Effects of Increased Horizontal Resolution on Simulation of the North American Monsoon in the NCAR CAM3: An Evaluation Based on Surface, Satellite, and Reanalysis Data

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ABSTRACT

Simulation of the North American monsoon system by the National Center for Atmospheric Research (NCAR) Community Atmosphere Model (CAM3) is evaluated in its sensitivity to increasing horizontal resolution. For two resolutions, T42 and T85, rainfall is compared to Tropical Rainfall Measuring Mission (TRMM) satellite-derived and surface gauge-based rainfall rates over the United States and northern Mexico as well as rainfall accumulations in gauges of the North American Monsoon Experiment (NAME) Enhanced Rain Gauge Network (NERN) in the Sierra Madre Occidental. Simulated upper-tropospheric mass and wind fields are compared to those from NCEP–NCAR reanalyses. The comparison presented herein demonstrates that tropospheric motions associated with the North American monsoon system are sensitive to increasing the horizontal resolution of the model. An increase in resolution from T42 to T85 results in changes to a region of large-scale midtropospheric descent found north and east of the monsoon anticyclone. Relative to its simulation at T42, this region extends farther south and west at T85. Additionally, at T85, the subsidence is stronger. Consistent with the differences in large-scale descent, the T85 simulation of CAM3 is anomalously dry over Texas and northeastern Mexico during the peak monsoon months. Meanwhile, the geographic distribution of rainfall over the Sierra Madre Occidental region of Mexico is more satisfactorily simulated at T85 than at T42 for July and August. Moisture import into this region is greater at T85 than at T42 during these months. A focused study of the Sierra Madre Occidental region in particular shows that, in the regional-average sense, the timing of the peak of the monsoon is relatively insensitive to the horizontal resolution of the model, while a phase bias in the diurnal cycle of monsoon season precipitation is somewhat reduced in the higher-resolution run. At both resolutions, CAM3 poorly simulates the month-to-month evolution of monsoon rainfall over extreme northwestern Mexico and Arizona, though biases are considerably improved at T85.

1. Introduction

During the twentieth century, the North American monsoon system became recognized as a significant climate regime in the southwestern United States and northwestern Mexico. Among the earliest written reports of a monsoonlike climate in this region are those of Campbell (1906) and Blake (1923), who document the existence of “Sonora storms” that prevail over the mountains and deserts of southern California from July into early October. As early as there have been documentations of the North American monsoon, there has been debate over its causes and its primary moisture source. Reed (1933) described the existence and movement of an upper-level anticyclone during the summer, around which air circulates to provide conditional instability over the southern Rocky Mountain states of the United States. Though several authors have cited the Gulf of Mexico as a primary moisture source for the monsoon (see Adams and Comrie 1997), other studies suggest that, at low levels, the Gulf of California provides most of the water vapor (Badan-Dangon et al. 1991; Schmitz and Mullen 1996; Higgins et al. 1997), while the Gulf of Mexico provides water vapor at upper levels (Schmitz and Mullen 1996; Higgins et al. 1997).

The region most influenced by the monsoon appears to be the western slopes and foothills of the Sierra Madre Occidental (SMO), where the rainfall change from June to July is largest (Douglas et al. 1993).

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ever, the monsoon is more than just a local phenomenon. It is a continental-scale system. The onset of the monsoon in northwestern Mexico and the southwestern United States is associated with broad belts of decreased precipitation to the north and east of this region (Barlow et al. 1998), such as over the Great Plains and northern tier of states, with expansion of the middle- and upper-level monsoon high (Higgins et al. 1997). A study by Reyes and Cadet (1988) suggests that even as far away as the South Pacific, the intensification of an anticyclonic gyre is a major forcing mechanism for the low-level monsoon moisture-laden circulation over western Mexico. And the monsoon is more than just a hydrologic forcing on the climate system. Barlow et al. (1998) show that in the monsoon’s mature phase diabatic heating over northwestern Mexico is as large as 1 K day\(^{-1}\), which influences continental-scale divergence and vorticity dynamics.

Accurately predicting the interannual variability of the monsoon system would aid water resource management, particularly in the arid southwestern United States and northwestern Mexico. It is fundamental that atmospheric models correctly capture the general features of the monsoon, including its onset and evolution, as reflected in the precipitation and other synoptic fields. In recent years, there have been many modeling studies of the North American monsoon, especially with mesoscale models. For global models, Arritt et al. (2000) evaluated its simulation in the second-generation Hadley Centre Model in which they found a slightly higher-than-observed standard deviation of monsoon onset date and an upper-level high simulated too far to the east and too weak. And Yang et al. (2001) tested the simulation of the monsoon in the National Center for Atmospheric Research (NCAR) Community Climate Model version 3 (CCM3) forced by climatological SSTs. Comparing against the National Centers for Environmental Prediction (NCEP)–NCAR reanalyses, they observed that CCM3 underestimates zonal-mean warm-season rainfall in Arizona and New Mexico and that the 200-mb anticyclone is displaced too far to the east.

Although mesoscale circulations make an important contribution to monsoon rainfall, as investigated by Berbery (2001), a global climate model cannot be expected to resolve them. It has been shown by Yang et al. (2001) that at T42 resolution, CCM3 cannot resolve well even the Gulf of California, a significant moisture source for the monsoon. The sensitivity of global model simulations to horizontal resolution has been the subject of much research over the past two decades (e.g., Boyle 1993; Kiehl and Williamson 1991; Boville 1991; Giorgi and Marinucci 1996; Senior 1995; Sperber et al. 1994; Lorant and Royer 2001), in which sensitivities of the simulations of different variables by different models have been investigated, all with widely varying results. In a recent study by Mo et al. (2005), simulations of warm-season rainfall over the United States and Mexico by the NCEP Global Forecast System (GFS) were analyzed. They showed that the model’s simulation of precipitation over the U.S. southwest and northwestern Mexico is improved when increasing the resolution from T62 to T126. The present study measures the sensitivity to horizontal resolution of CAM3 simulations of the warm-season rainfall associated with the North American monsoon system. It identifies any improvements or degradations in the simulation of the precipitation and upper-level height fields that result from increasing the horizontal resolution from T42 to T85, using the latest rainfall product of the National Aeronautics and Space Administration’s (NASA) Tropical Rainfall Measuring Mission (TRMM) satellite for northern Mexico, hourly surface gauge data for the United States, and monthly means from the NCEP–NCAR reanalysis for both. Additionally, this study evaluates the model’s simulation of monsoonal precipitation over the Sierra Madre Occidental region of Mexico, where complex topography and the presence of a narrow, yet significant, moisture source require a greater density of grid points for their adequate resolution. Specifically, it measures how much improvement can be realized in the simulation of this region by increasing the model’s resolution from T42 to T85, using for the evaluation a dense network of tipping-bucket rain gauges associated with the North American Monsoon Experiment (NAME).

2. Methods

a. Data

For the purposes of this study, several different datasets were used for validating the model. For evaluating the sensitivity of the CAM3 rainfall simulation to increasing horizontal resolution, hourly rainfall rates over four monsoon seasons from NASA’s Tropical Rainfall Measuring Mission (TRMM) satellite (Simpson et al. 1998; Kummerow et al. 1998, 2000) were used. Instantaneous rainfall rates were derived from the combination of measurements from the TRMM Microwave Imager (TMI) and the TRMM Precipitation Radar (PR). Available in the 3G68 (v.6) dataset, these rates have been averaged over 0.5\(^\circ\) × 0.5\(^\circ\) latitude–longitude boxes between 38\(^\circ\)S and 38\(^\circ\)N. Satellite sampling error is mitigated by averaging over multiple years, to ensure adequate coverage of all points in the monsoon domain. For locations poleward of 38\(^\circ\)N, a gridded hourly
precipitation rate dataset (hereafter referred to as HPD) was obtained from the National Weather Service (NWS)/Techniques Development Laboratory. The observations were derived from first-order NWS stations and cooperative observer stations and are gridded into $2.0^\circ \times 2.5^\circ$ boxes over the region $20^\circ$–$60^\circ$N, $140^\circ$–$60^\circ$W using a modified Cressman scheme. The data, available from 1963 to the present, are described in detail in Higgins et al. (2000). Both the TRMM data and the HPD were averaged onto a T42 grid. To evaluate the sensitivity of the CAM3 mass and wind fields to increasing horizontal resolution, monthly mean simulated 200-mb geopotential height and 200-mb horizontal wind, as well as 500-mb vertical velocity, were compared to those from the NCEP–NCAR reanalysis, as averaged over multiple monsoon seasons.

Measurements from surface rain gauges installed for the North American Monsoon Experiment (NAME) were used for a focused evaluation of the model’s monsoonal rainfall simulation over the Sierra Madre Occidental region of northern Mexico. The event-logging, tipping-bucket rain gauges are part of the NAME Enhanced Rain Gauge Network (NERN), as installed between 2002 and 2003 in major east–west transects of the Sierra Madre Occidental (Gochis et al. 2003a,b). Hourly rainfall accumulations from 83 of these gauges over three summers were used for the validation. Refer to Fig. 1, which shows the gauge sites in the SMO region. As these rain gauges are irregularly spaced, the measured rainfall accumulations were collected into an artificial $0.5^\circ \times 0.5^\circ$ mesh. For example, for hour $t$ of the NERN observational period, the rainfall totals from NERN gauges $g$ are denoted $r(g; t)$. For $g$ located in mesh cell $k$, $r(g \in k; t)$ was summed to obtain mesh cell total $R_k(t)$. The mesh cell totals were then averaged onto the model’s T85 horizontal grid to obtain $R(t)$. The phase and amplitude of the diurnal cycle were determined by fitting the regional-mean hourly mean precipitation rates $R(t)$ to the diurnal harmonic, such that

$$R(t) = f(t) + \delta(t), \quad (1)$$

where

$$f(t) = a \cos \left(\frac{2\pi t}{24}\right) + b \sin \left(\frac{2\pi t}{24}\right) + c, \quad (2)$$

and $\delta(t)$ is a residual. Estimates of the coefficients, $\hat{a}$ and $\hat{b}$, were determined by least squares regression, and formulas for the phase and amplitude estimates follow as

$$\hat{\rho} = \arctan \left(\frac{\hat{b}}{\hat{a}}\right) \quad (3)$$

and

$$\hat{A} = (\hat{a}^2 + \hat{b}^2)^{1/2}. \quad (4)$$

The phase and amplitude of the annual cycle were determined from visual inspection of the regional-mean monthly mean precipitation rates.

Since the distribution of NERN gauges within the SMO region is irregular, there is likely to be sampling uncertainty in estimating the phases and amplitudes characteristic of the entire region. Assuming the true values are known, the deviations of the estimates from the true values would be denoted $\epsilon_p$ and $\epsilon_A$. Since the true values are not known for this region, it is helpful to obtain some idea of the range of these deviations. Possible deviations $\hat{\epsilon}_p$ and $\hat{\epsilon}_A$ were measured via a Monte Carlo technique, described as follows.

Assume $\mathcal{N}$ is the population of gauges such that

$$g_1, g_2, \ldots, g_{83} \in \mathcal{N} \quad (5)$$

where $g_i, i = 1, \ldots, 83$ belong to the NERN network. Observation time $t$ is mapped to a specific hour of a specific month and year of the NERN observing period [i.e., $t \rightarrow (m, y)$] and is associated with a distribution of rainfall accumulations $D_N(t)$. A given subset of $\mathcal{N}$, $\mathcal{N}_k$, belongs to mesh cell $k$ and for a given time $t$ is associ-
ated with rainfall accumulation $R(t)$. Assume also that $\mathcal{M}$ is a population of gauges such that
\[ g_1, g_2, \ldots, g_{83}, h_1, h_2, \ldots, h_S \in \mathcal{M}, \]
where $h_i, i = 1, \ldots, S$ represent artificial gauges, spaced regularly so that their total number $S$ is inversely proportional to the spacing $\Delta$. A given subset of $\mathcal{M}, \mathcal{M}_k$, belongs to mesh cell $k$. At time $t$, the rainfall accumulations for artificial gauges $h \in k$ are determined by drawing randomly from $D_N(t)$. In this way, the synthetic rainfall accumulations at time $t$ are characteristic of the actual rainfall accumulations at this time, as measured in the NERN gauges. The synthetic rainfall accumulations are then summed over the mesh cell to provide $R_M^k(t)$.

From these two gauge populations, two realizations of regional-mean hourly mean rainfall can be defined: $R(t)$ and $R_M(t)$, where $R(t)$ is computed from the original NERN gauge measurements and $R_M(t)$ is computed from the measurements of both the original NERN gauges and the regularly spaced hypothetical gauges, as described above. Refer to Fig. 2 for the spatial distributions of the artificial networks. Note that the hypothetical gauges only fill the gaps in the SMO region, as shown with heavy solid lines. Both $R(t)$ and $R_M(t)$ represent averages over the SMO region on the model’s T85 grid. To estimate the spread of phase and amplitude deviations, four realizations of $R_M$ were generated for each of three horizontal spacings: $\Delta = 0.25^\circ, 0.5^\circ$, and $0.75^\circ$, corresponding to $S = 422$ (an approximate 500% increase in gauges), 92 (an approximate 100% increase), and 39 (an approximate 50% increase), respectively. The choices for the spacings were somewhat arbitrary; however, it was of interest to measure the sensitivity of the parameter estimates to the number of sample points. Thus, for each spacing, there were four estimates of deviations $\hat\varepsilon_p$ and $\hat\varepsilon_A$, and these were found to be sufficient for measuring the observational uncertainty.

b. Simulations

CAM3 is a three-dimensional global spectral model, which may be run at a number of different horizontal resolutions. Vertically, it uses 26 levels. Its parameterization for convection is the scheme of Zhang and McFarlane (1995), which relates the cloud-scale contributions of heat and moisture to the entrainment and detrainment rates of subgrid-scale convective plumes. Closure of the scheme follows the quasi-equilibrium theory of Arakawa and Shubert (1974) in which destabilization of the atmosphere due to large-scale processes is approximately balanced by stabilization due to

*Fig. 2. Maps showing the network of NERN rain gauges (heavy diamonds) and that of the artificial gauges (light diamonds) for gauge spacings of (a) $0.25^\circ$, (b) $0.5^\circ$, and (c) $0.75^\circ$. In network 1, 422 gauges are added to the existing network; in network 2, 92 are added; and in network 3, 39 are added. Dashed lines represent the model’s T85 grid lines. Solid lines represent a $0.5^\circ \times 0.5^\circ$ mesh.*
convection. The amount of convective rainfall is a function of the cloud-base mass flux, which itself is modulated by the change of CAPE. As for its boundary layer parameterization, for unstable or convective conditions, transport of heat, water vapor, or passive scalars by convective turbulence is related to the nonlocal eddy diffusivity for the quantity of interest, which itself is related to the convective velocity scale. The land surface model was initiated concurrently with the atmospheric model using 1 September climatological conditions, and it is interactive. For further details of the model’s formulation, the reader is referred to the paper by Collins et al. (2006). Various aspects of the global climate simulated by CAM3 have been documented in a special issue of Journal of Climate (Vol. 19, No. 11; see, e.g., Hack et al. 2006a,b; Rasch et al. 2006). Focused efforts on the analysis of the regional aspects of the model simulation such as the North American monsoon are yet to come. Recently, a modified version of the Zhang and McFarlane (1995) parameterization for rainfall simulation (Zhang 2002) was tested. It was shown that the modifications produce favorable results for simulation in the Tropics (Zhang and Mu 2005a,b). However, in the domain of this study, the modifications degraded the simulation of rainfall. For the two resolutions tested, the modified version of the model produced an unrealistically weak monsoon. Therefore, in this paper, the results from the simulations with the original version of the model are documented, and a more thorough analysis of the results obtained with the modified parameterization is reserved for a later study.

To isolate the model’s monsoon signal from its natural variability, ensembles of simulations were composed. Refer to Fig. 3, which provides a diagram of the design of the simulation ensembles as well as the periods of observational data. Note that IC indicates an initial condition dataset while BC indicates a boundary condition dataset. Pentagons indicate monsoon samples used for generation of the ensembles.

![Diagram showing the design of the T42 and T85 simulation ensembles as well as the overlapping periods of observational data. Note that IC indicates an initial condition dataset while BC indicates a boundary condition dataset. Pentagons indicate monsoon samples used for generation of the ensembles.](image-url)
Two simulations were generated independently in that the second one used a variation of the initial condition applied to the first, a 1 September climatology provided by NCAR. For example, T42 simulation A used IC\textsubscript{A,T42}, B used IC\textsubscript{B,T42}, and so on. At each of 26 vertical levels, the initial two-dimensional temperature field for simulation B was randomly perturbed from that of A such that

\[ T_B = T_A + (1 - 2R_B) \exp(k - 26), \]

where \( k \) is the vertical level index, increasing from 1 to 26 downward and \( R_B = 0.655436218 \) as prescribed by a random number generator. This formulation specifies a temperature perturbation on the order of \( 10^{-1} \) K at the lowest atmospheric level, decreasing to a perturbation on the order of \( 10^{-12} \) K at the highest atmospheric level. Note that the same value of \( R_B \) was used at T42 and at T85. Generating these independent simulations served to increase the ensemble size for validation. For validation of the mass and wind fields with the NCEP–NCAR reanalysis, each year’s monsoon was treated as an independent sample, such that the 9 yr of simulation A, which extends from September 1993 to September 2002, provided a nine-member ensemble. For validation of the precipitation simulation with the combination of HPD and TRMM, four monsoon samples were taken from simulation B, which extends from September 1997 to September 2002, and from simulation A from the same years to provide an eight-member ensemble for both T42 and T85. In a fifth simulation, CAM3 was run at T85 resolution for September 1998–September 2004 to overlap the observational period of the NERN rain gauges. This particular simulation used the same boundary conditions of the other T85 simulations and initial conditions of T85 simulation A (IC\textsubscript{A,T85}). The precipitation totals of the SMO region as averaged over the three monsoon seasons of 2002, 2003, and 2004 were compared to those as averaged from the NERN rain gauges, which themselves were checked against TRMM rainfall, available for the same time period.

3. Results

a. The North American monsoon system

The evolution of the North American monsoon is evidenced in Fig. 4, which displays the 4-yr monthly mean precipitation rates for the North American monsoon region, as outlined in the map.

As shown by Collier and Zhang (2006), the evolution of the monsoon varies for particular subregions inside the domain outlined in Fig. 4. However, observations from TRMM and the HPD show that, on average, the region as a whole experiences a monsoon onset in June.
with an almost doubling of precipitation over that of May, a mature phase in July and August, and a decay in September. Meanwhile, to the east, in the U.S. southern Great Plains region, the annual cycle of precipitation is out of phase with that of the monsoon region. This region, encompassing Kansas, Oklahoma, and Texas, experiences, on average, an annual maximum in precipitation during June, the month of monsoon onset, and a sharp decline in precipitation during the months of monsoon maturity, July and August.

Via use of the NCEP–NCAR reanalysis over the period 1979–94, Higgins et al. (1997) showed that evolution of the continental precipitation regime could be related to evolution of the tropospheric circulation and mean vertical motion fields. They observed that in the vicinity and south of a monsoon high, evident in the 200-mb wind field, divergence enhanced midtropospheric vertical motion and thus monsoon rainfall. Meanwhile, to the north and east of the monsoon high, they showed that the flow is more convergent, that there is midtropospheric subsidence, and that there is less rainfall. In addition, their study found that upper-tropospheric convergence in this region acts to limit rainfall related to the Great Plains low-level jet. Contour plots of seasonal-mean upper-tropospheric winds, geopotential height, and 500-mb vertical velocity, as well as 850-mb specific humidity and winds from the NCEP–NCAR reanalysis for 1999–2002, are shown in Fig. 5. Also, precipitation, as observed from the HPD dataset and from TRMM, is shown for the corresponding time period. There is a clearly defined clockwise circulation of the 200-mb winds over northern Mexico characteristic of the monsoon high. To the west and northwest of the high, the 200-mb winds are divergent, and the 500-mb vertical motion is upward. In the lower troposphere, the moisture tongue associated with the southeasterly flow from Mexico and the Gulf of Mexico and the southwesterly flow from the Gulf of California extends northward into the southwestern states of the United States. The large-scale forcing, together with the abundance of low-level moisture, produce the monsoon precipitation seen across northwestern Mexico and the southwestern United States in Fig. 5c. Note that, in the SGP region indicated in Fig. 4, upper-tropospheric winds are nondivergent and vertical motion is generally downward. These patterns are generally consistent with the findings of Higgins et al. (1997) and establish the connection of monsoon-circulation-induced large-scale forcing on convection and precipitation in the domain of this study.

The sensitivity of the simulation of the North American monsoon system to horizontal resolution was measured by comparing the monthly mean mass, wind, and precipitation fields among the T42 and T85 ensembles, NCEP–NCAR reanalysis, and surface gauge and TRMM rainfall data for a period encompassing the monsoon onset, maturity, and decay: the 6-month period of May–October, as averaged over 1994–2002. For this comparison, all simulated or observed fields were interpolated to T42, the resolution of the coarsest simulation.

Evaluation of the simulated upper-tropospheric mass and wind fields was facilitated by plotting the monthly mean 200-mb geopotential height, 200-mb horizontal winds, and the vertical pressure velocity at the 500-mb level for both the T42A and T85A ensembles, as well
as for the NCEP–NCAR reanalysis. These plots are shown in Figs. 6 and 7 for May–July and August–October, respectively.

In the premonsoon month of May, upper-tropospheric flow is predominantly zonal with a geographically uniform pressure gradient. Vertical motion is upward over much of the south-central and southwestern United States, while downward motion dominates over the western Gulf of Mexico, extreme northeastern Mexico, and the eastern subtropical Pacific. Both the T42 and T85 simulations capture the zonal flow with reasonable success this month. At both resolutions, the simulated 200-mb height is too large, and the model has trouble capturing the midtropospheric vertical motion pattern. In the onset month of June, the 200-mb height values are higher than in May everywhere in the domain, and the height pattern and winds show a more ridgelike orientation. The downward motion over the western Gulf of Mexico and southern Texas is weakened compared to that in May. In the simulations, the atmospheric thickness is even larger. At T42, there is a region of maximum 200-mb height over central Mexico, a clearly defined monsoon anticyclone. As for the vertical motion field, downward motion occurs over a small part of the southern Great Plains at both resolutions. North and south of this region, upward motion is found and is anomalously strong.

In July, the region of large-scale downward motion encompasses most of the southern Great Plains of the United States, along 100°W, or just to the east of the
ridge axis through western Texas and north-central Mexico. In terms of magnitude, the mean large-scale descent at both resolutions is too strong over Texas and Oklahoma. At T42, the region of downward motion does not extend far enough north, occupying only northern Texas and Oklahoma. This leaves Kansas in a region of anomalous large-scale ascent. Likewise, at T85, the region of downward motion does not extend far enough north. Instead, it extends too far south and too far west. At both resolutions, upper-tropospheric winds are reasonably well simulated, though they are slightly too strong both poleward and equatorward of the high. During August (Fig. 7), the mature phase of the monsoon, the upper-level pressure gradient is most relaxed, and the neighboring regions of midtropospheric upward and downward motion are qualitatively similar to those of July with a line of demarcation at roughly 105°W. Just as in July, at both resolutions, the region of downward motion over the U.S. southern Great Plains does not extend far enough north. At T85, it extends too far south and too far west and is anomalously strong. The upward motion tongue extending from Mexico into Arizona and New Mexico is, however, reasonably well simulated. In the waning monsoon months of September and October, the 200-mb geopotential height is lowered, the upper-tropospheric pressure gradient and wind speeds are larger, and the region of mean downward motion is expanded westward. The midtropospheric vertical motion and upper-level wind patterns are generally well simulated at both resolutions during these 2 months though simulated 200-mb heights are too large. In general, in terms of the overall evolution of the upper-tropospheric mass and wind fields during the onset, mature, and decay periods.
of the North American monsoon season, the most significant differences between the low- and high-resolution model biases are that, during July and August, the T85 simulation shows a region of large-scale descent over the U.S. southern Great Plains that extends much farther south and west than that at T42 and in the reanalysis. West of about 105°W, the direction of vertical motion agrees nicely among the simulations and the observations for these same months, though the model has a tendency to overestimate the magnitude of the vertical motion in some areas.

For the evaluation of the monsoon precipitation simulation, 1999–2002 TRMM satellite data and hourly gauge-based precipitation from the HPD were compared to the eight-member ensemble-average monthly means from the simulations (T42A, B and T85A, B), which consist of the monsoon samples from the same time period. Figure 8 shows the monthly precipitation for the months of May–July.

Simulated warm-season rainfall is quite sensitive to the change in horizontal resolution applied to CAM3. In May, while both the T42 and T85 simulations are too dry over eastern Texas and too wet over the western Gulf of Mexico, the T42 ensemble is noticeably wetter than the T85 ensemble over northeastern Mexico. In June, monsoonal rainfall begins for portions of northwestern Mexico. At both resolutions, the models show increased precipitation over northern Mexico during this month; however, the geographic distribution of rainfall is quite different between the two resolutions. At T85, the rainfall maximum is more linear, extending from southeast to northwest along the western coast of Mexico. This pattern is consistent with the pattern evident from the rainfall observations map. It is unlike the
pattern at T42, in which the rainfall belt is less well defined. At both resolutions, there is a dry bias over extreme northwestern Mexico during this month. Over the southern Great Plains, Texas remains relatively wet during June, with rainfall rates averaging between 2 and 3 mm day$^{-1}$ over the western part of the state and between 3 and 5 mm day$^{-1}$ over the eastern part. By contrast, the simulation at T42 is far too wet over western Texas, showing rainfall rates exceeding 4 mm day$^{-1}$ in a belt of relatively high rainfall extending along 100°W. While the simulation of western Texas rainfall is comparatively better at T85, rainfall is anomalously low over the central part of the state. This dry “hole” is expanded in all directions for T85 in July, while rainfall rates increase over western Mexico especially over the Sierra Madre Occidental. The increase in rainfall over western Mexico is in general agreement with the observations. While the geographic maximum at T85 is positioned anomalously south, the linear rainfall belt more closely resembles the observations than the broader feature at T42. Though T42 shows a wet bias over northeastern Mexico during July, it is more realistic over extreme western Texas, where T85 is anomalously dry. Once again, as during June, rainfall rates over extreme northwestern Mexico, and now extending into southeastern Arizona and southwestern New Mexico, are simulated too low relative to the observations.

Figure 9 shows the monthly mean precipitation rates for August–October. The dry bias over Arizona and extreme northwestern Mexico persists in the simulations for August, though at T85, the rainfall belt along the Gulf of California coast extends farther north, an improvement over T42. On the other hand, the dry bias at T85 over the southern Great Plains is expanded to include northeastern Mexico between 20° and 25°N. At the same time, however, the T42 wet bias around 40°N is removed at the higher resolution. During the months of monsoon decay in September and October, both

Fig. 9. Same as in Fig. 8 except for August–October.
simulations show a wet bias over extreme western Mexico and in the Gulf of California and a dry bias over Texas, particularly in October.

Thus, in terms of rainfall, its geographic distribution over western Mexico is better simulated at T85 than at T42, particularly during August. By contrast, however, at T85, rainfall is too low over Texas and northeastern Mexico during this same month. At both resolutions, CAM3 is too dry over extreme northwestern Mexico and Arizona during July and August. Refer to Fig. 10, which shows precipitation as averaged over this region.

It is evident that the T42 ensemble produces a late monsoon, beginning in August, peaking in September, and ending in November, while the T85 ensemble shows a weaker monsoon beginning in June, peaking in September, and ending in October. In addition, both ensembles show much wetter winters than do the observations, with the annual peak of rainfall occurring in December, as opposed to August. In fact, August represents the annual minimum in precipitation for the T42 ensemble. It is interesting to note from the table in Fig. 10 that the total June–September precipitation amounts from both ensembles reside within 25% of the observed total amount. However, month-to-month evolutions of rainfall do not agree as well between the model and the observations. In the observations, about 75% of the seasonal rainfall occurs in July and August, with August contributing the largest percentage. At T42, only 11% comes from July and August, with over 80% contributed during September. At T85, just under half of the seasonal rainfall occurs during these 2 months. While this proportion is still too low relative to
the observations, at T85, this region as a whole experiences much greater precipitation during these peak monsoon months than at T42. This is evident in the monthly means plot. For example, August precipitation at T85, still biased low by 50% of the observed mean, is at least 6 times greater than that at T42. This increase likely is due to the aforementioned northwestward extension of the western Mexico rainfall belt (see Fig. 9). In light of this improvement, CAM3 does not simulate well the evolution of rainfall in this region at either resolution. The model shows an anomalously late monsoon peak in September.

Comparison of Figs. 6 and 7 with Figs. 8 and 9 shows clear correspondence between the large-scale vertical velocity and precipitation in both the observations and simulations. Similar correspondence applies qualitatively to the differences between T42 and T85 simulations as well. For July and August, the differences are shown in Fig. 11.

Increasing the resolution of the model from T42 to T85 results in an increase in July and August precipitation for the western Gulf of Mexico and for northwestern Mexico and Arizona and a widespread decrease in precipitation over the south-central United States and northeastern Mexico. This pattern is, to some extent, similar to that shown in Mo et al. (2005) when they increased their model’s resolution from T62 to T126. However, the precipitation increase in northwestern Mexico and Arizona is much smaller here than in their case. This is probably due to the fact that their models are at higher resolutions.

The greatest amount of “drying” occurs over northeastern Mexico, just south of Texas. This region is also characterized by a relatively large net positive difference in the vertical pressure velocity (weakening of ascent or strengthening of descent from T42 to T85). For individual years, the patterns are similar. Figure 12 shows the scatterplot of precipitation difference versus 500-mb vertical velocity difference for both T85A-T42A and T85B-T42B. In the plot, each point represents either a July or August mean at a grid point in the domain of the map in Fig. 11. While this figure shows the combination of the differences for both sets of simulations, it should be noted that the relationship between the vertical velocity differences and the precipitation differences is qualitatively similar for each set. Generally speaking, positive changes in 500-mb vertical velocity are associated with negative changes in precipitation rate. The converse also is true.

In addition to its vertical motion field, the model’s water vapor field also is sensitive to horizontal resolution. This sensitivity may account for some of the rainfall differences between T42 and T85. To estimate this sensitivity and its contribution to the sensitivity of the mean precipitation, the ensemble-mean monthly mean precipitation at each grid point was decomposed into two parts: that owing to the import of moisture (mois-

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**Fig. 11.** The 1999–2002 mean differences between the T85 and T42 July–August mean precipitation rate (mm day$^{-1}$) and 500-mb vertical (upward) velocity (in mb day$^{-1}$). Where differences are positive, the quantity at T85 exceeds that at T42. The an ex (x) indicates the net T85 – T42 difference in 500-mb vertical velocity is downward (or less upward), while an open circle (o) indicates the net difference is upward (or less downward). Precipitation difference for the shaded area is shown in more detail in the inset.

**Fig. 12.** Scatterplot of 500-mb vertical velocity difference vs precipitation rate difference for July and August 1999–2002. T85A – T42A differences are shown as open circles while T85B – T42B differences are shown as filled circles. All points lie in the domain of the map in Fig. 11.
ture convergence) and that owing to surface evaporation. This was possible as the average change of specific humidity over a month is at least an order of magnitude smaller than the monthly mean moisture import, precipitation, and evaporation (not shown). Figure 13 shows the differences in precipitation, moisture convergence, and surface evaporation between T42 and T85. From the figure, both moisture convergence and evaporation differences contribute more or less equally to the negative precipitation difference over northeastern Mexico and Texas. By contrast, a positive moisture import difference explains most of the positive precipitation difference over the Sierra Madre Occidental region of Mexico and the western Gulf of Mexico. Just to the east of the SMO region, T85 shows less moisture import than T42. According to Fig. 14, which contours the 850-mb specific humidity and wind fields for the same months, regions of convergence appear to occur when southeasterly winds from the Gulf of Mexico meet or merge with northwesterly winds. At T42, during July and August, this region of convergence or confluence occurs at approximately 105°W. At 22°N, 105°W there is a maximum in low-level water vapor. At T85, the maximum water vapor is found at about the same position. However, the maximum wind convergence region is located closer to 110°W, and the high specific humidity values of greater than 10 g kg⁻¹ do not extend as far north and east as they do at T42. Comparing these contour maps to the corresponding contour map from the reanalysis in Fig. 5b, it is evident that the wind simulation at T85 for this region is slightly more realistic. However, note that 850-mb moisture is too low at both resolutions over northeastern Mexico and the U.S. southern Great Plains, particularly so at T85. In the reanalysis, the 10 g kg⁻¹ values extend north all the way into Kansas.

b. The Sierra Madre Occidental: Validation with NERN rain gauge data

Thus far, this study has focused on the large-scale dynamics and precipitation patterns associated with the North American monsoon system, and has shown that large-scale tropospheric motion over the North American monsoon region as simulated by the CAM3 is sensitive to model horizontal resolution. Additionally, moisture convergence is sensitive to the change in resolution and also affects precipitation, primarily in the Sierra Madre Occidental region. Since the Sierra Madre Occidental region of northwestern Mexico is generally regarded to be a core monsoon region, it is important to evaluate more closely the model’s precipitation simulation here and measure any improvement obtained by increasing the model’s horizontal resolution. It has become easier to evaluate precipitation in this region due to the recent installation of a fine network of rainfall gauges, the NAME Enhanced Rain Gauge Network (NERN). Installation of the NERN rain gauges over the SMO region has filled many voids in the mountainous western part of the country. In this section, CAM3 is validated at T42 and at T85 in the simulations of the diurnal and annual cycles of monsoonal precipitation in the SMO, using hourly rainfall accumulations available from the NERN gauges and those from TRMM. Refer to Fig. 1 for the location and boundary of the region of interest. To take advantage of as much data as possible for the model’s validation, the diurnal cycles represent averages over the June–September period.

The diurnal harmonic fits to the ensemble-mean regional-mean hourly mean precipitation rates for the four realizations of each network M are shown with the fit for the original NERN network X in Fig. 15. For the
three intergauge spacings tested, the diurnal harmonics are nearly indistinguishable from those of the NERN gauges alone, indicating relatively small uncertainty in estimating the phase and amplitude of the diurnal cycle over this region as a whole. For the SMO region in general, precipitation peaks between 1700 and 1900 LT. It should be emphasized that individual gauges may show earlier or later peaks. In fact, Gochis et al. (2003b) observed that the exact timing of the daily maximum depends on elevation, with lower-elevation stations showing peaks later in the afternoon, relative to higher-elevation stations. They found that the distribution of peaks from the first phase of gauges (installed in 2002) peaks between 1300 and 1800 local solar time (which should differ by at most an hour from local standard time, as shown in the results herein). The model’s regional average agrees with that of the observations as long as it falls within the bounds of the sampling uncertainty associated with the gauges. The NERN rain data indicate an amplitude of approximately 100% the daily mean, which appears to be relatively consistent with the network-mean diurnal cycle found by Gochis et al. (2003b) from phase 1 gauges. For individual locations, this amplitude may even be biased low, since the diurnal cycle shown here is partly representative of long periods with no precipitation. Using only wet days in the calculation, Gochis et al. (2003b) found a large increase in the peak mean hourly rain rate, particularly for the lowest elevation stations.

Figure 16 shows the diurnal harmonic fits of the regional-mean hourly mean precipitation rates for a June–September average. The two observational estimates, given by the solid curves, are from the NERN rain gauges and from TRMM for 2002–04. Shown for comparison is the fit from T85C, which is identical to T85A except that it covers the observational period of the gauges (2002–04), as well as the fits from the T42A and T85A simulations, as averaged over 1994–2001. At this point, it should be noted that, while the diurnal harmonics plotted represent averages over a 4-month period, the phase of the diurnal harmonic, as derived from regional-mean simulated rain rates, varies by at
most an hour within the months of June–September. Therefore, the diurnal harmonic derived from the 4-month average rain rates is a good representation of the model’s diurnal variability of rainfall over the SMO region during a given day of the monsoon season. It is reassuring to see that the TRMM-derived diurnal cycle compares well with that from the gauges, showing a peak at 1800 LT and a comparable amplitude. Furthermore, the phase of the diurnal cycle is consistent with that found by Gochis et al. (2004) for precipitation frequency in the NERN network mean. For the model runs, the diurnal cycles are tightly clustered together and precede the observations by 2–3 h. Such a result is consistent with the results of another diurnal cycle study, that of Dai and Trenberth (2004). Given the negligible uncertainty in estimating the phase of the diurnal cycle from the observations, it appears the model is biased early in its diurnal rainfall peak for this region. For the higher-resolution simulations, the phase bias is slightly reduced while the amplitude bias is enlarged. At T42, the amplitude of the diurnal cycle falls just barely outside the range of the amplitudes given by the NERN and the TRMM data. At T85, however, the amplitudes fall well outside the range of observational uncertainty.

The ensemble-mean regional-mean monthly mean precipitation rates for each network along with those from the original NERN network are shown in Fig. 17. Same as for the diurnal cycle estimation, any uncertainty in estimating the peak of the annual cycle from the NERN gauges is likely small. Neglecting subregional differences, precipitation over the SMO region as a whole is at a minimum in May and at a maximum in July. Also, the irregular spatial sampling of the NERN network does not appear to affect the observational estimate of when the monsoon begins on average and when it decays. Its onset, in May, is much more dramatic than its decay, which appears to be gradual. Also, there appears to be little uncertainty in estimating the amplitude of the annual cycle, or the strength of the monsoon. The NERN gauges alone indicate a large amplitude of over 2.5 times the annual mean. From the Monte Carlo simulations using the combination of the NERN and artificial gauges, the amplitudes are of comparable value.

The annual cycle of precipitation is shown in Fig. 18 for the two observational estimates: the T85C simulation over 2002–04, and the T42A and T85A simulations for 1994–2001. Between the two observational estimates, there is about a 20% difference in the amplitude of the annual cycle, with the NERN data showing the smaller peak. Among the simulations, there is a large spread of amplitudes as well. However, unlike that of the diurnal cycle, the amplitude of the annual cycle does not necessarily increase with increasing resolution. The amplitude of the T42 simulation falls between the amplitudes of the two T85 simulations. And the two T85 simulations differ in other ways as well. T85C shows greater winter precipitation than T85A. Second, while the monsoonal onset of T85C is more consistent with that of the observations, its decay is relatively more severe and less realistic. Nevertheless, both T85 simulations and the T42 simulation agree upon a monsoonal peak in July, which is consistent with the observational estimates. Thus, given the relatively large difference in amplitude between the gauge-based data and the satellite-based data and the even larger difference between the amplitudes of the two T85 runs, it is difficult to ascertain the effect of increased resolution on the overall strength of the monsoon. However, as previously shown, the geographic rainfall distribution in
the region is improved at the higher resolution during the core monsoon months of July and August. For the timing of the monsoon peak, there is no additional benefit to running the model at T85 rather than at T42.

4. Summary and conclusions

In this study, the simulation of the North American monsoon by the NCAR Community Atmosphere Model (CAM3) and its sensitivity to increasing horizontal resolution has been evaluated with observational data from surface-based gauges, TRMM satellite retrievals, and NCEP–NCAR reanalysis fields. For this experiment, the model was executed at two resolutions: the standard T42 (approximately 266 km × 266 km) and at T85 (approximately 133 km × 133 km). The results of the simulations provide important insights into the effects of increasing model resolution on the simulation of the monsoon system and the Sierra Madre Occidental region in particular. One of the major differences found between the relatively low-resolution and relatively high-resolution simulations is in the precipitation field. Increasing the resolution of the model from T42 to T85 results in a widespread decrease in precipitation during July and August over the U.S. southern Great Plains and northeastern Mexico. The area of reduced precipitation appears to correspond well with an enhanced region of large-scale descent, which, at T85, extends farther west and south than at T42. This difference in simulated dynamics may contribute to the negative rainfall bias here at the higher resolution. At the same time, the resolution increase results in an increase in monsoonal precipitation over northwestern Mexico and Arizona. Such an increase is somewhat consistent with the results of Mo et al. (2005). Despite the improvement, CAM3 is still anomalously dry in Arizona and extreme northwestern Mexico during the peak monsoon months. To the south of this region, over the Sierra Madre Occidental, the geographic distribution of rainfall appears to be better simulated at T85 than at T42 during July and August. Part of this rainfall difference can be attributed to a

**Fig. 17.** Regional-mean monthly mean precipitation rates from the original NERN gauges (dashed line) and from the real + synthetic gauge networks (solid lines) for 2002-04. The fit for a gauge spacing of (left) 0.25°, (middle) 0.5°, and (right) 0.75° is shown.

**Fig. 18.** Regional-mean monthly mean precipitation rates for the SMO region as averaged over 1994-2001 (T42A and T85A simulations) and over 2002-04 (NERN, TRMM, and T85C simulation).
difference in moisture import, which is higher at T85 than at T42. Averaging over the Sierra Madre Occidental region as a whole reveals that increasing the model’s resolution from T42 to T85 leads to little if any improvement in the simulation of the diurnal and annual cycles of precipitation here, though greater improvement may be found for individual subregions. For the diurnal evolution of rainfall, the model shows a 2–3-h phase bias at T42, and this is slightly reduced at T85. However, comcomitant with this reduction in phase bias is an increase in the diurnal cycle amplitude bias, which is relatively small at T42. As for the annual cycle of precipitation, the time of the monsoon peak in this region appears to be comparatively insensitive to increasing the horizontal resolution. It is unclear what effect increasing the resolution has on the monsoon peak amplitude.

The results of this evaluation indicate that tropospheric motions in and around the North American monsoon region are sensitive to horizontal resolution in CAM3. Increasing the resolution from T42 to T85 reveals an expanded region of enhanced large-scale descent north and east of the monsoon anticyclone, which appears to be consistent with a decrease in rainfall over Texas and northeastern Mexico. Meanwhile, the contribution of increased moisture import over the Sierra Madre Occidental region of northwest Mexico, as realized at T85, coincides with a more realistic geographic distribution of precipitation in this core monsoon region during August, as manifested in a more well-defined maximum in precipitation over the SMO, which is consistent with the observations. The causes of the differences in the simulated tropospheric dynamics requires further study and evaluation of the means of effecting changes in large-scale vertical motion, such as thickness advection and differential vorticity advection. As for the moisture transport differences, a subtle change in the low-level wind field, as caused by increasing the horizontal resolution, may influence the geographic distribution of available water vapor. Over Arizona and extreme northwestern Mexico, increasing the resolution of CAM3 from T42 to T85 does not completely eliminate the model’s dry bias here during the peak monsoon months, though it considerably reduces it. However, at both resolutions, the model shows an unrealistic evolution of monthly mean rainfall here, with an anomalously late monsoon peak in September. Further increasing the resolution of the model may show continued improvement for this important part of the monsoon domain. At an even higher resolution, the model may capture the Gulf of California low-level jet (Mo et al. 2005), a significant conduit for monsoonal moisture. However, it is unclear what effect a further increase in resolution will have on the precipitation simulation of surrounding regions.

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