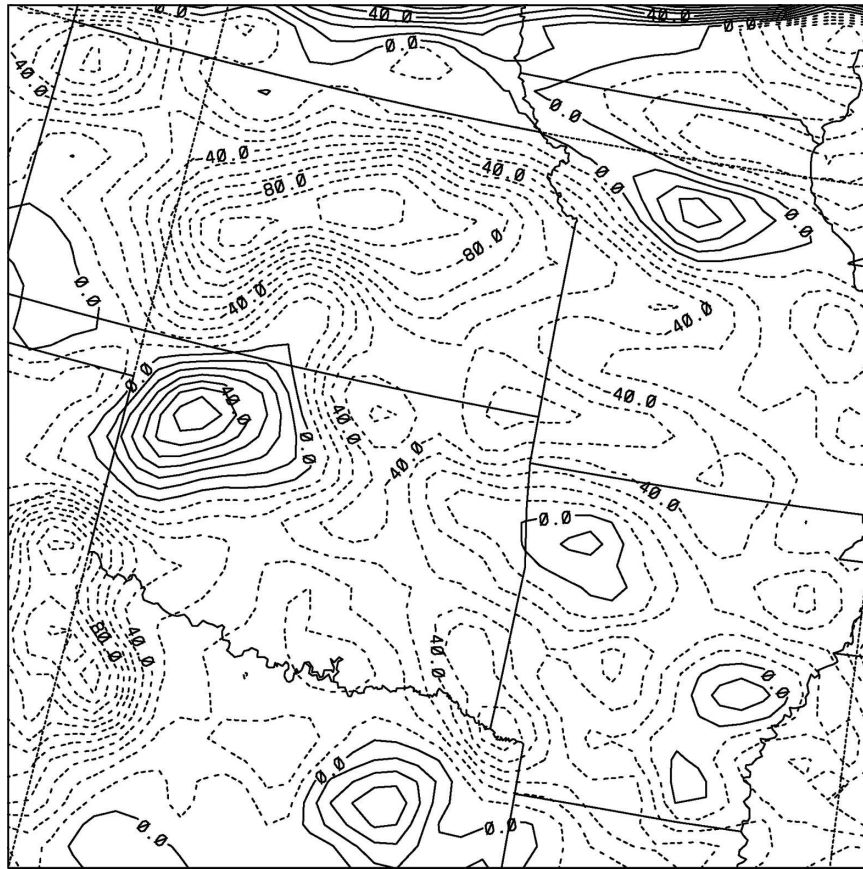


FIG. 15. Total monthly rainfall (mm) differences (control – no-TEW runs) averaged from all four months simulated over the Southern Great Plains region. Isolines every 10 mm. Solid lines are positive differences, and dashed lines are negative differences.

rainfall produced by the model in this region) when TEWs are present in the model boundary conditions. A similar change in the central U.S. precipitation distribution is seen in Higgins et al. (1997) in association with monsoon onset. They find that the precipitation anomaly over the Southern Plains states shifts from positive to negative several days before NAM onset in the southwestern United States, and the negative precipitation anomaly is very persistent after NAM onset. An analysis of precipitation anomalies associated with gulf surges is produced by Higgins et al. (2004), who show negative precipitation anomalies over the Southern Great Plains 2 to 4 days prior to gulf surge onset in Arizona. The only positive precipitation anomaly in the Southern Plains region is confined to the Texas coast 2 to 4 days after surge onset. The timing of the negative precipitation anomaly 2 to 4 days prior to surge onset in Arizona is consistent with the idea of a TEW influencing both the Southern Great Plains and the gulf surge in the NAM region in succession. Thus, the simulations indicate that TEWs may influence the amount of pre-

cipitation in the Southern Plains of the United States during the summertime, producing fairly broad regions with reduced rainfall from Kansas to Texas, in addition to their role in modifying the precipitation over the NAM region.

At least two possible reasons exist for the differences in central plains rainfall, both related to changes in meridional moisture flux (uq) over the region. It appears that the absence of TEWs generates periods of time where the northward moisture flux in the no-TEW run exceeds that in the control run. In general, TEWs alter the low- and midlevel wind trajectories over the Gulf of Mexico, at times significantly reducing the amount of moisture advected into the Southern Plains. This is consistent with the results of Mo and Berbery (2004), who show that decreases in Southern Great Plains rainfall are associated with the weakening of the low-level jet near the Texas coast and reduced northward moisture flux. However, moisture flux also can be altered in the control run through modification of the low-level water vapor. As the wave crosses the Gulf of Mexico, the

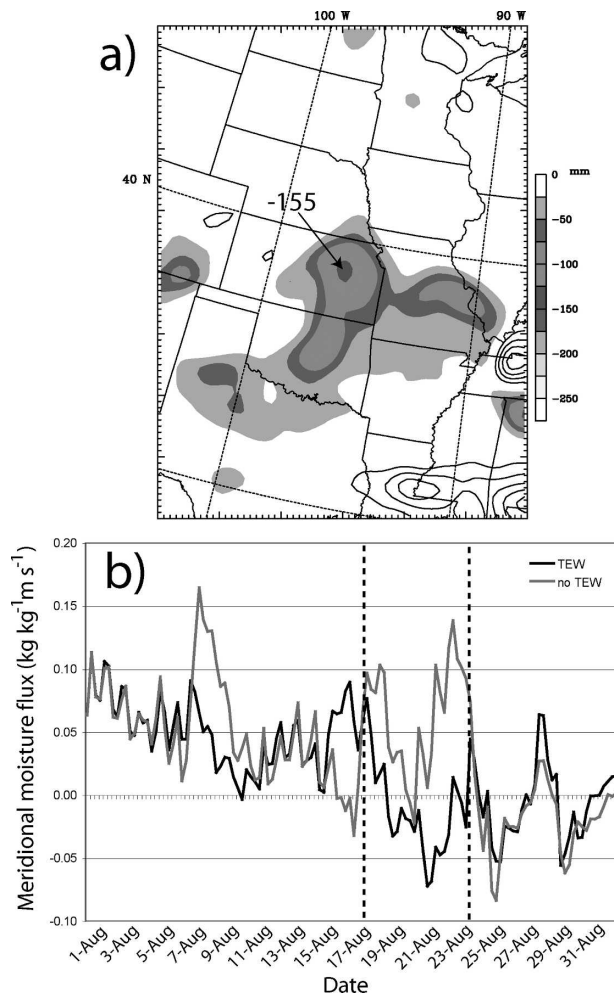


FIG. 16. Rainfall (mm) differences (control – no-TEW run) (a) over the central United States from 0000 UTC 17 Aug to 0600 UTC 23 Aug 1988, with (b) the meridional moisture flux at Dallas, TX, averaged from the surface to $\sigma = 0.675$. In (a), isolines (shading) indicate positive (negative) rainfall differences. Rainfall isolines are at 25-mm intervals, while the negative rainfall differences are shaded and their values are indicated by the key.

winds initially shift to northerly ahead of the wave and advect generally drier low-level air from over the land into the gulf region. Therefore, when the winds shift around to southerly on the other side of the wave, the air being advected into the Southern Plains from the Gulf of Mexico may not have as much moisture associated with it since its residence time over the warm gulf waters is short.

During a 6-day period in August 1988, rainfall in the no-TEW run increases in conjunction with increases in northward moisture flux at Dallas–Ft. Worth, Texas (Fig. 16). The introduction of more water vapor into the region allows values of convective available potential energy and precipitable water to increase, leading to an

enhanced potential for convection (not shown). However, increased rainfall in the no-TEW run does not always occur simultaneously with increased meridional moisture flux episodes. Sometimes the convective available potential energy is stored until a midlatitude disturbance provides the forcing for upward motion that helps to initiate deep convection. Less rainfall is produced in the control run since the values of convective available potential energy and precipitable water are less in this run compared to the no-TEW run.

6. Discussion

The mesoscale model MM5 is used to produce month-long simulations of the NAM during four separate months that have varying numbers of TEWs crossing western Mexico and different monthly precipitation distributions. All four months chosen for study are from the mature phase of the NAM (July and August). The simulations are able to reproduce a variety of surge events occurring over the Gulf of California, in good agreement with the available observations. These surges advect moisture into northwestern Mexico as well as the southwestern United States. Results further show that surge events fluctuate in strength, frequency, and location along the gulf from month to month, reaffirming that the NAM exhibits a great deal of interannual variability.

To gain an improved understanding of the relationship between TEWs and the NAM, a control run of the MM5 is compared to a simulation where TEWs have been removed from the model boundary conditions. A harmonic analysis is designed that successfully removes or dampens TEWs from the surface to the model top in the boundary conditions. The simulations produced using the boundary conditions without TEWs indicate that the TEWs are largely removed from the model, but that the model often creates disturbances over the Caribbean Sea and Mexico that still can influence the NAM.

The removal of TEWs from the model boundary conditions is shown to impact the gulf surges that develop over the NAM region. In most cases, gulf surges are still present when TEWs are removed; however, they are significantly weaker and their initiation is linked most often to westward-moving waves created within the model domain. The timing of the surges can also be altered by the absence of TEWs. These results indicate that gulf surges are strongly tied to TEWs during the four months studied, and although other disturbances can initiate gulf surges, the surges associated with TEWs appear to be stronger.

Simulations further show that not only are surges influenced by TEWs, but rainfall amounts over the

NAM region are influenced as well. In general, the presence of TEWs increases rainfall over the Gulf of California and northwestern Mexico and decreases rainfall along the slopes of the SMO within the region of heaviest average monsoon rainfall. In particular, rainfall is enhanced over the mouth of the Gulf of California where full gulf surges are often seen to begin. This result is consistent with the findings of Douglas and Leal (2003) and Higgins et al. (2004), who show that surges lead to increased rainfall in northwestern Mexico. In addition, rainfall is increased along the northern and eastern edges of the NAM region when TEWs are present in the model simulations. The model results, while only suggestive because of the challenges in simulating precipitation correctly, indicate that TEWs may act to expand the region of precipitation associated with the NAM with no apparent increase in total precipitation over the NAM region.

The presence of TEWs in the simulations also decreases rainfall over the U.S. Southern Plains region. Meridional moisture flux in the no-TEW run periodically exceeds that simulated when TEWs are present, thus possibly accounting for the increased rainfall over this region when waves are absent through changes in convective available potential energy and precipitable water. Results from Mo and Berbery (2004) support this conclusion, showing that decreases in Southern Great Plains rainfall are associated with the weakening of the low-level jet near the Texas coast and reduced northward moisture flux. In addition, Higgins et al. (2004) show negative precipitation anomalies over the Southern Great Plains 2 to 4 days prior to the onset dates of surges in Arizona, with positive precipitation anomalies 2 to 4 days after a surge being limited to the Texas coastal region. These anomalies are consistent with the effects of TEWs on the northward moisture flux.

These results suggest that TEWs play an important role in determining the intensity and northern extent of NAM, likely due to their relationship with gulf surges, as well as the U.S. Southern Plains rainfall through their ability to disrupt and reduce the northward meridional moisture flux. While the results regarding rainfall are only suggestive, it is clear that TEWs deserve greater attention since their predictability may be important to improving seasonal forecasts over North America during the summer. Studies examining the relationship between TEW frequency and both NAM and central U.S. rainfall over larger time periods are needed.

Acknowledgments. This study was supported by the NOAA/Office of Global Programs (now the Climate

Program Office) Joint PACS–GAPP North American Warm Season Precipitation Initiative. Funding to CIMMS was provided by the NOAA/Office of Oceanic and Atmospheric Research under NOAA–University of Oklahoma Cooperative Agreement Number NA17RJ1227, U.S. Department of Commerce. The authors greatly appreciate the constructive and helpful reviews provided by two anonymous reviewers. We also want to thank Drs. Alan Shapiro and Fred Carr of the School of Meteorology for their assistance during this research project. Finally, we thank Wayne Higgins for his assistance in using the CPC precipitation data and NCAR for providing the NCEP–NCAR reanalysis data.

REFERENCES

- Anderson, B. T., J. O. Roads, S. C. Chen, and H. M. H. Juang, 2000: Regional simulation of the low level monsoon winds over the Gulf of California and southwestern United States. *J. Geophys. Res.*, **105**, 17 955–17 969.
- Anthes, R. A., Y.-H. Kuo, D. Baumhefner, R. M. Errico, and T. W. Bettge, 1985: Predictability of mesoscale atmospheric motions. *Advances in Geophysics*, Vol. 28B, Academic Press, 159–202.
- Berbery, E. H., 2001: Mesoscale moisture analysis of the North American monsoon. *J. Climate*, **14**, 121–137.
- , and M. S. Fox-Rabinovitz, 2003: Multiscale diagnosis of the North American monsoon system using a variable-resolution GCM. *J. Climate*, **16**, 1929–1947.
- Brenner, I. S., 1974: A surge of maritime tropical air—Gulf of California to the southwestern United States. *Mon. Wea. Rev.*, **102**, 375–389.
- Burpee, R. W., 1972: The origin and structure of easterly waves in the lower troposphere of North Africa. *J. Atmos. Sci.*, **29**, 77–90.
- , 1974: Characteristics of North African easterly waves during the summers of 1968 and 1969. *J. Atmos. Sci.*, **31**, 1556–1570.
- Davies, H. C., and R. E. Turner, 1977: Updating prediction models by dynamical relaxation: An examination of the technique. *Quart. J. Roy. Meteor. Soc.*, **103**, 225–245.
- Douglas, M. W., 1995: The summertime low-level jet over the Gulf of California. *Mon. Wea. Rev.*, **123**, 2334–2347.
- , and J. C. Leal, 2003: Summertime surges over the Gulf of California: Aspects of their climatology, mean structure, and evolution from radiosonde, NCEP reanalysis, and rainfall data. *Wea. Forecasting*, **18**, 55–74.
- , R. A. Maddox, K. Howard, and S. Reyes, 1993: The Mexican monsoon. *J. Climate*, **6**, 1665–1678.
- , A. Valdez-Manzanilla, and R. G. Cueto, 1998: Diurnal variation and horizontal extent of the low-level jet over the northern Gulf of California. *Mon. Wea. Rev.*, **126**, 2017–2025.
- Dudhia, J., 1989: Numerical study of convection observed during the Winter Monsoon Experiment using a mesoscale two-dimensional model. *J. Atmos. Sci.*, **46**, 3077–3107.
- , 1993: A nonhydrostatic version of the Penn State–NCAR Mesoscale Model: Validation tests and simulation of an Atlantic cyclone and cold front. *Mon. Wea. Rev.*, **121**, 1493–1513.
- Fuller, R. D., and D. J. Stensrud, 2000: The relationship between

- tropical easterly waves and surges over the Gulf of California during the North American Monsoon. *Mon. Wea. Rev.*, **128**, 2983–2989.
- Gochis, D. J., W. J. Shuttleworth, and Z.-L. Yang, 2002: Sensitivity of the modeled North American monsoon regional climate to convective parameterization. *Mon. Wea. Rev.*, **130**, 1282–1298.
- Gutzler, D. S., and Coauthors, 2005: The North American Monsoon Model Assessment Project: Integrating numerical modeling into a field-based process study. *Bull. Amer. Meteor. Soc.*, **86**, 1423–1429.
- Hales, J. E., 1972: Surges of maritime tropical air northward over the Gulf of California. *Mon. Wea. Rev.*, **100**, 298–306.
- , 1974: Southwestern United States summer monsoon source—Gulf of Mexico or Pacific Ocean. *J. Appl. Meteor.*, **13**, 331–342.
- Higgins, R. W., and W. Shi, 2000: Dominant factors responsible for interannual variability of the summer monsoon in the southwestern United States. *J. Climate*, **13**, 759–776.
- , Y. Yao, and X. L. Wang, 1997: Influence of the North American monsoon system on the U.S. summer precipitation regime. *J. Climate*, **10**, 2600–2622.
- , K. C. Mo, and Y. Yao, 1998: Interannual variability of the U.S. summer precipitation regime with emphasis on the southwestern monsoon. *J. Climate*, **11**, 2582–2606.
- , W. Shi, and C. Hain, 2004: Relationships between Gulf of California moisture surges and precipitation in the southwestern United States. *J. Climate*, **17**, 2983–2997.
- Hong, S.-Y., and H.-L. Pan, 1996: Nonlocal boundary layer vertical diffusion in a medium-range forecast model. *Mon. Wea. Rev.*, **124**, 2322–2339.
- Kain, J. S., and J. M. Fritsch, 1990: A one-dimensional entraining/detraining plume model and its application in convective parameterization. *J. Atmos. Sci.*, **47**, 2784–2802.
- Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.*, **77**, 437–471.
- McCollum, D. M., R. A. Maddox, and K. W. Howard, 1995: Case study of a severe mesoscale convective system in central Arizona. *Wea. Forecasting*, **10**, 643–668.
- Mlawer, E. J., S. J. Taubman, P. D. Iacono, and S. A. Clough, 1997: Radiative transfer for inhomogeneous atmosphere: RRTM, a validated correlated-k model for the longwave. *J. Geophys. Res.*, **102**, 16 663–16 682.
- Mo, K. C., and E. H. Berbery, 2004: Low-level jets and the summer precipitation regimes over North America. *J. Geophys. Res.*, **109**, D06117, doi:10.1029/2003JD004106.
- , J.-K. Schemm, H. M. H. Juang, R. W. Higgins, and Y. Song, 2005: Impact of model resolution on the prediction of summer precipitation over the United States and Mexico. *J. Climate*, **18**, 3910–3927.
- Mock, C. J., 1996: Climatic controls and spatial variations of precipitation in the western United States. *J. Climate*, **9**, 1111–1125.
- NOAA, 1988: *Storm Data*. Vol. 30, No. 8, 66 pp. [Available from National Climatic Data Center, Environmental Data and Information Service, NOAA, Asheville, NC 28801.]
- Paegle, J., Q. Yang, and M. Wang, 1997: Predictability in limited area and global models. *Meteor. Atmos. Phys.*, **63**, 53–69.
- Reed, R. J., D. C. Norquist, and E. E. Recker, 1977: The structure and properties of African wave disturbances as observed during phase III of GATE. *Mon. Wea. Rev.*, **105**, 317–333.
- Saleeby, S. M., and W. R. Cotton, 2004: Simulations of the North American monsoon system. Part I: Model analysis of the 1993 monsoon season. *J. Climate*, **17**, 1997–2018.
- Schmitz, J. T., and S. L. Mullen, 1996: Water vapor transport associated with the summertime North American monsoon as depicted by ECMWF analyses. *J. Climate*, **9**, 1621–1634.
- Stensrud, D. J., R. L. Gall, S. L. Mullen, and K. W. Howard, 1995: Model climatology of the Mexican monsoon. *J. Climate*, **8**, 1775–1794.
- , —, and M. K. Nordquist, 1997: Surges over the Gulf of California during the Mexican monsoon. *Mon. Wea. Rev.*, **125**, 417–437.
- Tang, M., and E. R. Reiter, 1984: Plateau monsoons of the Northern Hemisphere: A comparison between North America and Tibet. *Mon. Wea. Rev.*, **112**, 617–637.
- Tribbia, J. J., and D. P. Baumhefner, 2004: Scale interactions and atmospheric predictability: An updated perspective. *Mon. Wea. Rev.*, **132**, 703–713.
- Wilks, D. S., 1995: *Statistical Methods in the Atmospheric Sciences: An Introduction*. International Geophysics Series, Vol. 59, Academic Press, 467 pp.