Numerical Simulation of Heavy Precipitation in Northern Baja California and Southern California

RUTH CEREZO-MOTA AND TEREZA CAVAZOS
Departamento de Oceanografía Física, CICESE-Ensenada, Ensenada, Baja California, Mexico

LUIS M. FARFÁN
CICESE-La Paz, La Paz, Baja California Sur, Mexico

(Manuscript received 20 January 2005, in final form 21 June 2005)

ABSTRACT

The fifth-generation Pennsylvania State University–NCAR Mesoscale Model (MM5) was used to simulate the heavy-precipitation events of 6–21 January 1993 during a moderate El Niño that produced severe flooding, landslides, and the loss of many lives near the border of California and Baja California, Mexico. The mean synoptic-scale condition consisted of a strong upper-level subtropical westerly jet over the U.S.–Mexico western border and the passage of three surface fronts, along with intense low-level advection of tropical moisture and convective activity over the region. The MM5 reproduced the mean observed circulation patterns of the study period but with less intensity. During the first extreme event on 6 January, the model sounding profile in San Diego, California, was more unstable and saturated than observed, the horizontal winds were weaker in the lower troposphere, and the simulated precipitation was slightly underestimated. The model precipitation was verified with hourly and daily precipitation records in California and Baja California, respectively, and regional errors were obtained for the 16-day period. The simulation showed an improvement when the resolution increased from 90 to 30 km, but there was not a significant improvement from 30 to 10 km; however, extreme rainfall events (>10 mm day⁻¹) were better resolved by the 10-km grid, possibly due to the finer-scale topography.

1. Introduction

Winter rainfall in northern Baja California, Mexico, and southern California significantly contributes to the annual precipitation (e.g., Minnich et al. 2000; Gershunov and Cayan 2003); in the Tijuana (Mexico)–San Diego (California) area winter rainfall supplies 90% of the annual precipitation (Cavazos and Rivas 2004). This includes certain periods of heavy precipitation that are partially associated with interannual fluctuations due to El Niño–Southern Oscillation (ENSO) anomalies in conjunction with North Pacific decadal and interannual variability (e.g., Gershunov and Barnett 1998; Pavía and Badán 1998; Gershunov and Cayan 2003; Cavazos and Rivas 2004). Atmospheric and oceanic teleconnections during strong and moderate El Niño events produce a southward displacement of the upper-level subtropical westerly jet stream over California and Baja California as well as onshore moisture advection that favors heavy rainfall events over the region (e.g., Minnich et al. 2000; Cavazos and Rivas 2004). Moreover, the transverse mountain ranges in southern California (e.g., San Bernardino; Fig. 1) enhance rainfall very efficiently when moisture is brought in by southerly winds, as it is common during some El Niño years (e.g., Kim 1997; Panday et al. 1999; Ralph et al. 2003). Heavy precipitation also occurs during neutral (non-ENSO) years and has been related to synoptic-scale conditions such as the Pineapple Express (Mo and Higgins 1998; Cavazos and Rivas 2004) and intraseasonal phenomena such as the Madden–Julian oscillation (MJO; Mo and Higgins 1998; Jones 2000). The interaction of these synoptic systems with the complex terrain of the region also plays an important role in the generation of heavy precipitation (e.g., Kim 1997; Leung et al. 2003; Ralph et al. 2003; Neiman et al. 2004).
The winters of 1992 and 1993 marked one of the longest periods of continuous warm El Niño conditions on record (Bell and Basist 1994). A prominent feature during the winter of 1993 was extremely heavy precipitation totals over California and the southwestern United States (Bell and Basist 1994). Long-term precipitation data from San Diego indicate that January 1993 is the wettest month in the last 50 yr (9.09 in., or 227 mm), and 1993 is the fourth wettest year since 1850. Similarly, the northern portion of the state of Baja California, Mexico, recorded several extreme precipitation events (>10 mm day⁻¹; Cavazos and Rivas 2004) in January 1993; in Tijuana, a border city with San Diego, the observed rainfall in January 1993 was 210 mm, just 23 mm below the total annual precipitation (Cavazos and Rivas 2004). During January 1993, flooding and landslides in the Tijuana area were associated with 40 casualties and great property damages (Bocco et al. 1993). Recently, in October and December 2004 and in January and February 2005, intense landfalling storms associated with the passage of several frontal systems, tropical moisture, and a southward displaced subtropical westerly jet produced flash floods, landslides, property damages, and fatalities from Santa Barbara to Tijuana.

To mitigate the impact of heavy precipitation it is necessary to improve the knowledge of atmospheric and oceanic conditions as well as to offer realistic regional forecasts several days in advance. Prediction with regional models can provide very useful real-time mesoscale weather features at fine temporal and spatial resolutions. The fifth-generation Pennsylvania State University–National Center of Atmospheric Research (NCAR) Mesoscale Model (MM5; Grell et al. 1995) has been successfully used in operational forecasts for the Pacific Northwest (e.g., Mass et al. 2003, and references therein). The nonhydrostatic MM5 resolves the primitive equations in sigma coordinates and includes several cumulus parameterization schemes (CPS) and explicit moisture schemes (EMS), which play an important role in the adequate simulation of the western coast precipitation (e.g., Colle and Mass 2000; Jones 2000; Mass et al. 2003). The ability of these parameterizations seems to be related to local convective processes, season of the year, grid size, and interaction with the topography. For instance, in a simulation over the Pacific Northwest mountains, Colle et al. (1999) found that the Reisner et al. (1998) EMS underestimated leeward precipitation and overestimated the precipitation along the windward slopes, but total precipitation improved when spatial resolution increased from 36 to 12 km.

The objective of this study was to simulate the heavy precipitation observed near the U.S.–Mexico western border during 5–21 January 1993 using the MM5. We utilized three spatial grids (90, 30, and 10 km) and tested for possible improvements in the simulation of daily precipitation when the resolution increased from 30 to 10 km. The model synoptic-scale circulation was compared with that obtained from the National Centers for Environmental Prediction (NCEP)–NCAR reanalysis dataset (Kalnay et al. 1996). To verify the thermodynamic and boundary layer conditions, we compared the observed sounding in San Diego with the nearest gridpoint sounding of the 10-km domain during the first heavy rainfall event of the period on 6 January. We validated the model gridpoint precipitation during the entire period using observed hourly precipitation data from stations in California and daily data from Baja California and calculated root-mean-square errors (rmse) and the model bias as described in section 2. Section 3 gives a brief description of the model configuration. Results of the simulations and model errors are discussed in section 4, and the conclusions derived from this study are presented in section 5.

2. Data and methods

The MM5 was initialized with the NCEP–NCAR reanalysis dataset, which has a spatial resolution of 2.5° latitude × 2.5° longitude (Kalnay et al. 1996). Upper-air soundings from San Diego (NKK; Fig. 1) were used...
to validate the model vertical structure of the atmosphere during the first heavy-precipitation event of the period (6 January). Radiosonde data were obtained from the National Oceanic and Atmospheric Administration’s (NOAA) Forecast Systems Laboratory (FSL) database.

Long-term daily precipitation for San Diego was obtained from the Western Regional Climate Center (WRCC) in Reno, Nevada. To evaluate the MM5 accuracy to simulate extreme precipitation we used hourly precipitation data from observation sites in the state of California from the National Climate Center cooperative (COOP) network. From the 250 sites available we used only 51 that had enough information for the study period. The COOP rain gauges have some systematic errors (NCDC 2003), which may lead to underestimation of precipitation by 5%–15% due to wind effects and evaporation with major problems in the mountainous terrain where frozen precipitation is common (e.g., Colle et al. 1999). Daily precipitation for eight Mexican stations located in Baja California was obtained from ERIC-II, a dataset provided by the Mexican Institute of Water Technology and the Servicio Meteorológico Nacional (SMN, in Mexico). Figure 1 shows the location of all the observation sites used in this analysis.

To verify the model precipitation with the observations, the model grid precipitation was first interpolated to each observation site using the Cressman inverse distance method (Cressman 1959; Colle et al. 1999). Then, for the 16 days ($n_{\text{obs}}$) of the study period (5–21 January 1993) rmse were calculated between daily model precipitation ($\text{ppt}_D$) in a particular domain and daily-observed precipitation ($\text{ppt}_{\text{obs}}$) in all the stations in Fig. 1 using

$$\text{rmse} = \sqrt{\frac{\sum_{n=1}^{n_{\text{obs}}} (\text{ppt}_D - \text{ppt}_{\text{obs}})^2}{n_{\text{obs}}}}. \quad (1)$$

We also estimated the total bias ($B$) score of the model precipitation interpolated to all the observation sites and for several precipitation thresholds: light precipitation (0–2.5 mm), moderate precipitation (2.5–10 mm), and heavy precipitation (>10 mm). The $B$ score is defined as

$$B = \frac{\sum_{n=1}^{n_{\text{obs}}} \text{ppt}_D}{n_{\text{obs}}} \cdot \frac{n_{\text{obs}}}{\text{ppt}_{\text{obs}}}. \quad (2)$$

A $B$ score $<1$ indicates that the model underestimated the observed precipitation, $B > 1$ indicates overestimation, and $B = 1$ indicates a perfect simulation.

3. Model description

The MM5 was run from 5 to 21 January 1993. Initial tests suggested that the model needed a spinup time of about 10 h; thus, the simulations were initialized every 72 h with an overlapping of 12 h between runs. These relatively long runs were chosen to test the skill of the model to predict extreme precipitation a few days in advance in order to provide longer lead times for flood forecasting, which in turn could eventually result in more advanced warnings of potential flooding to the general public. The model was operated with three numerical domains of 90-, 30-, and 10-km spatial resolutions, with the finest domain, D3, centered in the San Diego–Tijuana border area (Fig. 2). The two nested grids (D2 and D3) extend from California to the Baja California peninsula, as shown in Fig. 2, and used a two-way feedback. That is, each of the nested domains obtains their lateral boundary conditions from their outer domain during the integration and feeds the results back from the finest mesh to the coarser domain. Thus, the coarser domain (D2) is also influenced by the finer domain (D3), which may provide an upper bound in the ability of D2 to predict small-scale precipitation.

The mother domain (D1) extends from 0° to 60°N and from 170° to 100°W to include the possible teleconnections that usually occur between the equator and the North Pacific during El Niño events. Vertically, the
equations are solved on 23 sigma levels with finer resolution in the planetary boundary layer (PBL) using the Hong and Pan (1996) parameterization scheme. The top of the model atmosphere is set at 100 mb and an upper radiative condition was used to prevent reflection of energy from the model top in form of gravity waves (Klemp and Durran 1983).

From eight available CPS and eight EMS, and based on the results of the Pacific Northwest studies (e.g., Colle and Mass 2000; Colle et al. 2000), seven combinations of these schemes were tested. From these sensitivity analyses we found that for this region and season the Kain–Fritsch-2 CPS (Kain 2004) and the Schultz EMS (Schultz 1995) resolved precipitation better over the 30- and 10-km domains than other combinations available in the MM5. Since in mountainous regions an important part of the precipitation may occur in the form of snow, probably an EMS that considers ice and graupel processes such as the Schultz scheme could better resolve the microphysical processes involved in the formation of solid precipitation. On the other hand, the Kain–Fritsch-2 scheme includes shallow and deep convection. Shallow convection is typical of the west coast winter precipitation due to the subsidence influence of the North Pacific high and stabilization of the cold California Current, but deep convection may occur during extreme events (e.g., Neiman et al. 2004). The following results are based on the Kain–Fritsch-2 CPS and the Schultz EMS parameterizations, and the precipitation errors are presented only for the finest domain (D3), since little difference was observed between the errors in D2 and D3.

4. Results

a. Mean synoptic conditions

Weak sea surface temperature (SST) anomalies characterized the moderate El Niño 1992/93 episode (Bell and Basist 1994). As illustrated in Fig. 3a, in January 1993 positive SST anomalies (0.5°C) were observed in a band from the date line at the equator to the west coast of the Californias with maximum anomalies up to 1°C at the southern end of the California Current. As a partial response to this anomalous warming, the subtropical westerly jet was located 7° south of its long-term mean January position (30°–35°N), as demonstrated by the mean 200-mb zonal wind component in Fig. 3b. Maximum zonal wind anomalies at this level (>12 m s⁻¹; not shown) were noticed near the western coast of the U.S.–Mexico border; similar conditions have been documented for other heavy-precipitation events in the same region (e.g., Cavazos and Rivas 2004). At the 850-mb level, a persistent anomalous trough over the study area (not shown) and a strong meridional moisture flux from the tropical Pacific (20°N, 140°W) to the U.S.–Mexico border (Fig. 3c) favored intense convective activity, as indicated by the large negative anomalies of outgoing longwave radiation (OLR) up to −20 W m⁻² over the area of study (Fig. 3d). This OLR anomaly pattern was very persistent throughout December–February 1992/93 and was partially associated with intraseasonal oscillations from the Indian Ocean to the date line at the equator (Bell and Basist 1994).

The MM5 reproduces the mean atmospheric circulation patterns very well during 5–21 January 1993. Similar to the observations, the MM5 shows an intense low-level meridional moisture advection from the tropical Pacific (20°N, 140°W) to the coasts of California and Baja California (Fig. 4a), and a strong subtropical jet stream over the U.S.–Mexico border region (Fig. 4b), which favored the heavy-precipitation events of this period. Although the mean spatial structure of the model circulation during the period is very similar to that depicted by the NCEP–NCAR reanalysis in Fig. 3c, the simulated moisture advection into the study area is almost 50% weaker than observed and the model zonal wind speeds in the subtropical westerly jet are 10% weaker than observed.

b. January 1993 precipitation over the U.S.–Mexico border

Figure 5 shows the mean daily observed and simulated precipitation at local time (LT) for the stations shown in Fig. 2 for the 16 days of the analysis, as well as the rms errors for domains D2 and D3. The results of this figure are shown in LT because hourly precipitation was not available in the stations of Baja California. As described above, low-level southwesterly flow advected tropical moisture into the region and by 6–7 January all of the stations in Fig. 5 achieved the first heavy-precipitation event of the period, when a frontal system was observed off the west coasts of California and Baja California (NOAA 1993). A total of four (one stationary and three cold) fronts were observed during the period, and three of them coincided with the heavy-precipitation events in San Diego and Tijuana. Tijuana registered 85 mm on 6 January while San Diego documented 40 mm (Fig. 5). This large difference in the observed precipitation, which the model correctly simulates, could be due to the distance between the two locations (40 km; see Figs. 1 and 2) and the direction of the winds in the two sites in relation to the warm part of the frontal system (Fig. 6). At 1200 UTC San Diego experienced light southeasterly winds (Fig. 6); unfortunately there is not wind information available for
Tijuana, but the isobars along the coast suggest that the impact of the warm front may have been observed first in Tijuana and then in San Diego. The passage of the front and the instability that dominated the region (Fig. 7) on 6 January were partially responsible for the observed heavy coastal precipitation. However, the model overestimated the vertical velocity (dp/dt) over the study area on 6 January coordinated universal time (UTC; Fig. 7b), which may explain the early onset of the model precipitation, as indicated by the model heavy rainfall on 5 January local time in San Diego and Tijuana (Fig. 5).

The winds tend to loose moisture when passing through the mountains of California and Baja California; this is reflected in the lower amount of precipitation observed on 7 January in the two continental stations (El Centro and Mexicali) located in the Sonoran desert region, but the model was unable to capture on time any of the rainfall registered during the period in these two stations (Fig. 5). The fact that in many cases the model triggered the onset of precipitation with a lag of few hours compared to the observations could be partially attributed to the radiation, land surface, and/or PBL schemes selected. For example, the Hong and Pan (1996) PBL scheme may have problems with the timing and placement of synoptic-scale features, which affect surface-layer winds, temperature and moisture gradients, and precipitation. As will be discussed later, and also as described by other authors (e.g., Kim 1997; Panday et al. 1999), the onset of precipitation in this region shows a strong correlation between the wind direction and vertical shear. Therefore, if the model does not adequately capture any of these features, the resulting precipitation will be mistaken. Future tests with different PBL, radiation, and land surface parameterization schemes are needed to determine the physical sources of the mistiming of the precipitation onset.

An inactive period with no rainfall followed the first heavy event, but the model overestimated the rainfall of that period (8–11 January). The next active period

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**Fig. 3.** Mean observed conditions from 5 to 21 Jan 1993. (a) SST anomaly (°C); positive anomalies are shaded. (b) Mean 200-mb zonal wind component (m s⁻¹); wind speeds greater than 40 m s⁻¹ are shaded. (c) Mean meridional moisture flux at 850 mb (g kg⁻¹ m s⁻¹); positive values are shaded. (d) Surface OLR anomaly (W m⁻²); negative anomalies are shaded.
on 12–13 January was characterized by the passage of a cold front and heavy rainfall, which the model captured very well in San Diego and Tijuana. However, it missed the last heavy event of January (15–16) during the passage of another cold front. Both numerical domains (D2 and D3) produced similar rms errors along the U.S.–Mexico border region, as indicated in Fig. 5, suggesting that an increase in the model spatial resolution beyond 30 km may not necessarily improve the model precipitation.

c. 6 January 1993 extreme event

Hourly precipitation during the first heavy event on 6 January (Fig. 8) in San Diego and in the nearest model

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**Fig. 4.** MM5-simulated conditions in D1 (90-km resolution) from 5 to 21 Jan 1993. (a) Mean meridional moisture flux at 850 mb (g kg$^{-1}$ m s$^{-1}$); positive values are shaded. (b) Mean 200-mb zonal wind component (m s$^{-1}$); wind speeds greater than 40 m s$^{-1}$ are shaded.

**Fig. 5.** Daily observed precipitation (ppt obs, solid line with marker) in January 1993 in selected stations in the U.S.–Mexico border region (Fig. 2) and simulated precipitation (broken line) in the nearest grid point of D3 (10 km) at local time. Total observed precipitation in January, total simulated precipitation in D2 and D3, and rms errors (mm) in D2 and D3 are indicated. The first extreme event on 6 Jan is highlighted.
grid point of D3 indicates that the model triggered the onset of precipitation 8 h earlier than observed (on 5 January local time; Fig. 5) and that the model was unable to capture the largest observed peak at 2100 UTC (Fig. 8). An examination of the observed sounding at 0000 UTC of 6 January 1993 in San Diego (NKX) (Fig. 9a) reveals a dry and stable column, from the surface to the middle troposphere, with weak southerly winds below 700 mb, consistent with the lack of observed rainfall in the early hours of that day (Fig. 8). In contrast, the simulated sounding (Fig. 9b) shows a saturated and unstable profile as the lifted index (LI) is 1 and K index is 34, with south-southwesterly winds much stronger than observed along with a strong vertical wind shear in the first 2 km, explaining the early onset of the model precipitation at 0200 UTC.

Twelve hours later (1200 UTC) the conditions in San Diego changed dramatically as indicated by the observed sounding (Fig. 10a), which shows a nearly saturated profile from low to midlevels with 2.96 cm of precipitable water (PW) and strong south-southwesterly flow. The MM5 sounding at 1200 UTC (Fig. 10b) also depicts a marginally unstable column with little buoyancy (LI close to 0) and a moist environment (PW = 2.62 cm) that favored light precipitation as indicated by the model precipitation (Fig. 8). Low-level (1000 to 700 mb) southwesterly winds are weaker than observed and also weaker than the model winds at 0000 UTC. Bell and Basist (1994) document that intensity and direction of the wind are highly correlated with rainfall in this region; anomalous southwesterly flow is associated with above-normal precipitation. This may explain why the model slightly underestimated precipitation at this hour. At the end of the day, the model was unable to capture the observed maximum at 2100 UTC, but reproduced well the observed precipitation at 2200–2300 UTC (Fig. 8).

d. Regional precipitation and errors

As shown in Table 1, the model tended to overestimate light and moderate rainfall (<10 mm day⁻¹) and to underestimate extreme rainfall events (>10 mm day⁻¹). Although the bias score of the two domains is large in the first two intervals, the finest domain (D3) only underestimated the occurrence of extreme events by 30% (B = 0.7 in Table 1) as opposed to 45% by the 30-km domain.

Figure 9a shows the mean observed precipitation in southern California and Baja California, Mexico, during the period of analysis. For verification, southern California has a larger number of stations than Baja California, as indicated by the dots in Fig. 11a, but in spite of that all the southwestern portion of California lacks of information for this particular period. The
mean daily-observed precipitation maximum occurred over the San Bernardino Mountains (34°N, 117°W) in California, an indication of frontal and orographic precipitation. Consistently, the model grid D3 also produced a maximum along the mountains of California and Baja California from San Bernardino to Sierra de Juarez (Fig. 11b). Precipitation over topography is very sensitive to the direction of prevailing winds. The transverse mountains ranges in California (e.g., San Bernardino) enhance rainfall very efficiently when moisture is brought in by southerly winds (Kim 1997; Panday et al. 1999; Ralph et al. 2003). Model and observations show a relative agreement along the west coast with mean precipitation in the 10–30-mm range, while the model underestimates precipitation in the leeward side of Sierra de Juarez in Baja California. The mean precipitation of about 30–60 mm over San Bernardino is well captured by the model; however, on a daily basis the largest rms errors are seen over these mountains (Fig. 11c). Colle et al. (1999) noted a similar problem in the mountains of the Pacific Northwest, but observed an improvement in the estimation of model precipitation when the spatial resolution increased from 36 to 12 km. For instance, they found that the coarse grid produced better estimates of leeward precipitation, while the finescale grid produced better results in the windward side. This is also consistent with the results documented by Colle and Mass (2000) and Colle et al. (2000). In the present work we did not find a significant difference between the rmse at 30- (not shown) and 10-km resolution (Fig. 9); both grids produced smaller errors in the leeward side and larger values over the windward side. However, D3 with a more detailed topography improved the estimation of daily extreme precipitation events (>10 mm day⁻¹), consistent with the bias scores in Table 1.

The lack of a significant difference in the rmse between domains D2 and D3 could be the result of using a two-way feedback in this analysis, as one of our reviewers suggested. Based on this observation, we did additional 36-h tests for the first extreme event of the study period to compare the precipitation errors with one- and two-way feedbacks. However, the results did not reveal any significant differences in the mean precipitation for this particular case (not shown). Roebber et al. (2004) describe several cases from the literature where increasing model resolution does not necessarily lead to greater skill. These authors argue that a combination of high-resolution models, which assist in the development of conceptual models of various meso-scale phenomena, and ensemble forecast methods, which help quantify forecast uncertainty, may improve the forecast process.
5. Conclusions

The focus of this study was the numerical simulation of the heavy-precipitation events observed during 5–21 January 1993 in California and Baja California. During this period, below-normal height anomalies over the eastern subtropical Pacific and anomalous SST warming from the date line at the equator to Baja California were associated with an intensified and southward-displaced subtropical jet stream and persistent southwestern low-level moisture transport into the study area. These conditions and the passage of several frontal systems favored strong convective activity in the west coast and mountainous regions of the study area. Record precipitation totals were observed in southern California and northern Baja California, which produced severe floods and the loss of human life and damages valued in millions of dollars.

The MM5 reproduced reasonably well the structure of the mean synoptic atmospheric circulation during the study period, but with slightly less intensity than observed. The model captured the mean heavy precipitation (>10 m day\(^{-1}\)) in the coastal stations over the study period, but on average tended to underestimate heavy rainfall and to overestimate light and moderate rainfall (<10 mm day\(^{-1}\)). The underestimation of heavy rainfall could be partially associated with mistiming of the precipitation onset and the lower amounts of moisture flux simulated over the region. Other mesoscale models also tend to simulate too much light or moderate rainfall (e.g., Gershunov et al. 2000; Kato et al. 2001; Trenberth et al. 2003) partially due to cloud parameterization problems (Kain 2004).

On a daily basis, the model triggered the onset of the first heavy event (6 January) several hours before than observed, indicating that the vertical structure of the lower troposphere over San Diego was more unstable than observed; in addition, the winds and the vertical wind shear were also stronger than observed during the first few hours of the day. In many cases the model triggered the onset of precipitation a few hours before or after the observed precipitation, which could be at-
tributed to errors from the radiation, land surface, and/or planetary boundary layer (PBL) schemes selected. For example, the Hong and Pan (1996) PBL parameterization scheme utilized may have errors in the placement and timing of synoptic-scale features, which affects surface-layer winds, temperature and moisture gradients, and precipitation. Future tests with different PBL, radiation, and land surface parameterization schemes are needed to determine the physical sources of this bias.

The largest model errors were observed over the higher terrain (e.g., San Bernardino Mountains); this result is consistent with studies of the Pacific Northwest precipitation cited before. In general, the increase from 30- to 10-km resolution did not make a noticeable improvement in the estimation of mean precipitation, but there was an improvement in the estimation of extreme rainfall events by the finest domain, possibly due to a better representation of the regional topography. However, the lack of a larger number of stations in the mountains of southern California and Baja California makes it difficult to assess the true skill of the MM5 over complex topography.

The recent storms that produced flash floods, landslides, several deaths, and million-dollar damages from Santa Barbara to Tijuana from October 2004 to February 2005 highlight the importance of improving mesoscale predictions, model ensembles, and physical parameterizations to prevent and mitigate disasters. It is also very important to augment the number of surface observation sites and radiosonde data to improve model verifications in the study area.

Acknowledgments. This paper benefited from the constructive criticism and insightful suggestions of two anonymous reviewers and from the editor W. Kustas.

<table>
<thead>
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<th>Threshold (mm day^{-1})</th>
<th>Precipitation class</th>
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<th>MM5 D3 (10 km)</th>
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<tr>
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Fig. 10. Same as Fig. 9, but for 1200 UTC 6 Jan 1993.
This publication was funded by CONACYT-Mexico through an M.S. scholarship to Cerezo-Mota and is partially funded by NOAA Cooperative Agreement NA67RJ0155 to the University of Washington (UW) and Subcontract 739122 from UW to CICESE. We give special thanks to Julián Delgado of CICESE’s computing department for assistance during the performance of MM5 simulations. The MM5 is maintained by NCAR, which is sponsored by the National Science Foundation.

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Fig. 11. (a) Mean daily observed precipitation (mm day\(^{-1}\)) at local stations during 5–21 Jan 1993 and mean daily model precipitation in D3 (10 km) in contours, and (b) rmse between observed and model precipitation (mm day\(^{-1}\)). Dots indicate the location of the observation sites. Contour interval in (a) is 10 mm day\(^{-1}\), except for the 5 mm day\(^{-1}\) contour.


