# Hydrological Processes in Regional Climate Model Simulations of the Central United States Flood of June-July 1993


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ABSTRACT

Regional climate model (RCM) simulations of hydroclimate for the central U. S. are sensitive to RCM design, yet comparison of RCM results under common experimental conditions is rare. Thus the degree of and sources for inter-model variability are not well known. We have compared 60-d simulations of 1993 June-July from thirteen RCM simulations to each other and observations. Boundary data and initial conditions were supplied by the Project to Intercompare Regional Climate Simulations (PIRCS) experiment 1b. We have examined water vapor conservation and precipitation characteristics in each RCM for a $10^{O}$x$10^{O}$ sub-region of the Upper Mississippi River Basin (UMRB), containing the region of maximum 60-d accumulated precipitation in all RCMs and station reports.

Results showed that gross features of hydroclimate were well simulated in all RCMs. Specifically, all RCMs produced positive precipitation minus evaporation (P-E>0), and RCM recycling ratios were within the range estimated from observations. The range of P-E in RCMs enveloped the range of estimates of observed P-E, but most RCMs produced P-E below the estimated observed range. We found sensitivity of RCM E to radiation parameterization, including clouds, but inter-model variability of E was spread evenly about estimates of observed E suggesting little, if any, common errors of E among the simulations. In contrast, most RCMs produced P that was below the range of P from observations; thus a common dry bias of the simulations accounted for the low values of simulated P-E compared to observations.

Daily cycles of terms in the water vapor conservation equation revealed that P in most RCMs is driven by the dynamics of atmospheric circulation. In most simulations

nocturnal maxima of P and C (convergence) occurred simultaneously, consistent with observations of P and climatological studies of water vapor conservation. Three of the four driest RCMs had maximum P in the afternoon, while the time of maximum C was variable, suggesting that in these RCMs afternoon destabilization by insolation strongly influenced the precipitation process. When 60-d accumulated precipitation was decomposed as the sum of 3-h precipitation totals, a larger fraction of 60-d accumulated precipitation in all RCMs compared to station reports was from low 3-h totals. This tendency was exaggerated in the driest simulations. In station reports, accumulation from high 3-h totals had a nocturnal maximum, whereas accumulation from low 3-h totals had an early morning maximum. Satellite imagery suggests that this time lag between maximum accumulation from high and low 3-h totals occurred, in part, because many mesoscale convective systems had reached peak intensity overnight and had declined in intensity by early morning, while having significant overlap with the UMRB box. None of the RCMs contained this time lag between maximum accumulation from 3-h totals. We therefore recommend additional tests of the ability of RCMs to simulate the effects of mesoscale convective systems.

## 1. Introduction

Mesoscale processes and regional surface conditions influence the water cycle of the central United States (Rasmusson 1967, Fritsch et al. 1986, Higgins et al. 1997), suggesting that high-resolution models are necessary for detailed, physically based simulation of the region's hydroclimate. One approach to this problem is the use of a regional climate model (RCM) that nests a high-resolution limited-area model within the grid of a coarser-resolution analysis or climate model. A variety of RCM architectures exist, but systematic comparison of output from different RCMs is lacking (Giorgi and Mearns 1999). Understanding the sensitivity of hydroclimatic simulations to differences in RCM architecture advances our knowledge not only of hydroclimate but also of RCM design. RCM intercomparisons provide a common experimental framework to systematically identify processes that are simulated well or poorly, thereby either increasing confidence in RCMs as prognostic tools or indicating model components in need of improvement (Takle et al. 1999).

In the present study results from thirteen RCMs that participated in experiment 1b of the Project to Intercompare Regional Climate Simulations (PIRCS; Gutowski et al. 1998, Takle et al. 1999) are compared to each other and observations. The 60-d simulation period spans 1 June – 31 July 1993, overlapping the peak precipitation episode of the central United States flood (Arritt et al. 1997). An unusually high incidence of heavy precipitation, mesoscale convective systems (MCSs) and low-level jets (LLJs) contributed to this flood event (Kunkel et al. 1994, Arritt et al. 1997, Anderson and Arritt 1998), providing a test of the ability of RCMs to generate climatological features of mesoscale dynamics and precipitation. The relatively flat orography in our central U. S. target region allows us to focus on mesoscale factors other than terrain that contribute to hydroclimate (Giorgi 1990).

In our intercomparison, we emphasize sources of inter-model variability of precipitation, evapotranspiration and horizontal moisture flux. The intercomparison focuses on a $10^{O}$x$10^{O}$ latitude-longitude box ($37^{O}$-$47^{O}$N, $99^{O}$-$89^{O}$W) within the Upper Mississippi River Basin (UMRB). The location of 60-d maximum precipitation is contained within this region in all simulations and in observations. The following section contains a description of data sources and methodology. Results follow in Section 3, and a summary with discussion is given in Section 4.

## 2. Data Sources

### *a. Regional climate models*

Selected characteristics of the thirteen RCMs used in this study are listed Table 1. The continental U.S. and portions of adjacent oceans was included in the domain of each RCM. The nominal node-spacing was 50 km, but varied slightly in each RCM due to different map projections. Additional details of each RCM, as well as the PIRCS experimental design, are reported in Takle et al. (1999). The RCMs examined herein include six models developed outside the U.S. (DARLAM, ETH, HIRHAM, PROMES, SweCLIM), two adaptations of the NCAR-MM5 model (MM5-ANL, MM5-BATS), and two spectral models (NCEP RSM, Scripps RSM). Initial and boundary data for three RCMs (ETH, PROMES, SweCLIM-ECMWF) were generated from the ECMWF reanalysis. All other RCMs used the NCEP/NCAR reanalysis for their boundary data. Note that the NCEP RSM and Scripps RSM utilize information from the reanalysis over the inner domain as well as near the lateral boundaries through domain nesting (Juang and Kanamitsu 1994, Juang et al. 1997, Juang and Hong 2001). They are different only in convective parameterization scheme. Each simulation ran continuously from initialization on 1 June 1993 with lateral boundaries updated at 6-h intervals.

Diagnostic quantities were computed on the native lattice of each RCM before interpolation to a common $0.5^{O}$x$0.5^{O}$ latitude-longitude grid, which is approximately the nominal node spacing of the RCMs. We used a single pass Barnes scheme (Barnes 1964) with e-folding distance set to $0.5^{O}$ in order to remove signals less than twice the analysis grid spacing.

### *b. Observed precipitation*

Precipitation observations used in this study include station hourly precipitation, gridded hourly precipitation (Higgins et al. 1996), and gridded monthly precipitation (Legates and Willmot 1990). We derived station hourly precipitation form the hourly precipitation data (HPD) archive at the National Climatic Data Center (NCDC). Most station reports in the HPD had precision of 2.54 mm, but some reported precipitation to 0.254 mm. For consistency, we truncated the latter to 2.54 mm. Quality control procedures at NCDC removed stations that consistently failed to report both temperature and precipitation. We applied additional selection criteria, removing station records with gaps ≥ 24 consecutive hours. The data set used in this analysis contains 242 stations within the UMRB box. Domain average precipitation was the arithmetic mean of precipitation at stations within the UMRB box.

### *c. Diagnostic quantities*

#### 1) WATER VAPOR CONSERVATION EQUATION

Rasmusson (1968) and Peixoto and Oort (1992) have derived an area-average water vapor conservation equation:

$$S = (E + C) - P \qquad (1)$$

where S is the atmospheric water vapor storage, E is evapotranspiration rate, P is precipitation rate and C is convergence of vertically integrated atmospheric water vapor flux. The terms S and C are:

$$S = \frac{\partial}{\partial t}\int_{P_T}^{P_S} q \frac{dp}{g} \qquad (2)$$

$$C = -\nabla \bullet \vec{Q} \qquad (3)$$

where g is the gravitation constant, q is specific humidity, p is pressure, and integration boundaries $P_S$ and $P_T$ are pressure at the surface and top of the atmosphere, respectively. The vertically integrated atmospheric water vapor flux, $\vec{Q}$, is:

$$\vec{Q} = \int_{P_T}^{P_S} \vec{v} q \frac{dp}{g} \qquad (4)$$

where $\vec{v}$ is the two-dimensional velocity vector. The right hand side of (1) represents processes that can change the atmospheric water vapor content in a unit column. In this formalism, conversion to and from suspended liquid water and ice is neglected.

We applied the water vapor conservation equation (1) to the UMRB box for the 60-d period of the PIRCS simulations. Domain averages were the arithmetic mean of water vapor conservation components at each grid point within the UMRB box. Standard PIRCS output included 3-h accumulation of precipitation and surface latent heat flux, so that P and E in (1) were specified completely by dividing 3-h accumulation by 3-h and averaging over all 3-h periods. Output from most PIRCS RCMs included instantaneous precipitable water every 3-h, but for those that did not we computed instantaneous precipitable water every 6-h from instantaneous values of q and p. The difference of precipitable water between successive 3- or 6-h periods divided by the respective time period was the precipitable water tendency. The 60-d storage (S) was the average of precipitable water tendency taken over each 3- or 6-h interval.

Estimates of water vapor convergence in the central U. S. are sensitive to the frequency and spatial density of wind reports (Berbery and Rasmusson 1999) due to nocturnal acceleration of the low-level wind field over this region (Rasmusson 1968, Berbery and Rasmusson 1999). Since horizontal node spacing of the RCM output is approximately 50 km x 50 km, horizontal resolution should not be a large source of error in convergence estimates. However, PIRCS models archived wind components 4 times per day, which is half the frequency recommended by Berbery and Rasmusson (1999).

Equation (3) may be reformulated by use of Gauss' theorem as:

$$C^* = \oint \vec{Q} \bullet \hat{n}\, d\gamma \qquad (5)$$

where $n$ is the unit vector normal to the perimeter and $\gamma$ is a unit length along the perimeter. We computed the line integral along the perimeter of the UMRB box of the 60-d average of vertically integrated water vapor flux. The error, $\Delta$, of C* was:

$$\Delta = C - C^* \qquad (6)$$

where C is computed as the residual of the water vapor conservation equation (1). Typical values of $\Delta$ were less than 30% of the magnitude of C*. This is consistent with accuracy estimates for observed water vapor convergence in this region (Gutowski et al. 1997). In a few RCMs $\Delta$ was as large as twice the magnitude of C (DARLAM, MM5-ANL, MM5-BATS, RegCM2). Because of this disparity and since model P, E, and S are well represented in model output, we examined C as a residual rather than C*.

2) WATER VAPOR FLUX

The unique nocturnal maximum of summertime precipitation in the U. S. Midwest (Wallace 1974) temporally separates the daily maxima of P and E. This diurnal pattern, coupled with the nocturnal maximum of LLJs, raises questions about the diurnal cycle of water vapor flux in this region. To account for sampling errors discussed in the previous section, we applied an adjustment to 60-d averages of water vapor influx and efflux. The total influx, $F_{in}$, (or efflux, $F_{out}$) of water vapor was the line integral along the perimeter of the UMRB box for which the 60-d average of $\bar{Q}$ was directed inward (or outward). We adjusted $F_{in}$ and $F_{out}$ as follows:

$$F'_{in} = F_{in} + 0.5\Delta \qquad (7)$$

$$F'_{out} = F_{out} - 0.5\Delta \qquad (8)$$

3) RECYCLING RATIO

Estimates of water cycling in the central U. S. indicate that a small fraction of this region's precipitation originates as evaporated water vapor from within the region itself (Brubaker et al. 1993, Trenberth 1999). This characteristic is used as a gross diagnostic of hydroclimate in the PIRCS RCMs. A common quantification of water cycling is the two-dimensional recycling ratio derived by Brubaker et al. (1993), which has the form:

$$\rho = \frac{E'A}{E'A + 2F_{in}} \qquad (9)$$

where E' is area average evapotranspiration, A is area, and $F_{in}$ is water vapor influx. We computed ρ for each RCM using 60-d averages of E' and $F_{in}$. We computed ρ with $F'_{in}$ substituted for $F_{in}$ and found the difference to be inconsequential. The two-dimensional recycling ratio (for a complete review of recycling models see Burde and Zangvil 2001) was formulated for linearly varying fields under the assumption of a well-mixed atmosphere in steady state. If these assumptions were strictly met, the fraction of precipitation from evaporated water vapor within the domain would be exactly quantified. In the central U. S. LLJs, transient synoptic scale low-pressure systems, spatial heterogeneity of P and E, and temporal coherence between LLJs and precipitation are a few of many conditions that may violate these assumptions (Trenberth 1999, Burde and Zangvil 2001). Therefore, we suggest a cautious interpretation, following Trenberth (1999), in which $\rho$ is considered an index, rather than an exact measure, of recycling.

## 3. Results

### *a. Precipitation*

Observed accumulated precipitation for June-July 1993 as estimated using data from an archive initiated by Legates and Willmot (1990) exceeds 400 mm over Iowa, north central and northeastern Kansas, northern Missouri, southeast Nebraska, and southwest Minnesota (Figure 1a). Maxima exceeding 550 mm are located in north-central Kansas and central Iowa. The spatial pattern closely resembles a smoothed contour analysis of rain gauge data for June-August 1993 (Kunkel et al. 1994).

Precipitation fields of the RCMs have similar characteristics. RCM precipitation averaged over all models exceeds 300 mm in an area covering Iowa, southeast Minnesota, and western Wisconsin (Figure 1b). The area of maximum precipitation in the RCM composite is located northeast of the observed maxima. This characteristic reflects a common error in spatial location of simulated 60-d precipitation that occurs in all but three RCMs (plots for each RCM are available online at www.pircs.iastate.edu/hydrology/precipitation.html). The composite precipitation is much less than observed precipitation, due to variability in the exact location of heaviest rainfall in the RCMs. Large areas of precipitation exceeding 500 mm are apparent in all but four RCMs, although having less coverage than in the observations.

*b. RCM 60-d hydrology components*

In the climatological mean, summertime E exceeds P in the central U. S. (Roads et al. 1994, Kunkel 1990, Gutowski et al. 1997). The positive P-E of June-July 1993 was a large deviation from this climatological norm. Estimated P-E in May-June-July 1993 from global analyses or reanalyses was 2-3 mm $d^{-1}$ (Trenberth and Guillemot 1996, Gutowski et al. 1997). Positive P-E was produced in every RCM simulation, although values exceeded 2 mm $d^{-1}$ only in DARLAM and PROMES (Table 2). Nine RCMs (ClimRAMS, ETH, HIRHAM, MM5-ANL, MM5-BATS, NCEP RSM, SweCLIM-ECMWF, SweCLIM-NCEP, RegCM2) produced P-E within the range 0.5 mm $d^{-1}$ to 1.5 mm $d^{-1}$. The overall tendency to understate P-E is due to low bias of P in ten RCMs (ClimRAMS, CRCM, ETH, HIRHAM, MM5-ANL, MM5-BATS, PROMES, SweCLIM-ECMWF, SweCLIM-NCEP, RegCM2). Only DARLAM produced P greater than observed.

Precipitation rate depends on water vapor supply at the RCM boundaries and on precipitation processes internal to the RCMs. Sensitivity of P to the initial soil water content has been demonstrated in many RCM simulations of June-July 1993 (Paegle et al. 1996, Giorgi et al. 1996, Seth and Giorgi 1998, Bosilovich and Sun 1999, Hong and Leetmaa 1999, Hong and Pan 2000). Thus, different sources of initial soil water content (see sec. 2.a) may be related to differences of model P. It is also likely that some differences of model P are attributable to differences in model lateral boundary placement and methods for assimilating lateral boundary data (Seth and Giorgi 1998, Hong and Pan 2000). Lateral boundary nudging-zone width and dynamic constraints are unique to each RCM, and it is impossible to isolate the effect of these differences. Furthermore, the NCEP RSM and Scripps RSM have a unique nesting strategy of a domain nesting in physical space as well as a spectral nesting in spectral space (Juang and Hong 2001).

It is possible to examine differences due to lateral boundary data source. In order to do so, we have divided the models into two subgroups: one contained three RCMs (ETH, SweCLIM-ECMWF, PROMES) that were provided boundary conditions from the ECMWF reanalysis, and the other contained the remaining ten RCMs (ClimRAMS, CRCM, DARLAM, ETH, HIRHAM, MM5-ANL, MM5-BATS, NCEP RSM, SweCLIM-NCEP, RegCM2) that were given boundary conditions from the NCEP/NCAR reanalysis. The range of P of the ECMWF group is contained within the range of P of the NCEP/NCAR group. This suggests that P is more sensitive to RCM architecture than to boundary data source for this collection of RCMs. North America is a data-rich region compared with much of the rest of the globe, so that differences between ECMWF and NCEP-NCAR reanalyses should be relatively small; thus, our result may not apply to other regions.

One component of RCMs that contributes directly to P is the convective parameterization scheme. The convective portion of model precipitation varies greatly between the simulations, ranging from 97% to 39% (Table 3), and large variability occurs between RCMs that use similar convective parameterization schemes, e.g. SweCLIM, PROMES and CRCM. Thus, the influence of convective parameterization on P is RCM-dependent rather than a source of systematic inter-model variability for this collection of RCMs.

Observed E is difficult to ascertain, since it is measured in few locations. Trenberth and Guillemot (1996) have estimated that E during May-June-July 1993 was ~4 mm $d^{-1}$, which is nearly equal to estimates of its climatological value (Roads et al.

1994, Berbery and Rasmusson 1999, Gutowski et al. 1997). Kunkel et al. (1994) concluded that potential evapotranspiration during June-July 1993 was slightly less than its climatological value due to enhanced cloudiness. Thus, a climatological value for E may be an appropriate estimate for June-July 1993. The RCM-average E is 3.9 mm $d^{-1}$. Ten RCMs (ClimRAMS, DARLAM, ETH, HIRHAM, MM5-ANL, MM5-BATS, SweCLIM-ECMWF, SweCLIM-NCEP, RegCM2) produce E within 15% of the RCM-average (Table 2). It is suspected that relatively low E in PROMES is directly related to unrealistically low soil water content in the initial conditions. Other extreme values of E are produced by models having relatively high insolation (NCEP RSM, Scripps RSM, CRCM) despite deriving initial soil conditions from the same source (NCEP/NCAR reanalysis), suggesting radiation parameterization schemes (including cloud effects) are a source of uncertainty in RCM simulations of regional hydrology.

*c. Water cycling*

Climatological estimates of summertime $\rho$ in the central U. S. range from 0.15-0.25 (Brubaker et al. 1993, Eltahir and Bras 1996, Trenberth 1999), reflecting the strong low-level water vapor transport that characterizes this region's summertime hydrology. Dirmeyer and Brubaker (1999) have estimated that $\rho$ in the central U. S. during June-July 1993 was within the range 0.05-0.10. The decrease from its climatological value is due to intensified low-level moisture flux (Trenberth and Guillemot 1996). All RCMs produce $\rho$ within the estimated observed range, except PROMES for which $\rho$ is less than the minimum of the estimated observed range (Table 2). The low value of $\rho$ in PROMES is caused by low E, which occurs despite its relatively high insolation. The agreement between the range of $\rho$ in the RCMs and observations further suggests the collective dry bias is due to internal RCM precipitation processes rather than difference in water vapor supply.

*d. Daily cycle of hydrological components*

While examining the daily cycle of terms of the water vapor conservation equation (1) we found that daily cycles of C and P exhibited nocturnal maxima in most but not all RCMs. We formed two subgroups based on this distinction. Group A is composed of the 9 RCMs (DARLAM, ETH, MM5-ANL, MM5-BATS, NCEP RSM, SweCLIM-ECMWF, SweCLIM-NCEP, RegCM2, Scripps RSM) for which daily cycles of P and C both contained a nocturnal peak. The remaining four RCMs (ClimRAMS, CRCM, HIRHAM, PROMES) formed group B.

In the composite daily cycle of group A (Figure 2a), P and C simultaneously increase gradually during 1030-2230 LST, remain nearly constant during 2230-0430 LST, and decrease rapidly during 0430-1030 LST. (The daily cycle for each RCM may be viewed at www.pircs.iastate.edu/hydrology/daily/watercycle.html.) The broad maximum of composite P between 2230 and 0430 LST is caused by differences among RCMs of group A of the timing of maximum P; some RCMs have a sharp peak at 0130 LST compared to other RCMs in which maximum P is more persistent, lasting from 2230 to 0430 LST. A sharp decrease of P immediately follows maximum P in all RCMs, and although the rate of decrease varies between 1.0 and 3.0 mm $d^{-1}$, minimum P is achieved in all RCMs of group A between 0730 and 1030 LST. Similar trends are found for C. As a result, inter-model spread of P and C is large, and both the time and magnitude of maximums of composite P and C are not representative of the RCMs. However, the composite contains a close correspondence between trends in P and C that is also a characteristic of each RCM.

The composite daily cycle for group B shows less correspondence between trends of P and C. Maximum composite C occurs at 1630 LST, prior to a broad maximum of composite P that occurs between 2030 and 0130 LST. The time of maximum C in each member of group B occurs either in the afternoon (HIRHAM, ClimRAMS, and PROMES) or overnight (CRCM), unlike group A in which all members had a nocturnal maximum of C. Thus, the range of C is very large, providing little confidence that the time and magnitude of maximum composite C is representative. Further, since all models of group B have a daytime maximum of P, the relation between trends of P and C varies among the RCMs. In fact, maximum P lags maximum C in ClimRAMS and PROMES but leads maximum C in CRCM and HIRHAM. The essential characteristics of daily cycles of P and C for each RCM that is reflected in the composite are that P and C reach maximums at different times and P is maximum during the afternoon.

The composites illustrate that climatological precipitation in this collection of RCMs is sensitive to different model processes. The simultaneous maxima of P and C in RCMs of group A indicate a precipitation response to the grid-scale circulation. As discussed in section 3a, the partition of precipitation into convective and stable parts fluctuates greatly among the RCMs. Yet, daily cycles of convective precipitation for members of group A contain similar trends that are reflected in the daily cycle of composite convective precipitation (Figure 3a). Composite convective precipitation is largest at 2230 LST, which is more than an hour after sunset, and remains relatively high through 0430 LST. (The relatively large value of composite convective precipitation at 1630 LST is due to an outlier.) The spread of individual RCM convective precipitation about composite convective precipitation is large in the evening, owing to disparity in fraction of stable and convective precipitation among the RCMs rather than shifting in the time of maximum convective precipitation. Thus, large-scale moisture convergence may be more important than destabilization by insolation in sustaining sub-grid precipitation in these RCMs.

Composite stable precipitation for group A also reaches a maximum overnight. It is noteworthy that composite stable precipitation and convective precipitation contribute about equally to total composite precipitation during 0430 to 1030 LST. In fact, in MM5-ANL, MM5-BATS, NCEP RSM, and SweCLIM-NCEP stable precipitation contributes >50% of total precipitation during this period. Since convective precipitation increases during the afternoon and evening prior to maximum stable precipitation in many of the RCMs of group A, one might hypothesize that a feedback mechanism between the grid-scale and sub-grid moisture fluxes might be at play. One plausible scenario for the relatively large fraction of stable precipitation in some RCMs is that grid-scale relative humidity was increased by detrainment of water vapor from the convective scheme, thereby increasing the likelihood that stable precipitation would develop as grid-scale moistening and lifting caused the relative humidity to exceed the specified threshold for stable precipitation. This mechanism cannot be verified with the PIRCS database, but would constitute a sort of upscale growth of RCM precipitation processes that would be qualitatively consistent with observed precipitation processes in MCCs (McAnnelly and Cotton 1989) if it were confirmed.

The daily cycle of composite convective precipitation of group B shows a very different trend from that of group A (Figure 3b). Maximum composite convective precipitation is reached at 1630 LST and sustained through 2230 LST, although only ClimRAMS peaks after 2030 LST. Since maximum convective precipitation in these RCMs occurs during daytime and is not clearly related to the time of maximum convergence, the results suggest a greater sensitivity to destabilization due to daytime heating by insolation than large-scale convergence in these RCMs.

Similar to composite stable precipitation for group A, composite stable precipitation for group B is largest at night (figure 3b). Despite the disparity among the RCMs of group B in timing of maximum C, a positive value for C is evident overnight in all RCMs of group B except HIRHAM. Thus, stable precipitation likely develops in response to nocturnal convergence in these RCMs, though the response is not nearly as large as in group A.

Because the results suggest that a factor contributing to inter-model variability of precipitation is inter-model variability of convergence, daily cycles of composite $F'_{in}$ and $F'_{out}$ for group A and for HIRHAM and PROMES are plotted in figure 4 (moisture flux fields were unavailable for ClimRAMS, CRCM, and MM5-ANL). The daily cycle of both composites have maximum $F'_{in}$ at 00 LST. This is the time of maximum LLJ frequency in all RCM output (not shown). The time of maximum LLJ frequency derived from hourly reports from the NOAA Profiler Network is slightly later than 00 LST (Arritt et al. 1997). Much of this disparity between RCM output and observations can be attributed to the lower sampling frequency of the RCM output than for the profilers (6-h versus 1-h). While the time of maximum $F'_{in}$ is identical in the two composites, the magnitude of the nocturnal maximum is larger in the composite for group A than for HIRHAM and PROMES, which leads to greater nocturnal convergence in the composite for group A.  Since maximum $F'_{in}$ is related to LLJ frequency, it is also related to the dynamic evolution of LLJs, which contains a substantial divergent (ageostrophic) component (Blackadar 1957, Uccellini and Johnson 1979, Chen and Kpaeyeh 1993). One plausible explanation for the disparity in amplitude of the diurnal cycle of $F'_{in}$ might be the magnitude of horizontal diffusion caused by the computation routines. Greater diffusion would tend to reduce the magnitude of moisture flux, especially the divergent component.  Unfortunately, this hypothesis is untestable with the PIRCS data set.

The presence of afternoon convergence in the composite for HIRHAM and PROMES is difficult to explain.  The daily cycle of composite $F'_{out}$ is very similar for both composites, except at 18 LST when composite $F'_{out}$ for HIRHAM and PROMES is greatly reduced compared to the composite $F'_{out}$ of group A.  This behavior is atypical compared not only to the other RCMs in this collection but also to climatological studies of LLJs and moisture transport (Higgins et al. 1997).

*e.   Daily cycle of observed Precipitation*

The daily cycle of station P contains a single nocturnal maximum during 0130 to 0430 LST and sharp decrease during 0430 to 1330 LST (Figure 5).  These features resemble those of the climatological daily cycle for precipitation in Iowa (Takle 1995), though the amplitude of the daily cycle in 1993 is larger.  The timing of maximum P in group A is much closer to the observed time of maximum P than in group B. This further suggests that the relationship between the large-scale circulation and precipitation is incorrectly simulated by members of group B.  Furthermore, three members of group B (ClimRAMS, HIRHAM, CRCM) rank as the three driest RCMs of this collection, suggesting that incorrectly relating precipitation to the resolvable-scale circulation affects not only the daily cycle of precipitation but also the time-average water balance.

### *f. Three-hour precipitation totals*

#### 1) HISTOGRAM OF 3-H PRECIPIATION TOTALS

Heavy precipitation events were unusually frequent during the peak flood period in late June and early July 1993 (Kunkel et al. 1994). Such events were mesoscale in nature, so that the ability of RCMs to simulate heavy mesoscale precipitation events is important to their ability to add information to large-scale analyses or GCM results. Takle et al. (1999) give evidence that large MCS can be simulated in RCMs even when synoptic forcing is weak. Here, we examine the statistics of 3-h precipitation totals, which is influenced by the integrated effect of heavy precipitation events and MCS. Station precipitation is accumulated over 3-h intervals identical to the archived intervals of the RCMs. The lowest observable precipitation amount (2.54 mm) determined the lowest 3-h total and bin increment in the histograms (see Section 2.b).

In Figure 6 the cumulative frequency of 3-h precipitation totals is expressed as the fraction of 60-d accumulated precipitation for each RCM and the station data. The fraction that is produced by 3-h totals ≤ 12.70 mm is larger in all RCMs than in the station data. The tendency for models to produce too much rain at low precipitation rate is reported at many different time scales in many climate simulations (Giorgi et al. 1996, Kunkel et al. 2001). An explanation for this tendency is that different horizontal scales are represented by precipitation in station data and RCMs. Rain gauge measurements are point observations (Legates and Willmot 1990) while the RCMs in this sample cannot resolve processes having horizontal scale smaller than several times their nominal grid spacing of 50 km. Systematic model error may contribute as indicated by an excessive occurrence of 3-h totals ≤ 12.70 mm (Figure 7) in relatively dry models (Table 2).

A different perspective emerges if we define "heavy 3-h precipitation" as those 3-h rates that contribute the upper 10% of 60-d accumulated precipitation. By this definition heavy 3-h precipitation contribute equally to 60-d accumulated precipitation in the simulations and station data, but the threshold that defines heavy 3-h precipitation may vary. In fact, thresholds under this definition range from 2.54 mm (ClimRAMS) to 53.34 mm (ETH), although 8 of 13 RCMs are within a smaller range of 10.16 mm to 35.56 mm. Thresholds for the simulations are generally lower than for the station data (43.18 mm). This suggests that inadequacy in representing heavy 3-h precipitation in the simulations may have contributed to the collective tendency for the simulations to exhibit dry bias. In support of this inference, three of the four driest RCMs (ClimRAMS, HIRHAM, CRCM) also have relatively low thresholds for heavy 3-h precipitation.

More recent work with ClimRAMS (version 4.3) has introduced the Kain-Fritsch (KF) cumulus parameterization scheme as an alternative to the Kuo schemed used in the PIRCS study. Preliminary RAMS-KF simulations improve the low precipitation biases, especially in the central U. S. (Castro et al. 2001). Results with RAMS-KF will be presented in a future paper on the North American Monsoon Model Intercomparison Project (NAMIP).

2) DAILY CYCLE OF FREQUENCY OF 3-H TOTALS

In the station data different daily cycles of frequency of 3-h totals were found for ranges of 3-h totals of 2.54-5.08, 7.62-10.16, and 12.70-101.60 mm. We defined low, moderate, and high 3-h total categories corresponding to these 3-h total ranges. We applied the same categorical analysis to the simulations. Arguably, 3-h total ranges should be redefined due to the disparity between station and RCM climatologies. However, high rate precipitation is well-defined by 3-h totals ≥ 12.70 mm for all but one simulation. In order to associate meteorological features with the daily cycles, we examined cloud-top characteristics in GOES-8 infrared (IR) imagery.

Daily cycles of accumulated precipitation in each category in the station data have a single peak, but the peak accumulation of high 3-h totals occurs at 09 LST while for moderate and low 3-h totals the peaks occur at 12 and 15 LST, respectively (Figure 7). Widespread high 3-h totals are associated with mature MCSs in GOES-8 IR imagery, whereas low 3-h totals are associated with either the decay of an MCS or (less often) coverage by low-level stratus clouds. Therefore, the time shift of maximum accumulated precipitation is associated with the frequent development and decay of nocturnal MCSs within the UMRB box.

In the simulations daily cycles of accumulation for low, moderate, and high 3-h totals each contian a single peak (except in DARLAM; Figure 8). In the composite of group A, accumulated precipitation from high 3-h totals peaks overnight. (The daily cycle for each simulation may be viewed on the PIRCS webpage at www.pircs.iastate.edu/hydrology/daily/threehourtotals.html.) Accumulated precipitation from moderate and low 3-h totals peaks simultaneously with high 3-h totals. In the composite of group B, maximum accumulation from low and moderate 3-h totals is greater than and *leads* that of high 3-h totals. (Recall that the peak accumulation of low 3-h totals *lagged* that of high 3-h totals in station data.) Thus daily cycles of 3-h totals in the simulations lack a lagged-correlation signal that is consistent with that of observed, recurrent MCSs.

## 4. Summary and Discussion

We have compared output from 13 RCM simulations of June-July 1993 to each other and to observations. Our comparison focused on the atmospheric water cycle over the portion of the Upper Mississippi River Basin (UMRB) where flooding was most intense. Following are the main results of this intercomparison.

- All RCM simulations had P-E>0, but in only DARLAM and PROMES was P-E as large as estimates of observed P-E; the general tendency to understate P-E was caused by low bias of P that occurred in most RCMs.
- Inter-model variability of P was more closely related to internal RCM precipitation processes than to differences in either boundary data or recycling ratio.
- RCM values for E were not consistently greater than or less than estimated values of observed E, but extreme values of RCM E corresponded to extreme values of RCM insolation.
- Nine out of the 13 RCMs produced qualitatively similar daily cycles of P and C, having maximum P and C occur simultaneously at night. This behavior is consistent with observations of precipitation in 1993 and climatological studies of the relationship between observed P and C. The other four RCMs produced daily cycles differing from the first group and each other. Among these simulations, consistent relationships between maximums of P and C were not found, even though maximum P occurred during the afternoon in all four RCMs.
- The low bias of P that occurred in most RCMs is related to overproduction of low 3-h precipitation totals at the expense of high 3-h precipitation totals. RCMs with afternoon maximum of P consistently had greater dry bias and excessive frequency of

low 3-h precipitation totals.

- All RCMs failed to emulate a time lag between maximum accumulation of high 3-h precipitation totals and low 3-h precipitation totals that occurred in station precipitation due to precipitation from MCSs.

A key indicator of the ability of RCMs in this collection to add realistic climatological detail is their ability to simulate the observed nocturnal maximum of P. This feature is absent in the climatology of the NCEP/NCAR reanalysis output (Higgins et al. 1997) that served as driving data for ten of the RCMs. In fact, global climate models and reanalyses, which typically have been run at horizontal node spacing much coarser than the RCM simulations analyzed here, usually do not exhibit the observed nocturnal maximum of precipitation (Ghan et al. 1995, Higgins et al. 1997).

Despite this improvement, additional tests over extended periods are needed to determine the impact of model bias on climate statistics. In particular, the absence of a realistic MCS signal in RCM precipitation suggests that many systems are simulated inaccurately. Although the dynamical scale of such systems may be at or slightly less than the Rossby radius of deformation (Zhang and Fritsch 1987, Cotton et al. 1989), it may be important to simulate such details of climate correctly in order to have confidence in simulations made as forecasts. Mesoscale models that have reproduced many of the dynamical features of MCSs (Zhang and Fritsch 1987, Stensrud and Fritsch 1994) generally use node spacing that is at most one-half of the spacing in this collection of RCMs, suggesting that a first step might be sensitivity analysis of PIRCS1-b results to horizontal node-spacing.

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