A Performance Evaluation of the NGM and RAMS Models for the 29-30 March 1991 Front Range Storm

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ABSTRACT

A PERFORMANCE EVALUATION OF THE NGM AND RAMS MODELS FOR THE 29-30 MARCH 1991 FRONT RANGE STORM

This study investigates the performance of both the Nested Grid Model (NGM) and Colorado State University's Regional Atmospheric Modeling System (RAMS) for the 29-30 March 1991 Front Range storm. Through this investigation, a better understanding of both models efficiency and limitations can be assessed as they simulate the multicomponent nature of winter storms in Colorado's mountainous terrain. In addition, this study has focused much of its attention on initialization procedures in RAMS, and how these procedures affect model output.

In a RAMS control run, which was set-up to mimic NGM's grid structure, it was evident that RAMS was able to outperform NGM for this one storm. In six subsequent RAMS sensitivity simulations, it became clear that it is extremely important for the model to be initialized with realistic topography and surface properties, such as a reasonable soil temperature profile. To date, Colorado Front Range winter storms still pose difficulties in forecasting snowfall amounts and the presence of supercooled liquid water aloft. This RAMS simulation has begun to focus attention, at least from a modeler's perspective, on what parameters are of critical importance in these simulations and what parameters are of lesser significance.

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INTRODUCTION

1.1 Overview

The National Meteorological Center's Nested Grid Model (NGM) is one of the most widely utilized meteorological products since it is used by both forecasters and research scientists. With a 80 km × 80 km horizontal grid spacing, and with 16 vertical levels ranging from the surface to a height of about 18 km, the models resolution (~320 km) is too coarse to resolve many of the mesoscale circulations that develop over the mountainous regions of the western United States. For example, along the Colorado Front Range during winter precipitation events, it is well known that linear shaped snowbands (100 km × 25 km) often develop and are responsible for moderate to heavy snowfall totals in select areas (Rasmussen et al., 1990). With NGM's coarse resolution these mesoscale snowbands remain unresolved and often lead to erroneous forecasts of snowfall amounts.

Accurate prediction of snowfall totals along the Front Range is an ongoing challenge for regional forecasters. Snowfall accumulation has a very high areal variability. For example, the storm of 14-16 January 1987 brought 50 cm of snow to Colorado Springs, 25 cm in the Denver area, and about 10 cm in Fort Collins (Wesley and Pielke, 1990). Snow depth across the area can have as large a gradient as 30 cm to 40 cm over a distance of some 50 km, especially in and adjacent to the foothills. NGM is relatively consistent in predicting deep cyclonic storms moving across the Rockies, however it is of little help to the forecaster who looks to it for determining regional snowfall patterns and amounts.

This thesis is an attempt to better understand the dynamics and microphysics of Front Range snowstorms. In order to accomplish this goal, the Colorado State University's Regional Atmospheric Modeling System (RAMS) has been utilized. RAMS output has

been directly compared with NGM output and then both models output have been checked against observations. It is hoped that through the comparison and contrast of model results that we are able to gain new insight into the importance of realistic topography, initialization with appropriate surface fluxes, and microphysics in these types of models. The 29-30 March 1991 storm was selected as a case study because the NGM greatly over-predicted the extent and amount of snow that fell along the Front Range.

Late in the evening of March 28, this particular storm had all of the traits of a deep cyclonic storm that should have produced widespread and moderate snowfall across the area. At 0000 UTC on 29 March, a 500 mb low was positioned over the Texas panhandle (Figure 1.1). Over the next 12 hours a 500 mb low developed over southwest Colorado, while a 700 mb low formed over northeast New Mexico. In addition, there was a surface high over the upper Great Plains that by 0000 UTC was causing northeasterly flow of colder air into the Front Range. NGM's 24- and 36-hour forecast made at 0000 UTC 29 March, predicted moderate amounts of snowfall (10–20 cm) from northern Wyoming to northern New Mexico (Figures 1.2 and 1.3). Figure 1.4 shows observed total precipitation for the storm across the Colorado region. It is evident that there were select areas which received 2.0 cm or more of water equivalent precipitation, but those regions were of very limited area, and certainly not to the extent and amounts predicted by the NGM.

1.2 Previous Research

There has been considerable research and interest in wintertime snowstorms along the Colorado Front Range over the last 15 years. Boatman and Reinking (1984) synthesized much of the early work on these storms; they also were among the first to draw a clear distinction between deep cyclonic storms that frequent the area, and arctic outbreaks of cold air that moves into the Front Range from the east. In reality, many Front Range snowstorms have characteristics of both deep cyclonic and shallow arctic air masses.

Wesley (1991) distinguished between the blocking of cold air as it tries to move up the eastern slope of the Rockies, and cold air damming. In the former, as the cold air is orographically lifted, it cools even more, creating a meso-high pressure dome at the

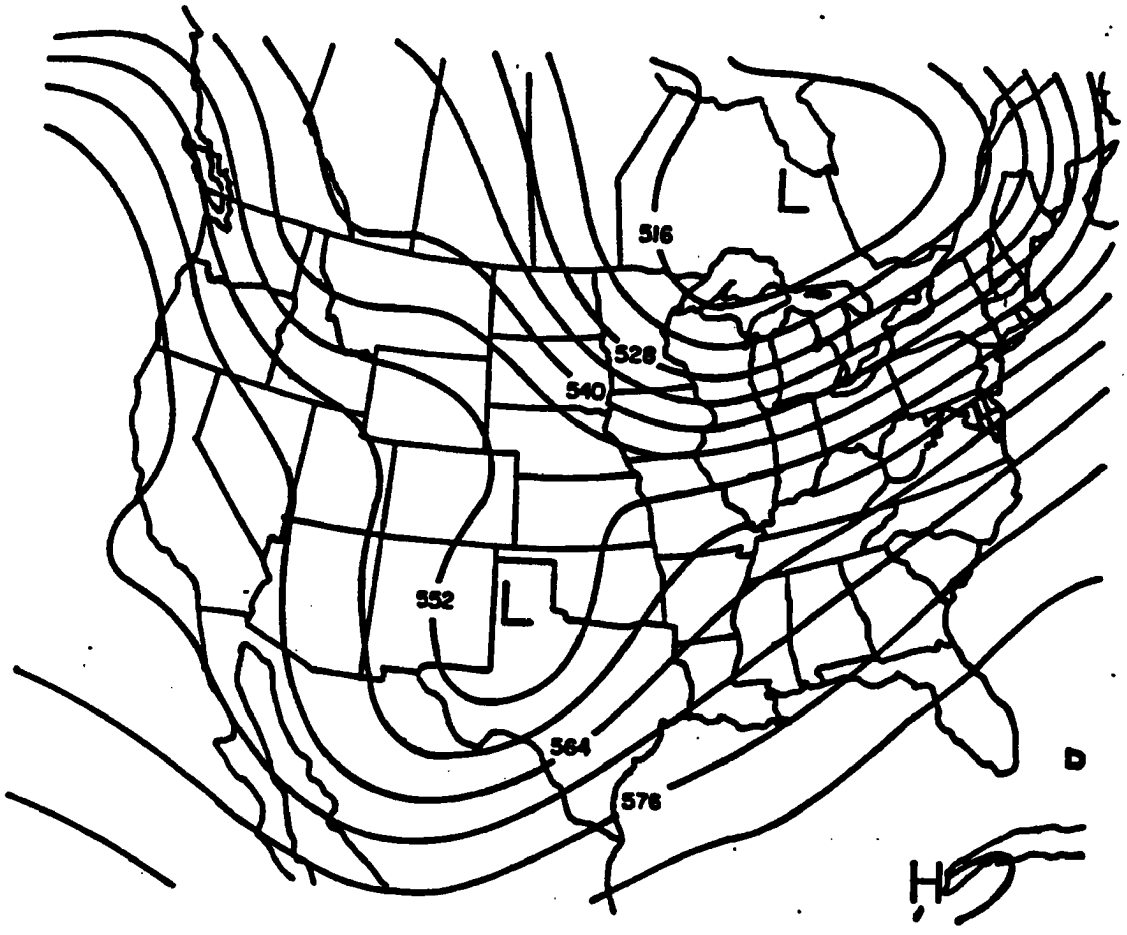
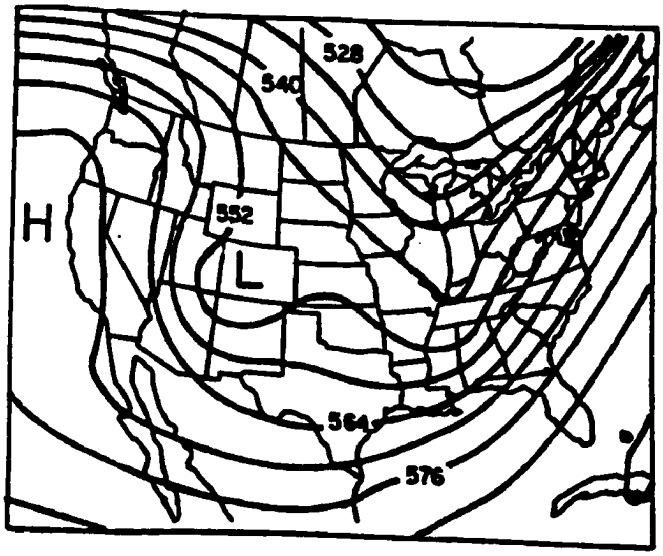


Figure 1.1: 500 mb height field at 0000 UTC 29 March (60 m contour interval).



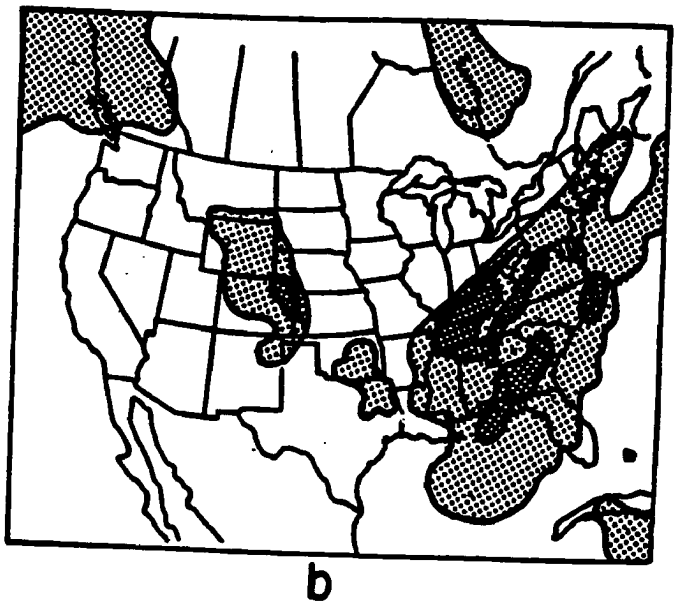
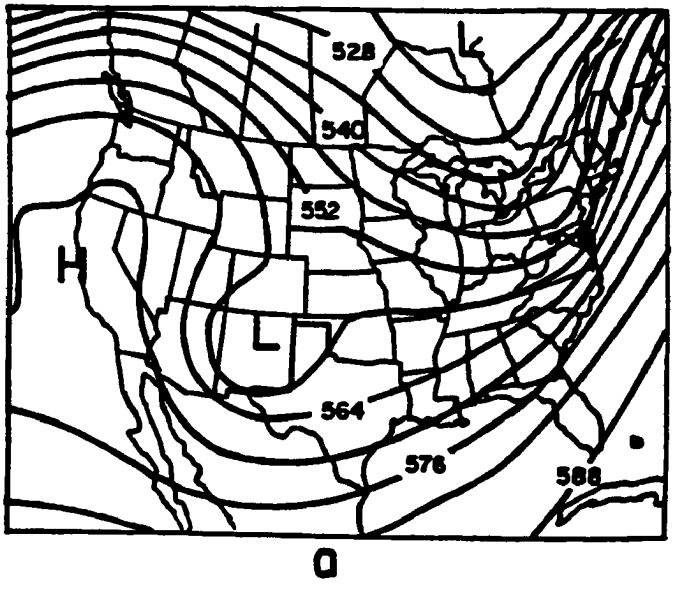


Figure 1.2: NGM 24-hour forecast, valid 0000 UTC 30 March for (a) 500 mb height field (60 m contour interval), and (b) precipitation (light stipple – 0 to 1 inches, dark stipple – 1 to 2 inches).



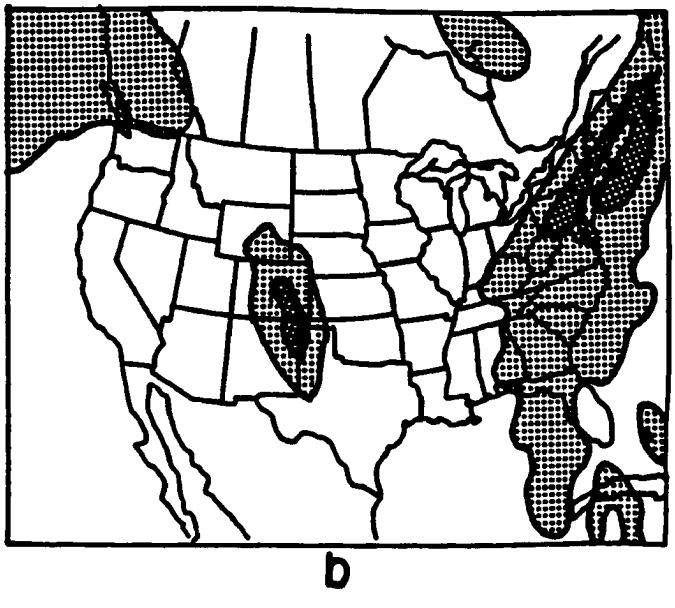


Figure 1.3: NGM 36-hour forecast, valid 1200 UTC 30 March for (a) 500 mb height field (60 m contour interval), and (b) precipitation.

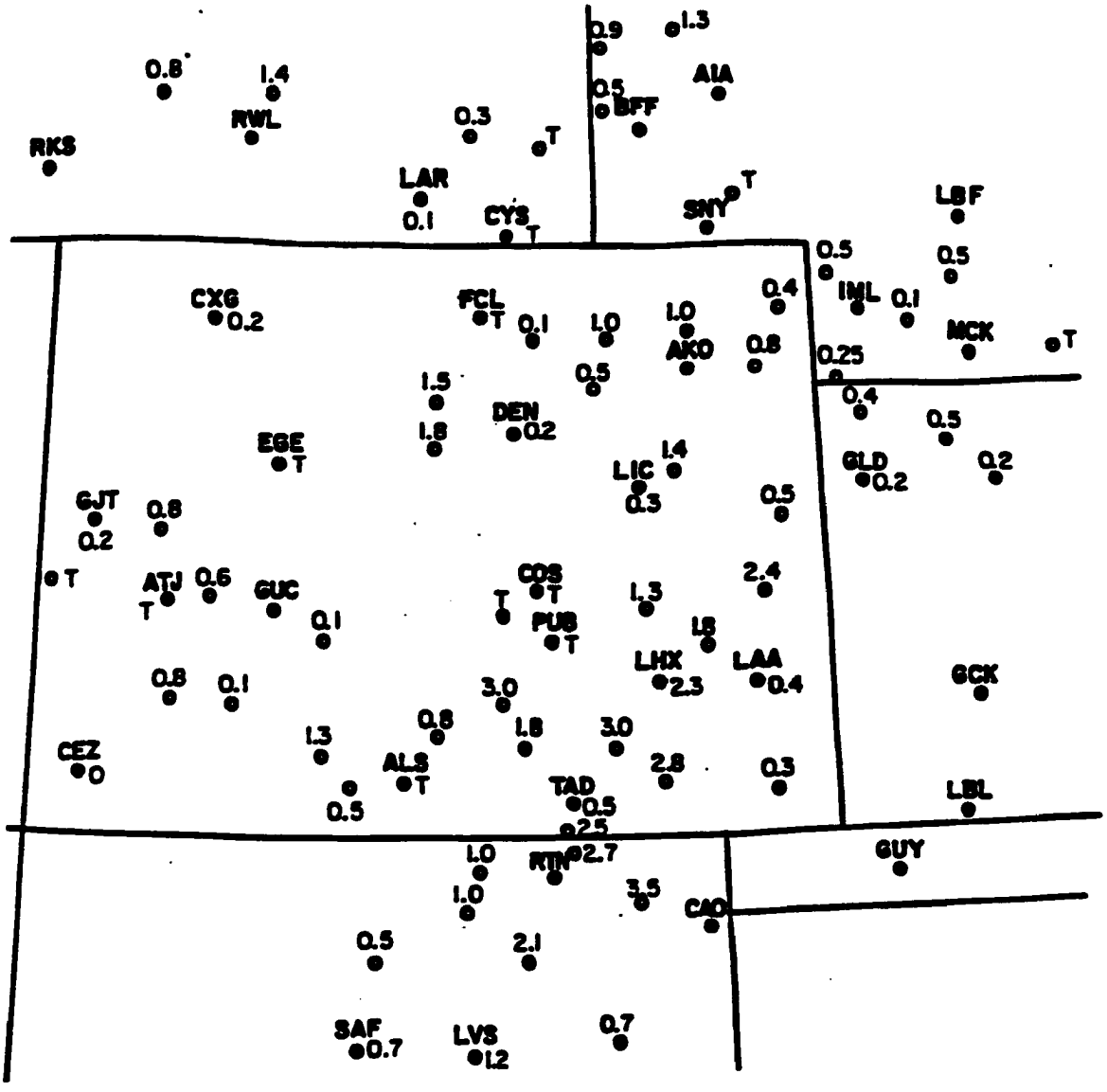


Figure 1.4: Storm totals 29-30 March 1991 (cm water equivalent).

surface which restricts further upslope flow. Damming occurs when easterly flow is forced up and over the eastern edge of a cold dome. This often leads to a mesoscale convergence zone and a subsequent area of moderate to heavy snowfall some 50 to 100 km east of the foothills. Terrain-induced blocking is found in almost all Front Range snowstorms, while damming is found in a much smaller subset of storms.

Blocking occurs in a stability stratified environment when the ratio of the inertial to buoyant force is less than one (Wesley, 1991). The Froude number $fr = U/Nh$ is a measure of that ratio, where U is the flow velocity, L is the barrier height, and N is the Brunt-Väisälä frequency. The Froude number is also related to the ratio of kinetic to potential energy. The stronger the temperature stratification is in lower levels, the more difficult it becomes for low-level flow to ride over the crest of a topographic barrier. Since the Front Range is several hundred kilometers long, and some 2000 m higher than the eastern Plains of Colorado, blocking events are frequent when the winds have an easterly component.

Boatman and Reinking (1984) found that in the case of an arctic outbreak when low-level easterly flow (upslope) is forced upward due to the regional topography, upslope clouds often form. The arctic airmass is often only about 100 mb deep, with the upslope clouds forming at the top of the cold airmass in an inversion zone. A well-known shortcoming of NGM, as reported by Junker et al. (1989), is its inability to accurately predict the evolution of low-level arctic air masses as they move across the high plains toward the Rockies.

Recently, there has been considerable interest in the microphysics and thermodynamic processes taking place within a upslope cloud layer. This is an ideal location for the presence of supercooled liquid water (SLW), and appears to be responsible for much of the in-flight aircraft icing along the Front Range. The National Center for Atmospheric Research's Winter Icing and Storms Project (WISP) is an ongoing research project, dedicated to understanding the problem of aircraft icing, especially the hazards that exist within these upslope clouds. Typical temperatures within this layer range from $-7°C$ to $-14°C$ (Reinking and Boatman, 1986), with liquid water contents from 0.05 gm m^{-3} to

as high as 0.7 gm m^{-3} (Rasmussen et al., 1991). One of the interesting findings of the Rasmussen research group (1991) is that regions within upslope clouds that are low in SLW are high in ice crystal mixing ratios. In addition, using the Clark anelastic mesoscale model, this group found that areas of SLW were organized into bands oriented NW to SE with a 50 km spacing. This is of interest because it appears to be very similar to the snowbands that are commonly observed by local radar (Wesley et al., 1990).

The formation and propagation of these snowbands is crucial to understanding and forecasting snowfall amounts as well as icing potential along the Front Range. With a typical size of 100 km by 25 km (Wesley, 1991), these mesoscale features can only be resolved by high resolution models. Their origins are not yet fully understood, however, it is presently thought that topography plays a prominent role in their development as well as convective instability (Rasmussen et al., 1990), gravity waves, and upper level jet streaks (Powell, 1992).

Low-level upslope flow along the Front Range is vital to local snowstorms because it causes decoupling from the dominate westerly flow, not allowing adiabatically warm air to reach the surface (Wesley, 1991). Secondly, in some cases, a bulge forms along the western edge of the cold pool, directly over the foothills. This forces the westerlies to rise over the bulge, with the subsequent vertical motion leading to increased production of ice crystals, and increased snowfall (Wesley et al., 1990). Thirdly, it is quite common to have multiple cold surges propagate from the east during the life of a storm (Dunn, 1987). Once a cold pool is entrenched along the Front Range, secondary surges often are forced over the cold top, creating the aforementioned mesoscale convergence zone (Wesley and Pielke, 1990). In addition, Wesley (1991) speculates that convection induced by the lifting of the easterlies over the cold pool may be important for precipitation processes.

The decoupling of shallow easterly flow from the westerlies is important because in many respects it extends the breadth of the mountain barrier causing temperatures in the intervening layer to remain colder, causing greater precipitation efficiency by increasing the time scale for crystal growth (Peterson et al. 1989). With a larger concentration of crystals in the middle and upper levels, there is a higher probability that the seeder-feeder

mechanism is in operation. Wesley and Pielke (1989) report that many of the reported crystal types for the 30-31 March 1988 storm were dendrites that were heavily rimed. The middle and upper levels produce plate, columnar, or dendrites which become rimed as they fall through the SLW region of the cold dome.

As noted by Boatman and Reinking (1984), shallow upslope clouds produced by anticyclonic storms located in the northern Great Plains, produce stratus type clouds that contain relatively little moisture, hence these clouds produce little precipitation on their own. However, if moisture is advected in from the west or southwest, then the potential for moderate to heavy snowfall is possible. This is exactly what often happens when there is a 500 mb low over the Four Corners region of the southwest. In this scenario, moisture from the Gulfs of California and Mexico is advected over the southern Rockies. The seeder-feeder precipitation enhancement process (Reinking and Boatman, 1986) comes into play as orographically produced middle and upper level clouds seed the lower levels with seeder crystals. As the ice crystals fall through the low-level cold pool, they are often rimed, indicating the presence of a layer of SLW.

In a damming scenario, as easterly flow rides over the leading edge of the cold pool, snowfall totals and the areal distribution of that snow seem to be related to the moisture content of the inflow (Wesley, 1991). One must also account for any movement of the convergence zone during the life of the storm, as the cold pool increases in depth or if the leading edge propagates eastward. One must also allow for complex crystal trajectories as these particles fall through several layers, each one with a different wind speed and direction. In some cases, the source area can be a large distance from the target area.

The topography of the Front Range not only causes blocking, it also forces air upward, creating local convergence zones. The north-south oriented Rocky Mountains provide the primary barrier to both westerly and easterly flow. However, smaller terrain features such as the Cheyenne Ridge and Palmer Divide play prominent roles in the local production of snowfall (Figure 1.5). These terrain features cause enhanced low-level vertical motion, often resulting in a high correlation between areas of maximum snowfall and terrain features (Wesley and Pielke, 1990).

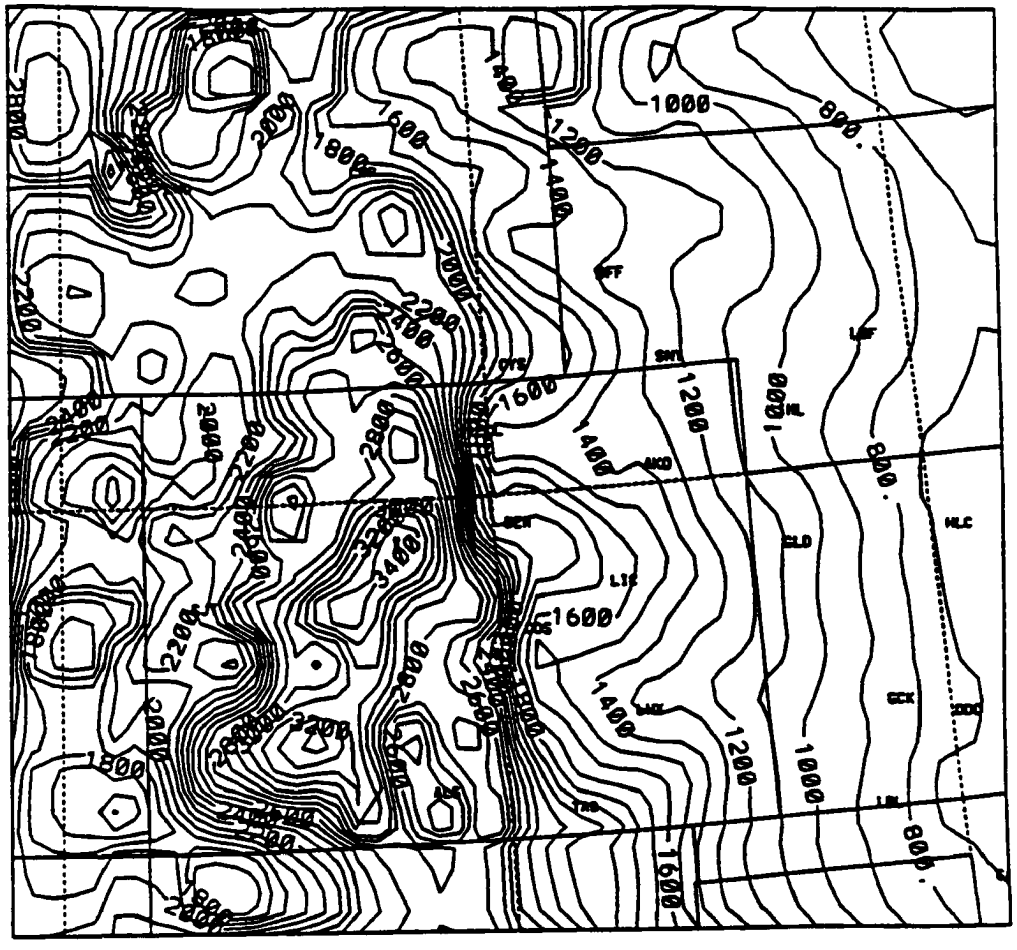


Figure 1.5: Topography of the Colorado region (contours 100 m).

At certain times non-classical mesoscale circulations such as a 'snowbreeze' can also create a localized snowstorm. Johnson et al. (1984) documents a case in April of 1983 where a large snow-face area in the north Front Range was snow-free, but surrounded by snow-covered terrain. Rapid heating of the snow-free area along with a large temperature gradient across the snow-covered/snow-free boundary, allowed for the generation of a low-level wind from the snow-covered ground on to the snow-free ground in the same fashion as a sea breeze. Vertical motions and an abundant supply of moisture caused clouds to form over the snow-free area, with a subsequent snowfall over the previously snow-free area. This snow-breeze effect is only possible in the fall and spring when the thermal forcing is large enough. This type of mesoscale circulation in addition to the mountain-plain circulations do show the multi-faceted nature of events that can induce or enhance local snowstorms.

1.3 Modeling Studies

Abbs and Pielke (1987) simulated two upslope snowstorms with a model that was a precursor to RAMS. They concluded that the effects are strongly dependent on the orientation of the prevailing wind with respect to terrain orientation. Hence, southeast winds that descend from Palmer Divide into Denver often result in little snowfall; northeast winds however, are forced upslope towards Denver. In the process, the airmass becomes cooler often leading to large amounts of snowfall.

Meyers and Cotton (1992) were able to simulate a snowstorm which took place in the Sierra Nevada Range of California. They found that orographic blocking caused an upstream (40 km) cellular structure as ambient flow was forced upward through a deeper layer because the blocked air was acting as an 'equivalent' barrier. Cloud bands formed in the upward portion of these cells. These bands did propagate downstream and had a large effect on precipitation processes. The authors note that the vertical motion upstream of the Sierra Nevada barrier was due to orographic blocking but that the trigger responsible for the cellular band structure was wet potential instability. They also conducted a sensitivity experiment where they took out the Coast Range. The Coast Range lies some 120 km

west (generally upstream) of the Sierra Nevada. Their results indicate that a seeder-feeder production mechanism was triggered by the Coast Range.

Lee at al. (1988a) modeled flow over a cold pool which exits on the lee side of a mountain barrier. In this case they found that the cold pool can modify mountain waves very significantly, but is dependent on the depth of the cold pool. Their results indicated that the top of the cold pool acts very much like terrain with similar shape. In the absence of strong surface heating, the cold pool was flushed by the mountain wave only when the cold pool was shallow (less than half of the barrier height).

The Peterson et al. (1991) model simulations and field work in the Yampa Valley of north-central Colorado, reveal that in the case of low-level decoupled flow, a pool of dammed air does not necessarily "decrease the effective height of the barrier, but rather acts as an extension of the mountain for orographic lift purposes." The net effect is an enhancement in the amount of precipitation, as well as an upstream shift in snowfall totals. The authors attribute the enhancement to a longer trajectory in which the hydrometeors have more time to grow.

Modeling sensitivity experiments conducted by Cotton et al. (1986) on aggregation rates in orographic snowstorms suggest that small terrain features "upstream of major mountain barriers can significantly alter the ice/water budgets of the clouds forming over the main orographic barrier." It should be noted that these simulations were conducted with clouds which had low liquid water content.

The RAMS model has been previously used to study winter precipitation events along the Front Range. Wesley (1991) simulated both a deep cyclonic and a shallow anticyclonic event. He found that topography played a major role in the distribution of snow along the Front Range due to small-scale orographic lifting as well as cold air damming. Wesley (1991) simulated the 30-31 March 1988 storm with varying degrees of success. RAMS had some difficulties in predicting 700 mb flow as well as the strength of the 500 mb cut-off low. As a result, the model-produced temperature fields were 1°C to 3°C too warm along the Front Range. In addition, the simulation was not able to create blocked flow along the foothills/I-25 corridor. The model did, however, accurately produce many of the synoptic-scale features, as well as an enhancement of orographic precipitation. Wesley attributes

some of RAMS difficulty in producing certain flow fields to the model's sensitivity to initial conditions as well as to boundary conditions.

Two other factors that could lead to erroneous output are the difficultly in accurately portraying surface fluxes, especially over a large domain that is for model purposes considered horizontally homogeneous in terms of the surface landscape, and secondly improper terrain representation. High resolution terrain data sets are important for simulations in complex terrain. As an example, Wesley (1991) added a second telescoping grid (22 km grid spacing) after the initial simulation. In this second grid domain, he used a higher resolution terrain data set (30 seconds vs. 10 minutes for the coarse grid), which resulted in the model being able to predict more representative precipitation values in the mountains of Colorado.

Wesley's (1991) two-dimensional simulations indicated the effect of a cold pool on the vertical motion of the westerlies as they cross the barrier. In the case of a deep cold pool (2.8 km MSL), a strong gravity wave signature was apparent in the westerlies overlying the cold pool. Within the cold pool itself, there was little vertical motion. When the cold pool was removed from the simulation, the gravity wave signature was almost totally absent, and vertical velocities had almost increased two-fold. These simulations seem to indicate that there is an increase in areal extent of the region of ascent (although of smaller velocity) at the western edge of the cold pool, as contrasted with when the cold pool is very shallow or absent. The net effect of the cold pool is not only to retard adiabatically warm winds from reaching the surface but precipitation is enhanced due to increased upward motion above the cold pool.

1.4 Summary

These modeling studies reveal the multi-component nature of orographically influenced Front Range snowstorms. Figure 1.6 is a flow chart that attempts to capture the salient components. Such a chart summarizes many of the factors that a forecaster must consider and evaluate before issuing a snowfall or icing forecast.

The first factor is what I call synoptic favorability. What is meant by this is that there must be a source of moisture, as well as cold surface temperatures. Moderate to

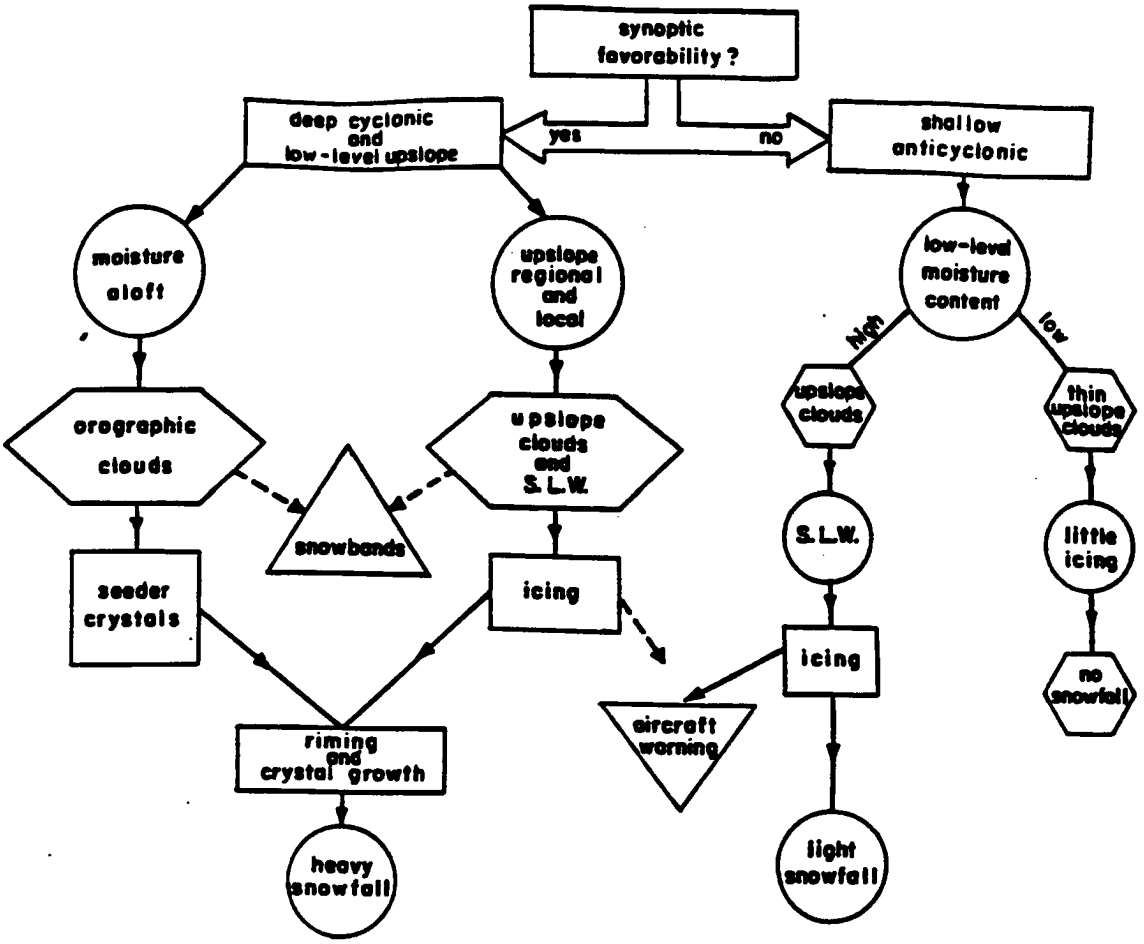


Figure 1.6: Flow chart showing multi-component nature of Front Range snowstorms that lead to aircraft icing and either light or heavy amounts of snowfall.

heavy snowfall across the Front Range are associated with middle and upper level moisture advection from the southwest or west. Cold surface temperatures result from upslope flow. The forecaster must first ask if this is a deep cyclonic storm, with abundant moisture, or a shallow anticyclonic storm with little moisture.

If the synoptic pattern is conducive to a deep cyclonic storm, then orographic clouds will probably form in the moist flow aloft. There is also the possibility of snowband formation. Will the snowbands remain stationary or propagate? Once a cold pool forms, will a convergence zone east of the foothills/I-25 corridor form?

Near terrain features such as Cheyenne Ridge and Palmer Divide, the forecaster must ask if the surface winds are downslope (northeast into Fort Collins) or upslope (northeast into Denver)? In the former case, no snow may fall at all.

A forecaster must also have access to data that indicates the presence of SLW at the top of the cold pool inversion. If SLW is present in abundance, what kind of icing hazard exists, as well as what increase in snowfall could occur due to riming and crystal growth? One must also be aware that very cold surface temperatures may not be representative of the overlying airmass. It is generally thought that temperatures below $-20°C$ are too cold for any sustainable snowfall. In this shallow cold pool, however, it has been known to snow heavily when the surface temperatures have been in the $-25°C$ to $-30°C$ range (Wesley, 1991).

Since these components change with time, and some, such as surface winds can change very rapidly, an accurate snowfall forecast must be constantly updated. Herein lies the challenge.

Chapter 2

THE NGM AND RAMS MODELS

2.1 Introduction

With a large increase in computer speed and storage capabilities, numerical models are quickly becoming the cornerstone for both forecasters and researchers. One would assume that higher resolution could be achieved by simply reducing the grid spacing within the model. To some degree this is true, however, Weygandt and Seaman (1988) show that when the grid resolution increases, it is equally important to have microphysics and initialization parameterizations that are appropriate at that particular scale. In short, if the grid resolution increases without modifications to model physics, which in some cases are grid dependent, then the overall model accuracy is compromised. Ramage (1982) notes that improvements in the prediction of large-scale features did not necessarily bring improvements in the prediction of precipitation.

Using the Penn State/NCAR mesoscale model, Weygandt and Seaman (1988) investigated cyclogenesis in the eastern United States. The purpose of their study was to see how grid spacing affected the quality of the prediction. They set-up the model with three grids. The coarse grid had a 160 km spacing, the second grid 80 km, and the fine grid 26.7 km. Simulating several cold air damming events along the western side of the Appalachian Mountains, they found that after 72 hours of simulation, mesoscale features were indeed better resolved on the finer grid, at least for phenomena that were related to geographic features (i.e., stationary). Next they examined the model-resolved propagating mesolows. They found that pressure fields were predicted more accurately as grid spacing decreased, but position errors did not show the same pattern. In some cases, position errors increased as grid size decreased. They attribute these errors to the fact that most

propagating phenomena enter the domain of the fine grid from a much lower resolution coarse grid. There is also the possibility of distortion at the interface between the two grids.

The need for more studies of this nature is evident. Each model has its own strengths and weaknesses that must be recognized. It is easy to forget that numerical models are not exact representations of the real atmosphere, although the goal of the modeler is to create a 'model atmosphere' where research can be conducted without leaving the office. These models are only as good as the model equations and parameterizations allow them to be. In reality, models should be constantly evolving as new parameterizations and ideas are incorporated into them.

To date, most high resolution mesoscale models are used primarily as research tools, but the day when they are used to make forecasts is on the horizon. Their continued modification and upgrading can only be accomplished if we know how they perform with respect to the real atmosphere and other models. This chapter is an overview of both the NGM and RAMS models, their set-up, and some of their deficiencies and strengths.

2.2 The Nested Grid Model

The nested grid model (NGM) is a primitive equation grid-point model, containing three nested grids which provide hemispheric coverage (northern hemisphere), the highest resolution being over North America (Tuccillo, 1988). The horizontal grid spacing for the three grids is 320 km, 160 km, and 80 km with 16 stretched vertical levels ranging from 1000 mb to 10 mb. The timesteps for the three grids are 300 seconds, 150 seconds, and 75 seconds, respectively.

The National Meteorological Center (NMC) periodically updates and introduces new procedures and new parameterizations into the NGM. In July of 1986, the radiation scheme, cloud model, surface energy budget, and boundary layer parameterizations were updated (Tuccillo, 1988). The radiation parameterization utilizes both shortwave and longwave radiation, and includes the effects of water vapor, clouds, ozone, and carbon dioxide. This routine is called once per forecast hour.

The interactive cloud model allows clouds to form within a grid volume when the relative humidity reaches 80%. Only three of the 16 vertical layers are allowed to contain clouds at any one time.

Boundary layer parameterization of surface fluxes uses the bulk aerodynamic approach proposed by Louis (1979). These equations are then incorporated into a mixed layer model developed by Phillips (1986). As many as six layers may be considered as mixed layers within the model. Mixed layers responded to an upward flux of heat or moisture (buoyancy) from the surface and mixing by the wind. There is also a stability dependent vertical diffusion equation for momentum. The surface energy budget equation is solved every 300 seconds. Surface parameters are grid dependent with surface roughness, albedo, and moisture availability being the most important. Albedo is increased in those grids which are covered with any amount of snow. Moisture availability represents the parameter with the greatest uncertainty (Tuccillo, 1988). Moisture data is obtained from an annual climatological vegetation index.

Areas covered by snow are determined from satellite surveys. All grid volumes whose areas are at least 50% covered by snow are considered to be fully covered by snow, while those with less than 50% coverage are considered snow-free (Petersen and Hoke, 1989). Snow coverage is updated within the NGM once a week, usually on Tuesdays or Wednesdays. This routine is static in the sense that snow depth and areal coverage do not change during the week-long forecast cycle. This can lead to major forecast errors when there are rapid and large areal changes in snow cover, especially during the Spring and Fall. As an example, Petersen and Hoke (1989) state: "errors in the forecasted surface temperature of nearly 10°C are not uncommon when the actual snow cover and that used by NGM differ."

On 7 November 1990, additional modifications were introduced into NGM as reported by Petersen et al. (1991). Noteworthy modifications include enhanced orography and subsoil temperature. The old topography lacked insufficient resolution for many important features; there were many misrepresentations along mountainous coastlines, as well as differences in the detail and quality of data used for the United States and that used for

Canada and Mexico. The new topography is based on the US Navy 10-minute resolution terrain data set. Filtering produces a field with T80 resolution. Triangular 80 filters the terrain isotropically. Terrain represented in physical space is converted into spectral space where all wave numbers larger than 80 are truncated. It is then transformed back to physical space, where there are 80 wavenumbers per latitude circle. Here there is higher resolution at high latitudes, where the circumference of the latitude is smaller than near the equator. Figure 2.1 shows the new topography. Test runs by Petersen et al. (1990) indicate that NGM now produces more realistic orographic precipitation and a slightly reduced tendency for erroneous lee side cyclogenesis.

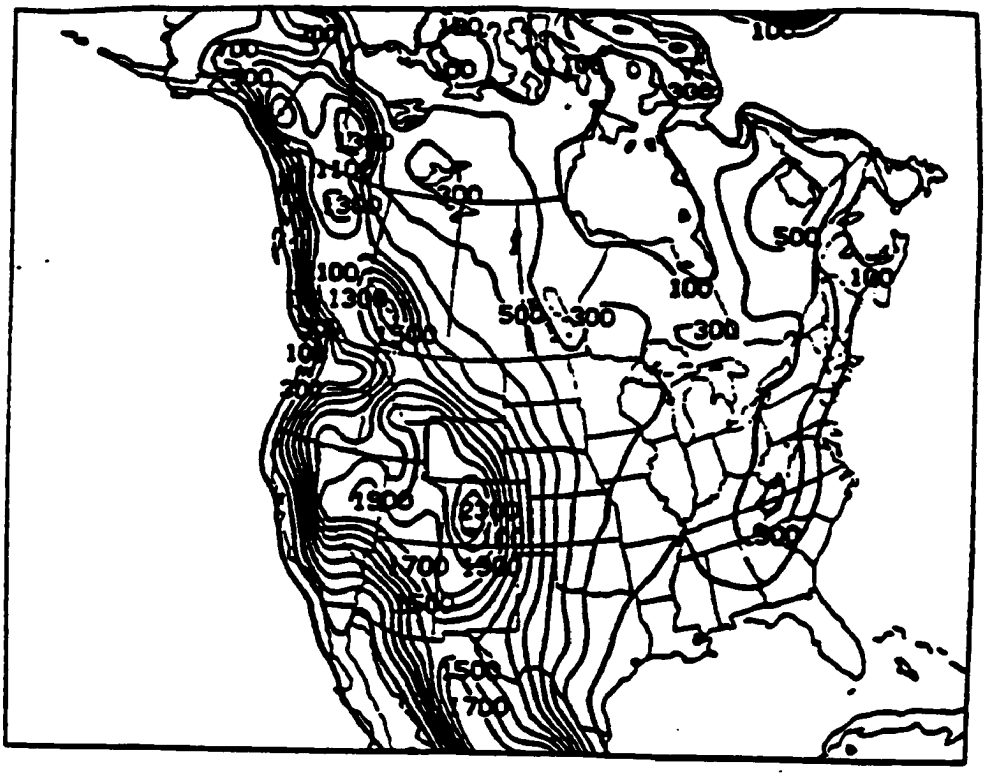


Figure 2.1: NGM 10-minute data set after smoothing.

2.2.1 NGM Performance Evaluation

Junker et al. (1989) reports on the strengths and deficiencies of the model using data from several NMC test simulations and then comparing that output with observations. The NGM has a realistic diurnal temperature cycle, however, the cold bias prevalent over the Rockies is a problem. This cold bias would tend to cause an overprediction in precipitation over much of Colorado as air parcels reach premature saturation.

In a study by Hoke et al. (1985), they were able to show that during a three-month period (November 1984–January 1985), NGM's 48-hour forecast for surface lows had a mean distance error of 322 km. Junker et al. (1989) also found that NGM often overdeepened cyclones coming out of the Rockies. The authors attribute part of this problem to the model's inability to capture shallow cold air blocking/damming events along the lee side of the Rockies. For anticyclones, the mean distance error over a three-month winter period were 261 km for the 24-hour forecast and 334 km after 48 hours.

From the aforementioned studies it is clear that the NGM is somewhat suspect in its 24 to 48-hour forecasts in and near the Rockies during the wintertime. Positioning of lows and highs is crucial to forecasting Front Range snowstorms. For example, if the NGM positions a low on the Colorado-New Mexico state line, but in reality the low is some 250 km to the south, then there is a very real possibility that NE Colorado will receive very little precipitation while the forecast may call for moderate snowfall.

The fact that NGM has real problems with arctic outbreaks, which are important in almost all of the snowstorms along the Front Range, means that it is of limited use in these types of situations. There is always the possibility that NGM's various errors (cold bias, cyclone positioning, arctic outbreaks) somewhat cancel each other, so that the model forecast verifies for a given storm. On the other end of the spectrum, there are times when NGM gives a highly inaccurate forecast. When it comes to snowfall, a badly blown forecast can spell trouble for travelers and those caught in the middle of a 'sudden' snowstorm. With NGM's (and other NMC models) limited usefulness in areas of complex terrain, forecasters are now beginning to use higher resolution models, such as RAMS to forecast and study these snowstorms.

2.3 The Regional Atmospheric Modeling System

The RAMS model (Tremback et al., 1986; Tremback, 1990; Pielke et al., 1992) was born out of the merger of two mesoscale models and a cloud model. The result is a telescoping two-way nested atmospheric model which contains many user-specified options (Pielke et al., 1992). These options allow the model to be used for a range of applications over a wide range of scales. These options include the use of hydrostatic or non-hydrostatic primitive equations, variable or horizontally homogeneous initialization, cumulus parameterization or explicit microphysics, as well as radiation and surface budget parameterizations. All of the simulations in this thesis were conducted using RAMS Version 2C with variable initialization. Table 2.1 summarizes the pertinent model parameters used in these simulations.

Table 2.1: Summary of Model Parameters Used

Primitive equations	Nonhydrostatic
Grid points	70, 62, 17 in x, y, z
Grid spacing	80 km in x, y and variable in z ranging from 200 m to 17250 m
Timestep	90 seconds
Radiation	Chen parameterization, updated every 1200 seconds
Top boundary	Wall
Albedo	0.2
Surface Roughness	0.05 m
Soil Parameterization	Tremback and Kessler, 11 soil levels
Microphysics	Explicit
Initialization	Variable with 12-hour nudging
Diffusion	Deformation

2.3.1 Grid structure

In order to mimic NGM as close as possible, an 80 km by 80 km horizontal grid was set-up. The vertical coordinate used in the model is a terrain-following σ_z system. 17 vertical levels were used in all of the simulations except ST 5 which was a high resolution run with 30 vertical levels. The first model level is actually below soil surface so that

there were 16 levels in the atmosphere, ranging in height from 200 m to 17250 m. Table 2.2 gives NGM and RAMS vertical levels. Figure 2.2 shows the extent of the coarse grid

Table 2.2: NGM and RAMS vertical levels in meters. (NGM heights approximated from standard atmosphere.)

	NGM (m)	RAMS (m)
1	0	200
2	300	594
3	700	1037
4	1150	1582
5	1675	2292
6	2300	3044
7	3050	3830
8	3900	4793
9	4800	5800
10	5900	6800
11	7075	7800
12	8500	8756
13	10250	10201
14	12300	12369
15	15250	14730
16	19000	17250

domain. It is important to have a domain large enough to capture propagating features, especially wintertime disturbances that move on shore from the North Pacific.

2.3.2 Radiation parameterization

The Chen and Cotton (1983) parameterization for both longwave and shortwave radiation was used in every simulation. This scheme allows for the radiative effects of water vapor, cloud water, ozone, and carbon dioxide. It was felt that radiative flux divergences updated every 1200 seconds was sufficient for this type of simulation which lasted a total of 36 hours. The radiation and microphysics modules are the most computationally costly parts of the model, hence a realistic but efficient timestep is required. Shortwave radiation varies with longitude across the width of the domain, and is adjusted to account for sloping topography (Cram, 1990).

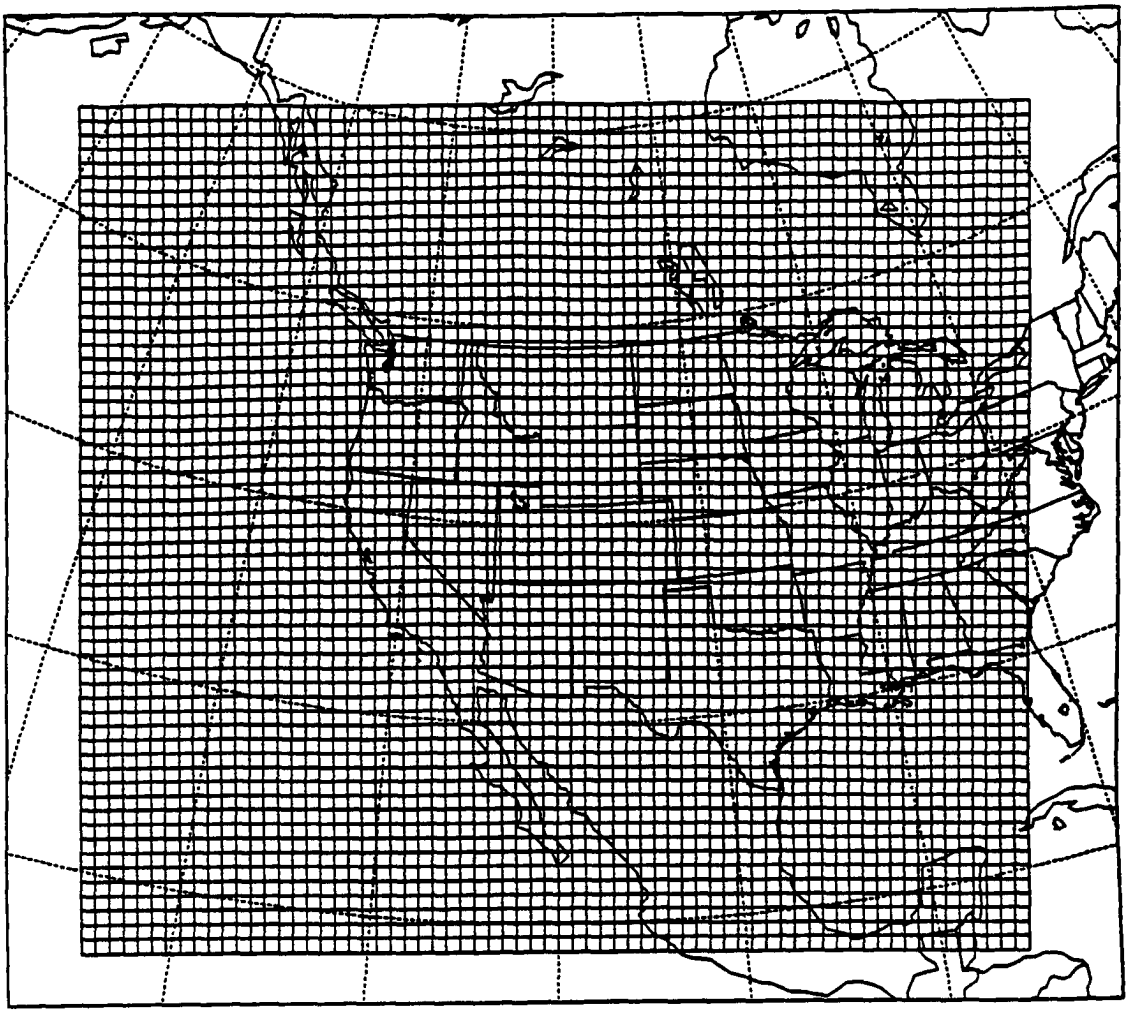


Figure 2.2: RAMS domain with 80 km × 80 km grid overlayed.

2.3.3 Surface parameterization

Each simulation utilized a multi-soil parameterization developed by Tremback and Kessler (1985), where the user specifies the number of soil levels, level spacing, initial moisture content, and temperature. In these simulations, loam was chosen because it was felt that it was probably the most representative considering that the domain stretched across most of the North American continent. The soil module is initialized horizontally homogeneous across the entire land mass. It is possible to make the soil parameters location dependent, but since regional data on soil properties is scarce (and moreover, is not expected to have an important role in winter cyclogenesis for the case studied in this thesis), it was felt that a horizontal homogeneous initialization was necessary.

The soil module calculates a surface energy budget which includes conduction into and out of the soil, sensible and latent heat fluxes, as well as shortwave and longwave radiation fluxes. The module then predicts soil temperature and moisture content at each model timestep.

There were 11 soil levels specified for these simulations, from the surface down to 0.5 m, with five levels within the upper 10 cm. Over a 36-hour simulation, only the upper 10 cm of soil will have a significant impact on the lower atmosphere. Longer simulations would of course involve much deeper soil levels. For a climate study, it would be important to have actual soil moisture and temperature data, or at least realistic data which includes temperature at depths below 30 cm. There is also a vegetation option available in RAMS, although it was not activated for these simulations.

It should also be noted that at present, there is no snow cover parameterization available in RAMS, although it is planned for future versions. The effect of snow cover on the boundary layer is very important in the calculation of surface fluxes, as eluded to in Chapter 1. It affects albedo, surface temperatures, as well as soil moisture. In Fall and Spring, when large areas of the North American continent experience rapid changes in snow cover, this can have a major impact on regional weather.

Johnson et al. (1984) as well as Segal et al. (1991) illustrate the impact of a mesoscale 'snow breeze' on Front Range weather. This snow breeze acts much like a sea breeze

in that it can generate a mesoscale convergence zone above a snow-covered/snow-free boundary. These convergence zones can lead to the formation of clouds and additional snowfall (Johnson et al., 1984). RAMS, lacking any snow-cover parameterization, misses these features altogether. NGM has a snow-cover parameterization in operation, but the model's coarse resolution (i.e., 4 grid increments are 320 km) allows it to only resolve moderate to large-scale snow-cover patterns.

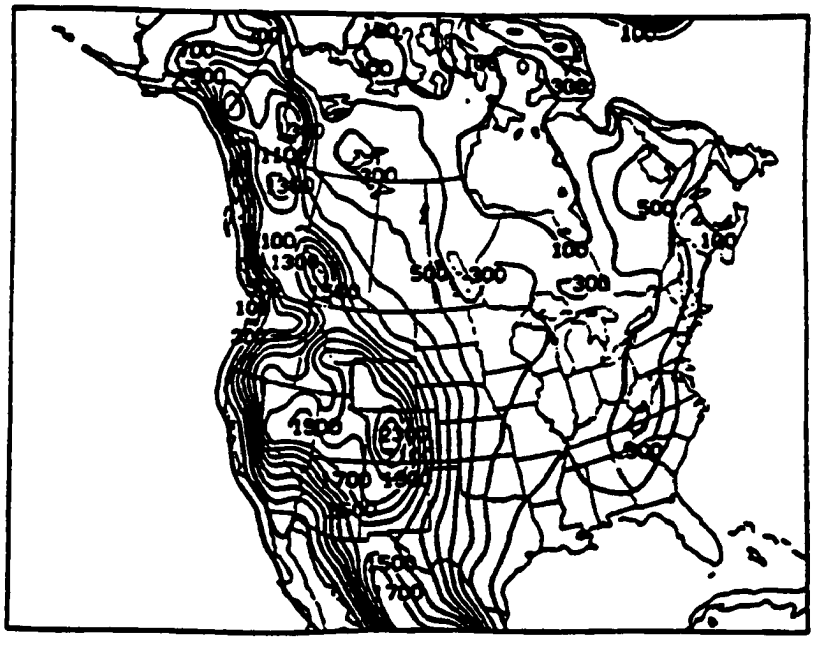
There is the added fact that several storms in the two weeks prior to the 29-30 March simulations had deposited various amounts of snow in the mountains of the western United States, as well as across much of the northern Great Plains. The Colorado Rockies were blanketed by snow, although prior to 29-30 March, there was little snow left below 2000 m along the Front Range.

2.3.4 Microphysics parameterization

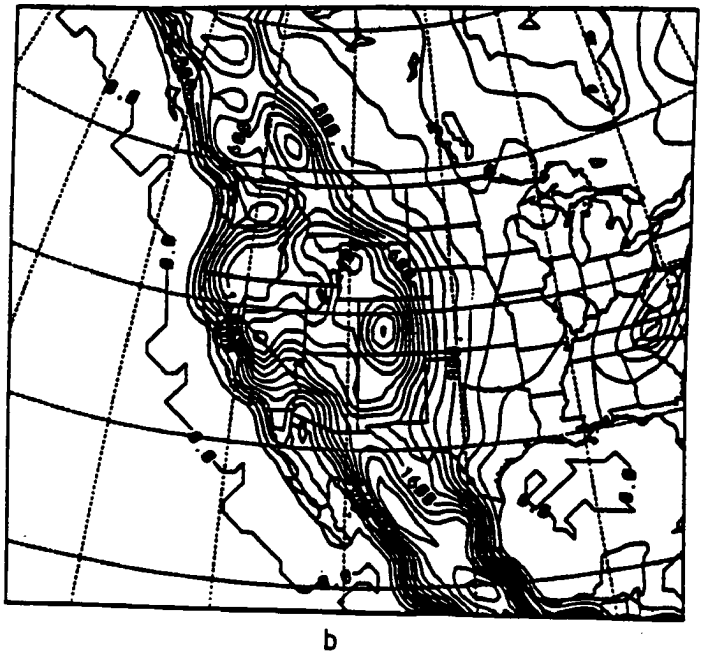
Version 2C microphysics module uses a bulk parameterization which assumes that rain water, pristine crystals, snow, graupel, and aggregates may be represented by a continuously specified size distribution (Flatau et al., 1989). Diagnostic concentrations were used for all species except pristine crystals which used a prognostic scheme. Each species can acquire mass through vapor condensation/deposition, self-collection, or interaction with another species. The model predicts the mixing ratios of each species, while the distribution of a particular species is diagnosed. In all simulations undertaken in this study, the minimum pristine crystal mass was set to be 1.0×10^{-11} kg, a cloud condensation nuclei concentration of 3.0×10^{5} per liter, and a homogeneous nucleation temperature of 233 K.

2.3.5 Topography

The US Navy 10-minute terrain data set was used for model topography (Figure 2.3). This data set was then in turn passed through a silhouette-averaging routine which tends to fill in deep valleys and round-off mountain tops. From Figure 2.3, one can see that this topography is a coarse representation of the actual terrain. Within Colorado, the Rocky Mountains are reduced to one large mountain, centered just west of the center of the state.



a



b

Figure 2.3: (a) NGM 10-minute topography, and (b) RAMS 10-minute topography (200 m contour interval).

The height of the barrier is some 3200 m, when in reality the barrier height is actually around 4000 m. At this resolution, neither the Cheyenne Ridge and Palmer Divide are present. One should also note the terrain gradient on the eastern slope. Instead of a steep barrier along the Front Range, the slope is stretched out to the Colorado-Kansas border. This has the effect of reducing the barrier effect of the mountains. It should be mentioned that NGM also uses the Navy 10-minute terrain data (Hoke et al., 1989), however, their smoothing routine reduces the height of the Rocky Mountains to about 2400 m (compare Figure 1.5 to Figure 2.3).

2.3.6 Model initialization

Data used for the initial and boundary condition were derived from the RAMS Isentropic Analysis package (ISAN) (Pielke et al., 1992, Cram, 1990, Tremback, 1990). This package utilizes three NMC data sets; the NMC 2.5° global analysis, rawinsonde and upper air soundings, plus surface observations. These three data sets are archived at NCAR. Data analysis takes place in isentropic coordinates. The last stage of the ISAN package interpolates data to the model grid.

2.3.7 Nudging and boundary conditions

RAMS lateral boundaries are nudged every 12 hours to observed fields through the ISAN package. In the nudging scheme, an extra tendency term is added to each prognostic equation which forces the predicted variable towards the observed value (Pielke et al., 1992). Since NGM output is a true forecast, direct comparison between it and RAMS is limited, because RAMS is not a true forecast model beyond the first 12 hours of simulation. However, the extra tendency term also includes a nudging weighting function that can be adjusted by the user in order to reduce/increase the influence of the observed fields on the interior of the model domain. This nudging takes place in a zone along the lateral boundaries which is specified once again by the user. A five grid point wide nudging zone was used in each of the simulations in this thesis. The weights for all but one of the simulations (ST2) was a sensitivity simulation that used different weights), were .75, .45, .25, .10, and .05.

Chapter 3

RAMS SIMULATIONS

RAMS' ability to simulate winter storms along the Front Range was addressed in part by Wesley (1991) in his dissertation. As mentioned in Chapter One, the model did quite well overall. He also investigated the effects of complex terrain on the efficiency and production of snowstorms. In many respects this thesis is a continuation of that work. The purpose of these additional simulations is to learn more about model initialization and scales of resolution, as well as some of the dynamic structure of these storms.

Table 3.1 is a listing of the simulations contained in this study, as well as some of the key model options. All of the simulations were begun at 0000 UTC 29 March and ran for a total of 36 hours.

Table 3.1: List of simulations and key model options.

Type	Designation	Comments
Control Run	CR	Full microphysics, one grid 80 km horizontal; soil moisture = 0.2; nudging = .75, .45, .25, .10, .05
Sensitivity 1	ST1	Snow variable turned off in microphysics module
Sensitivity 2	ST2	Same as CR except nudging set to .10, .10, .10, .10, .10
Sensitivity 3	ST3	Same as CR except temperature offset -5° all levels
Sensitivity 4	ST4	Same as ST3 but Silhouette – averaging = 0.1
Sensitivity 5	ST5	Fine grid 20 km spacing in x, y; 30 vertical levels, otherwise same as CR
Sensitivity 6	ST6	Same as CR but with flat terrain at sea level

3.1 Control Run

This simulation contained only one grid, with 70, 62, and 17 grid points in the x, y, and z directions. The timestep was 90 seconds and the upper boundary was a wall. The upper 10 cm of the soil had the same temperature as the lowest level in the atmosphere at start time. Soil moisture was 9% of total soil volume. For this simulation it was decided from previous RAMS research that the microphysics module should be used with all five water species activated. Figure 3.1 shows the wind vectors at 5800 m (MSL) at both 12 hours, 24 hours, and 36 hours into the simulation, while Figure 3.2 shows the wind vectors at 3045 m at the same three times. The model geostrophic winds (not shown) closely correspond with the geostrophic wind speeds calculated from the NMC 500 mb analysis. The 500 mb flow was in general 5 m s^{-1} to 10 m s^{-1} stronger at Grand Junction during 29-30 March than they were at Denver. Figures 3.3 – 3.5 are plots of the perturbations in the Exner function at 12, 24, and 36 hours into the simulation at a height of 5800 m as compared with NMC's 500 mb height fields. What these plots essentially show is perturbations in the pressure field. At 1200 UTC 29 March there is a substantial pressure drop over north central Wyoming. This low pressure center then tracks south-southeast over Colorado, where it remains stationary for the duration of the simulation. Figures 3.3b – 3.5b are the NMC analyzed fields for comparison with the previous three figures. The agreement is quite good, indicating that RAMS did a good job with the synoptic-scale flow. It is important to remember that nudging only takes place in a zone five grid points wide around the boundary, the center of the domain is left to evolve on its own. The only problem that RAMS exhibits is shown in Figure 3.5a, where the low pressure center remains in Colorado, while NMC data (Figure 3.5b shows it in central New Mexico.

Figure 3.6 shows the evolution of surface winds (200 m AGL) over the length of the simulation. Initially there is a slight northeast flow (upslope) along the Colorado Front Range, but the upslope is short-lived. These plots agree quite well with the mesonet data. During most of the period the flow is northwesterly or northerly which is not conducive to snowstorms. East-west wind vector plots at Denver are shown in Figure 3.7. These show that at 0000 UTC 29 March a fairly deep layer of upslope flow was present. By 1200 UTC

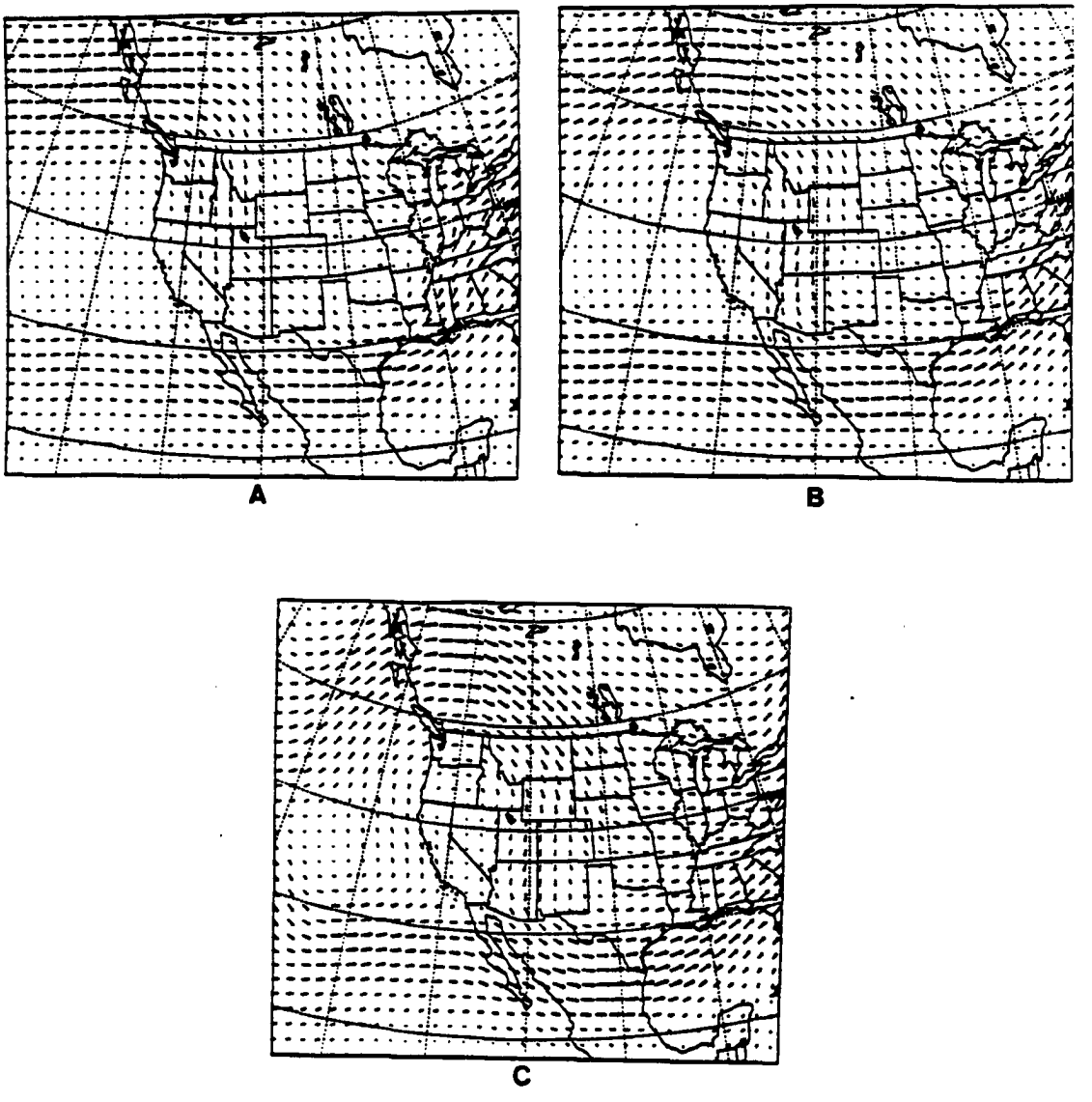


Figure 3.1: Wind vectors at 5800 m (MSL) for (a) 1200 UTC 29 March, (b) 0000 UTC 30 March, and (c) 1200 UTC 30 March.

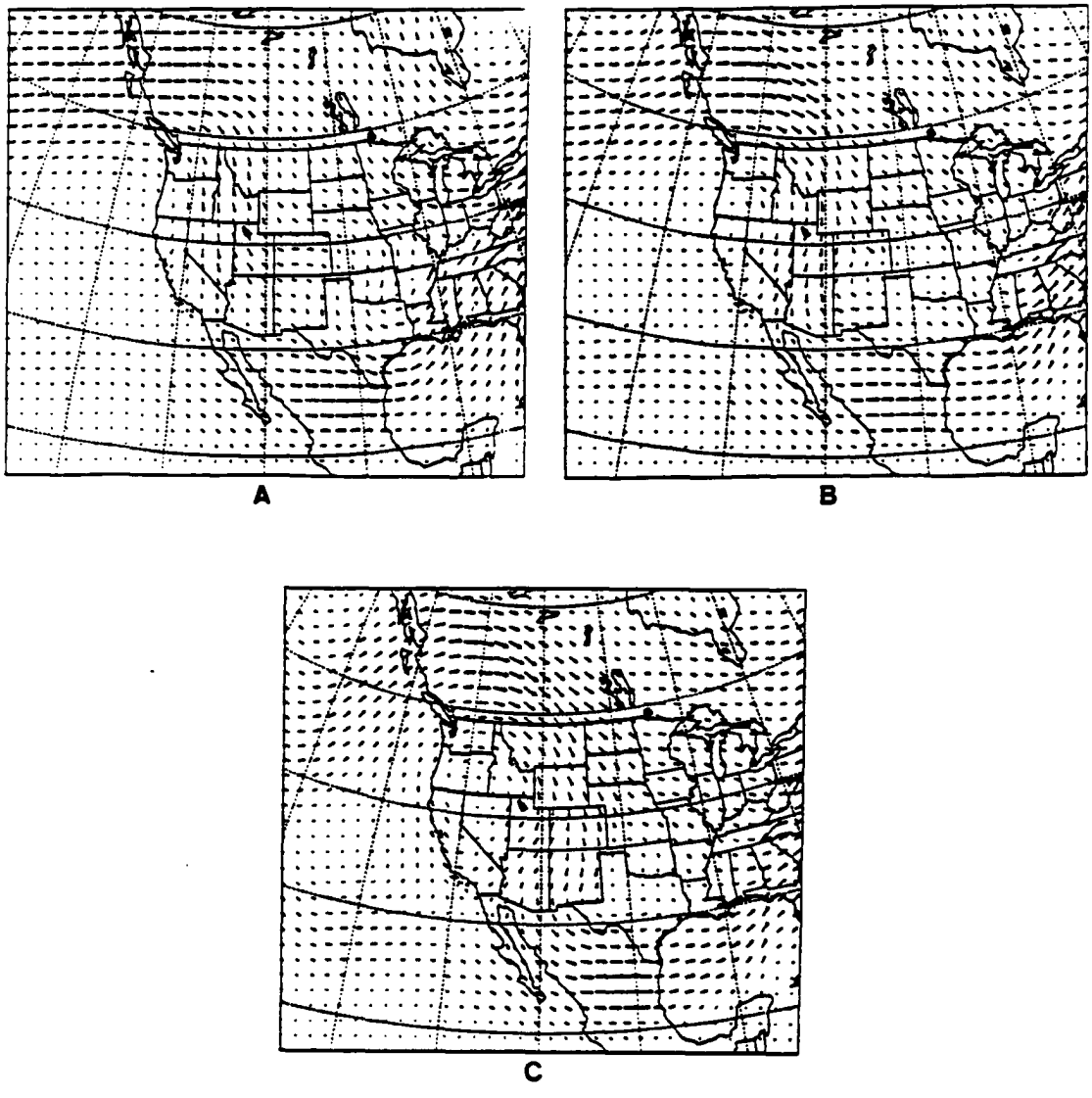
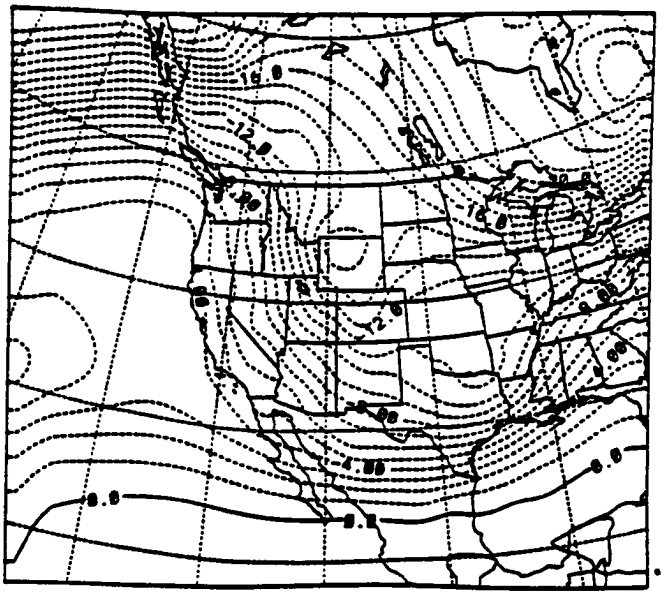


Figure 3.2: Wind vectors at 3045 m (MSL) for (a) 1200 UTC 29 March, (b) 0000 UTC 30 March, and (c) 1200 UTC 30 March.



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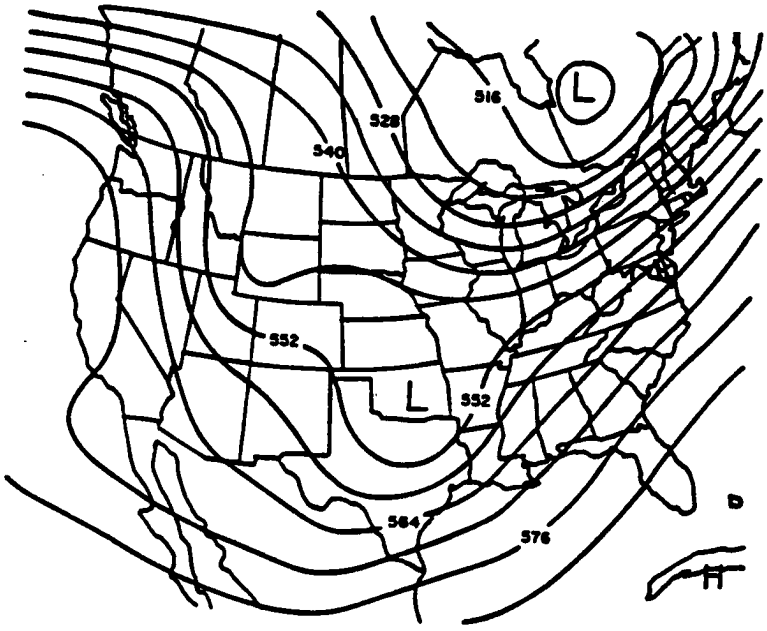
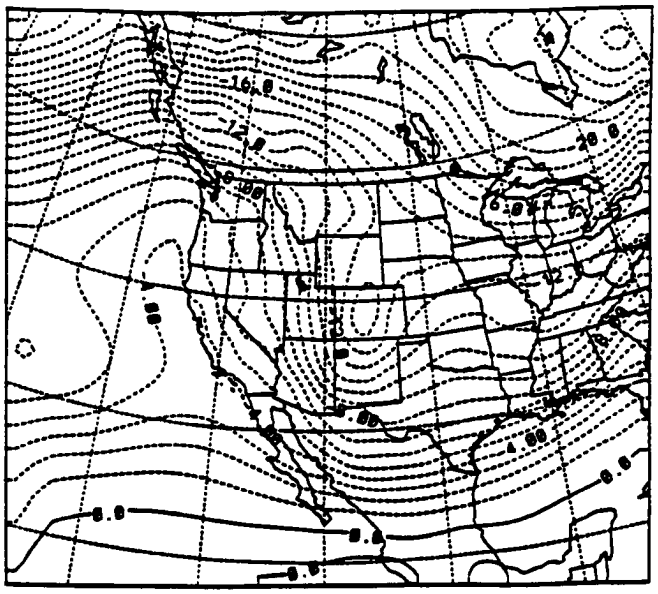


B

Figure 3.3: (a) Perturbation in Exner function at 5800 m (RAMS), and (b) NMC 500 mb heights (contours 60 m) at 1200 UTC 29 March.



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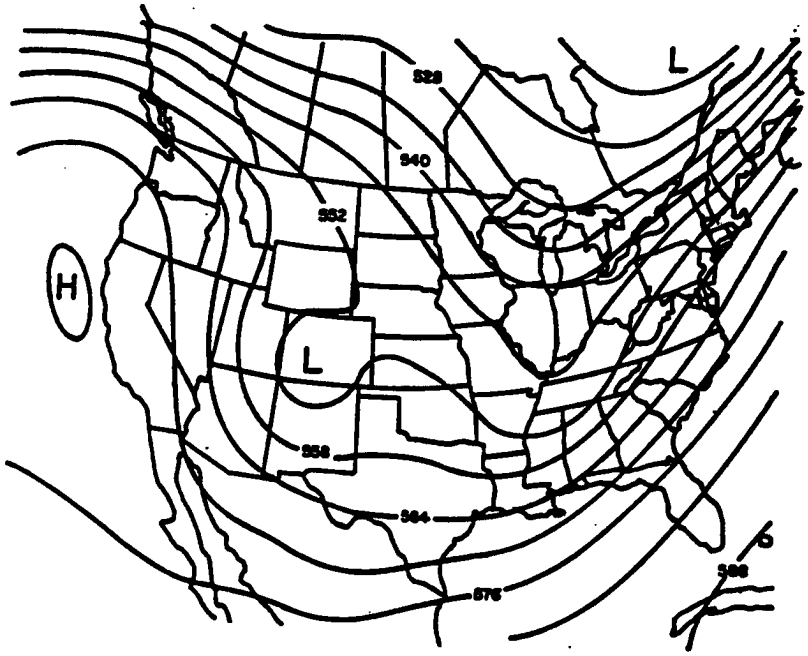
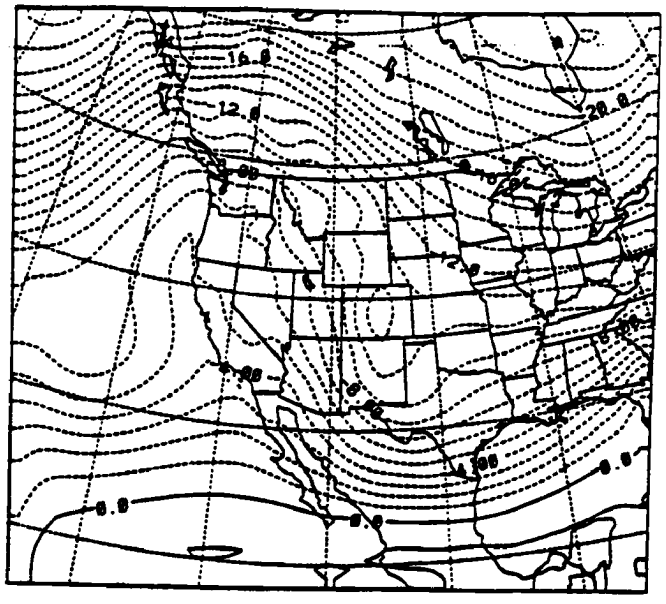


B

Figure 3.4: (a) Perturbation in Exner function (RAMS), and (b) NMC 500 mb heights (contours 60 m) at 0000 UTC 30 March.



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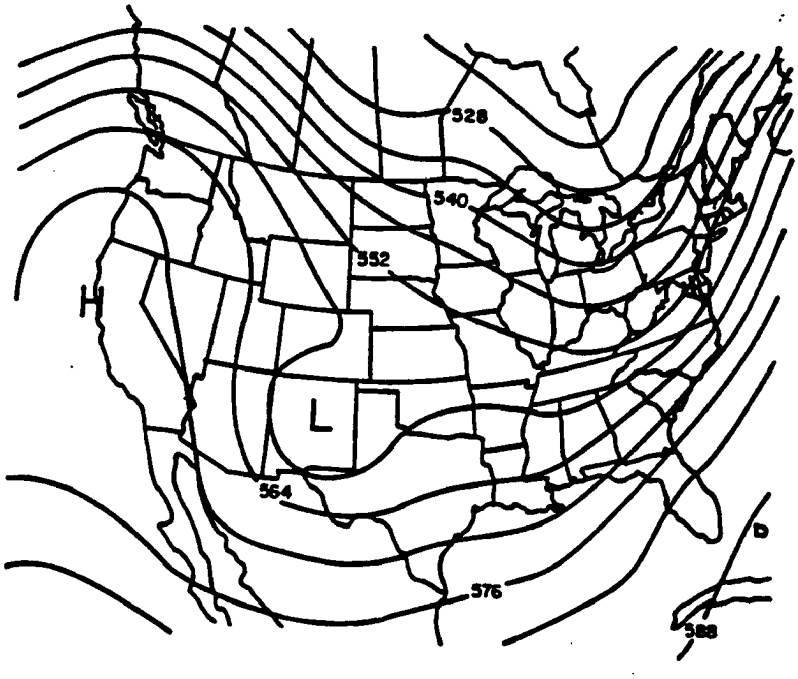


B

Figure 3.5: (a) Perturbation in Exner function (RAMS), and (b) NMC 500 mb heights (contours 60 m) at 1200 UTC 30 March.

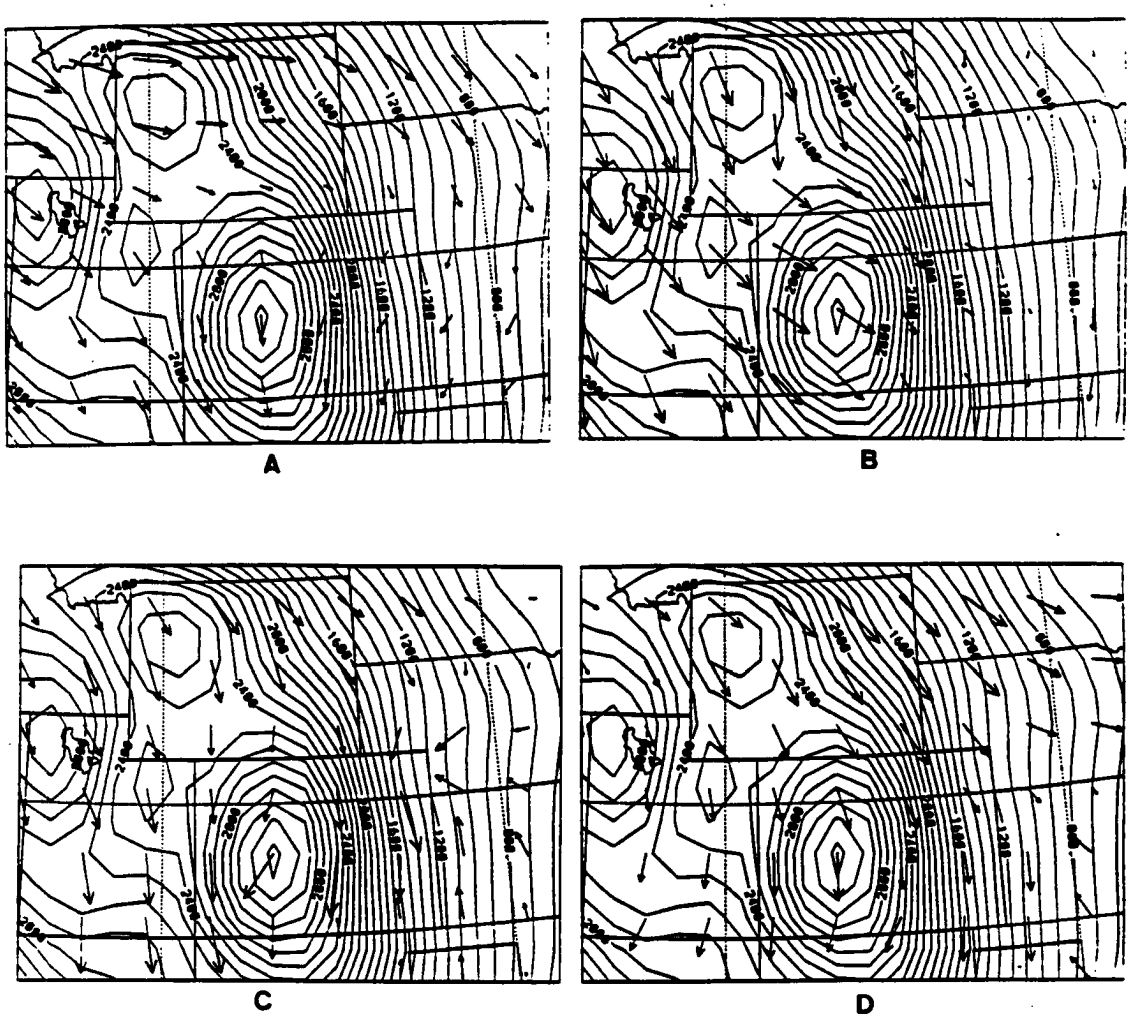


Figure 3.6: Surface winds (200 m AGL) for (a) 0000 UTC 29 March, (b) 1200 UTC 29 March, (c) 0000 UTC 30 March, and (d) 1200 UTC 30 March.

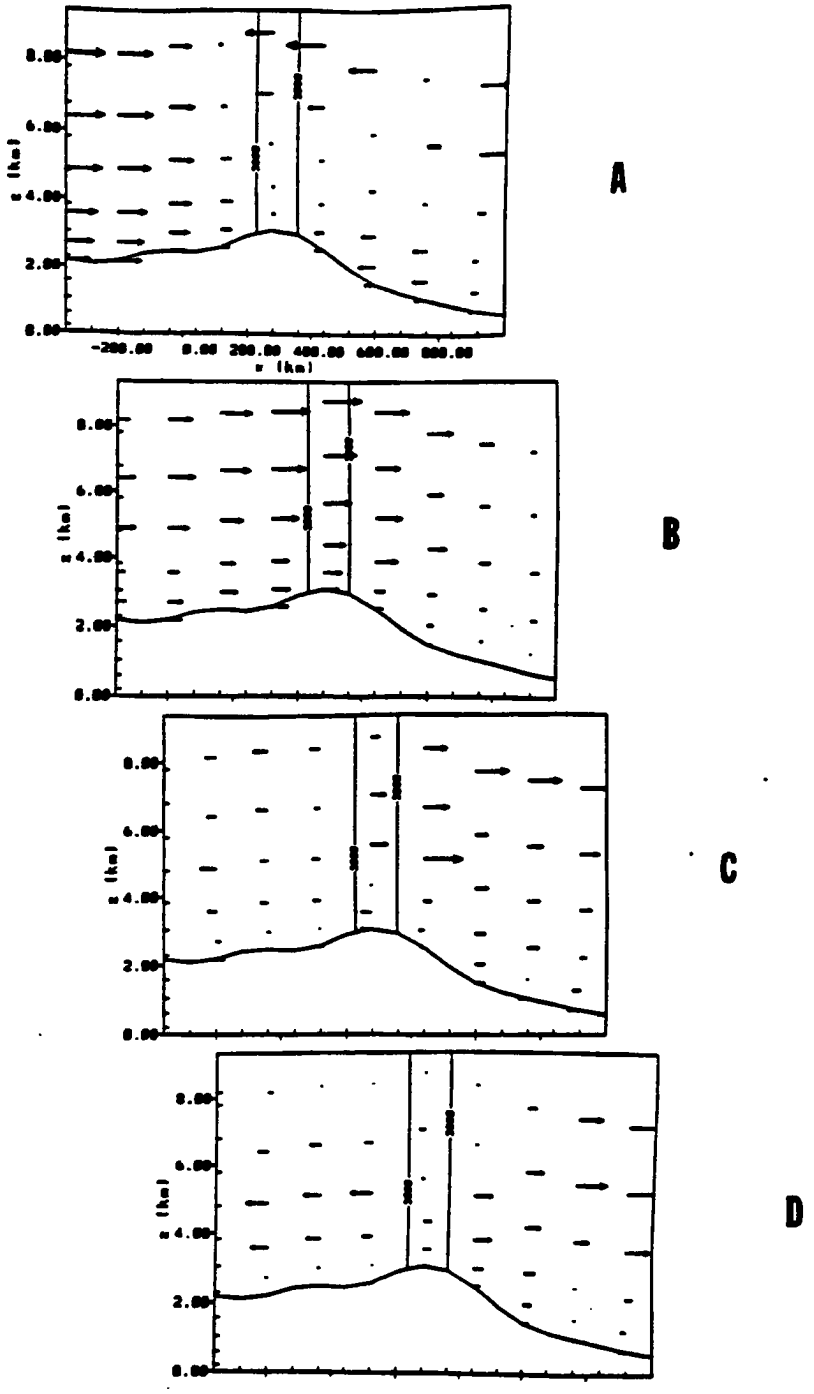


Figure 3.7: $x - z$ cross section of winds near Denver at (a) 0000 UTC 29 March, (b) 1200 UTC 29 March, (c) 0000 UTC 30 March, and (d) 1200 UTC 30 March.

the flow consists of westerlies at all levels, but 12 hours later, 0000 UTC 30 March, there is a shallow upslope flow once again. This is also revealed by plots of potential temperature which at 0000 UTC 30 March show a surge of colder air moving into NE Colorado (Figure 3.8).

Figures 3.9 indicates the total accumulation of precipitation and NGM forecast after 24 and 36 hours. Figure 3.10 shows the observed 24 hr precipitation ending 12 Z 30 March. The broad band of precipitation in the Appalachians and Ohio Valley was handled fairly well, except that RAMS generated a sharp cut-off of the precipitation on the eastern edge. Observations from the Carolinas indicate that as much as 8 cm of rain fell during this storm in central South Carolina. Why RAMS truncated the precipitation the way it did seems to be explained by the fact that it lies close to the lateral boundary where the microphysics module often has trouble due to nudging.

Figure 3.11 shows the Colorado region in much greater detail. After 36 hours the heaviest precipitation, approximately 2.4 cm, is along the Nebraska, Kansas, Colorado border area with two smaller maxima centered over the central Rockies and southeast Colorado. It does appear that RAMS greatly overestimated the amount of precipitation that fell in the Nebraska, Kansas, Colorado area by as much as 1.5 cm (compare with Figure 1.4). This is probably due to the fact that model topography is too coarse, with the terrain gradient along the eastern edge of the Rockies extending too far out into Kansas and Nebraska (i.e., the barrier of the Front Range is not realistic). Another factor in this overestimation of precipitation may be that the low pressure center, stalled over Colorado and hence provided a mechanism for enhanced precipitation. Vertical velocities over the Nebraska, Kansas, Colorado border area were in the range of 0.2 m s^{-1} to 0.3 m s^{-1} at 0000 UTC 30 March, which led to an increase of 1.4 cm of water for the 12-hour period between the 24 and the 36 hour of simulation. If the aforementioned theories are valid, then it appears that the overprediction in precipitation was not a microphysics problem, but an error in the positioning of the synoptic low. This illustrates the importance of positioning errors in these types of low resolution models. An error of several hundred kilometers (two or three gird points) in turn carries over to the microphysics of the storm and leads to erroneous forecasts much of the time.

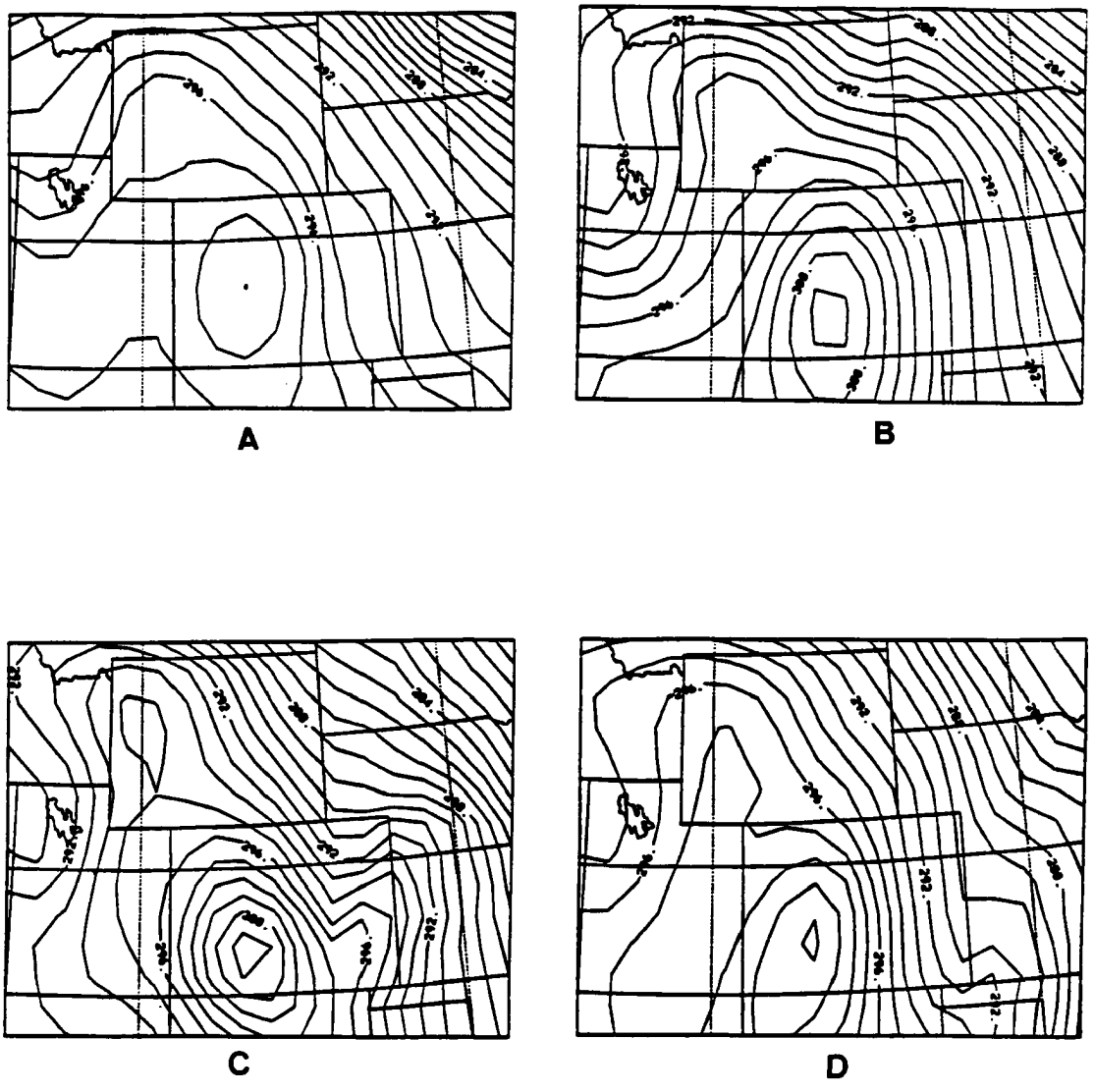


Figure 3.8: Potential temperatures at 200 m (AGL) at (a) 0000 UTC 29 March, (b) 1200 UTC 29 March, (c) 0000 UTC 30 March, and (d) 1200 UTC 30 March (contours 1°K).

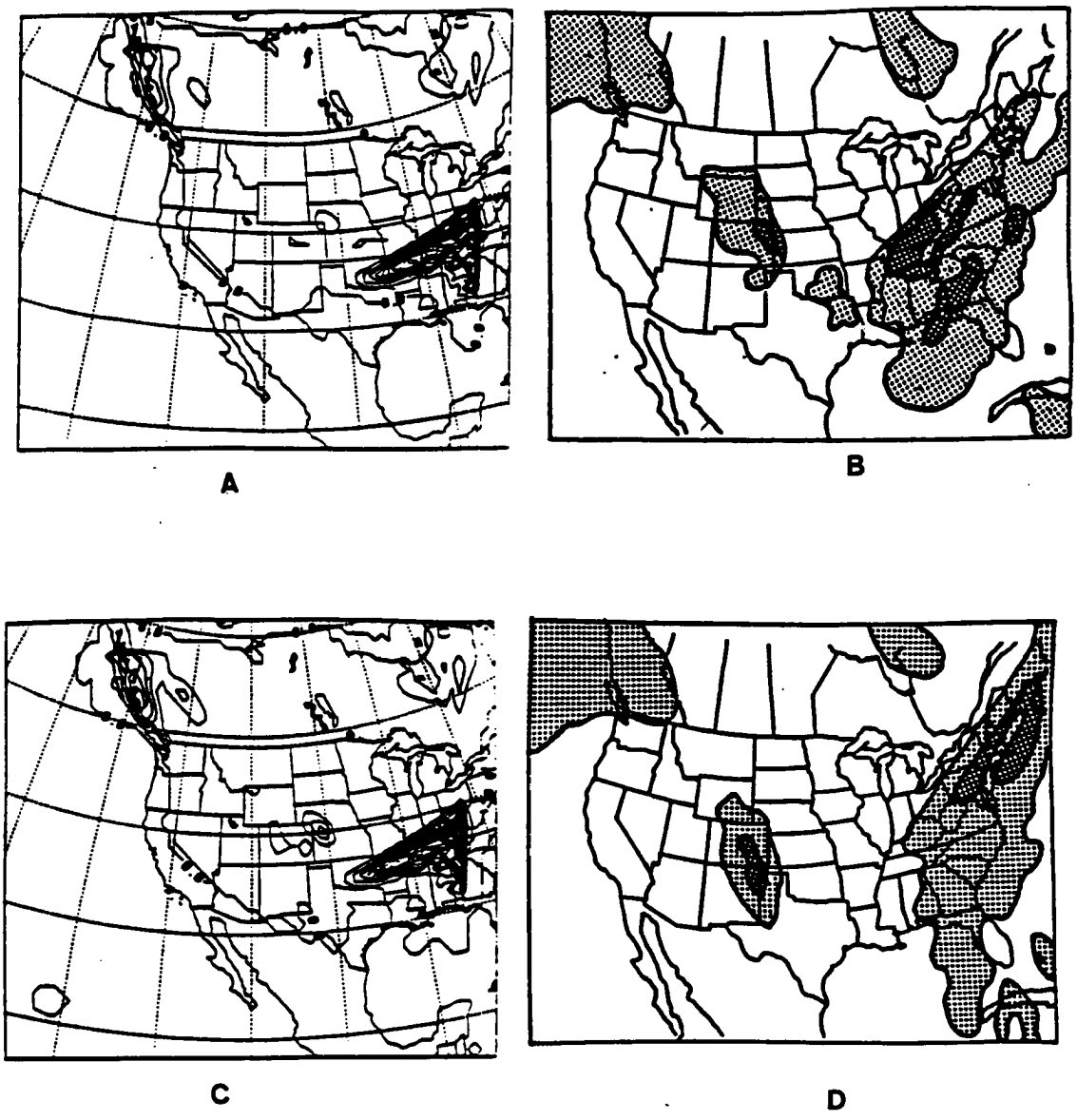


Figure 3.9: (a) RAMS total precipitation at 0000 UTC 30 March, (b) NGM 24-hr forecast for total precipitation (valid 0000 UTC 30 March), (c) RAMS total precipitation at 1200 UTC 30 March, and (d) NGM 36-hr forecast for total precipitation (valid 1200 UTC 30 March). Light stipple 0-1 inches, dark stipple 1-2 inches.

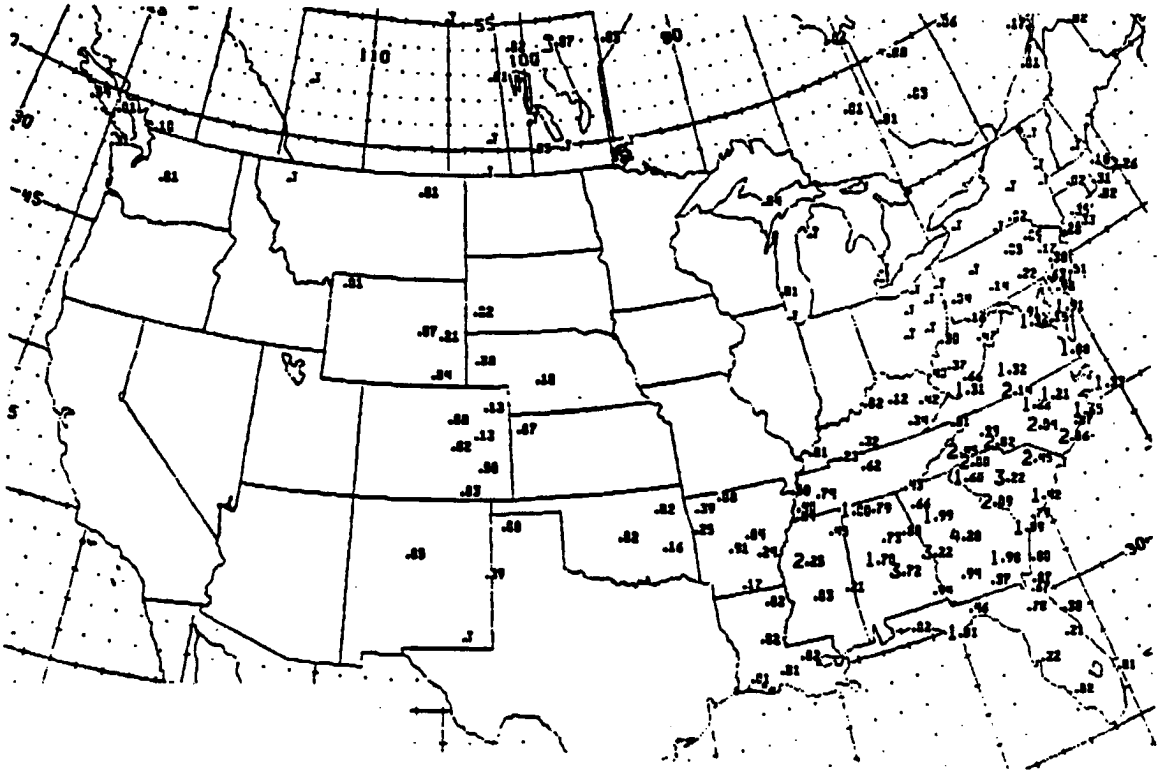


Figure 3.10: 24-hour precipitation totals (inches) for the period 1200 UTC 29 March to 1200 UTC 30 March.

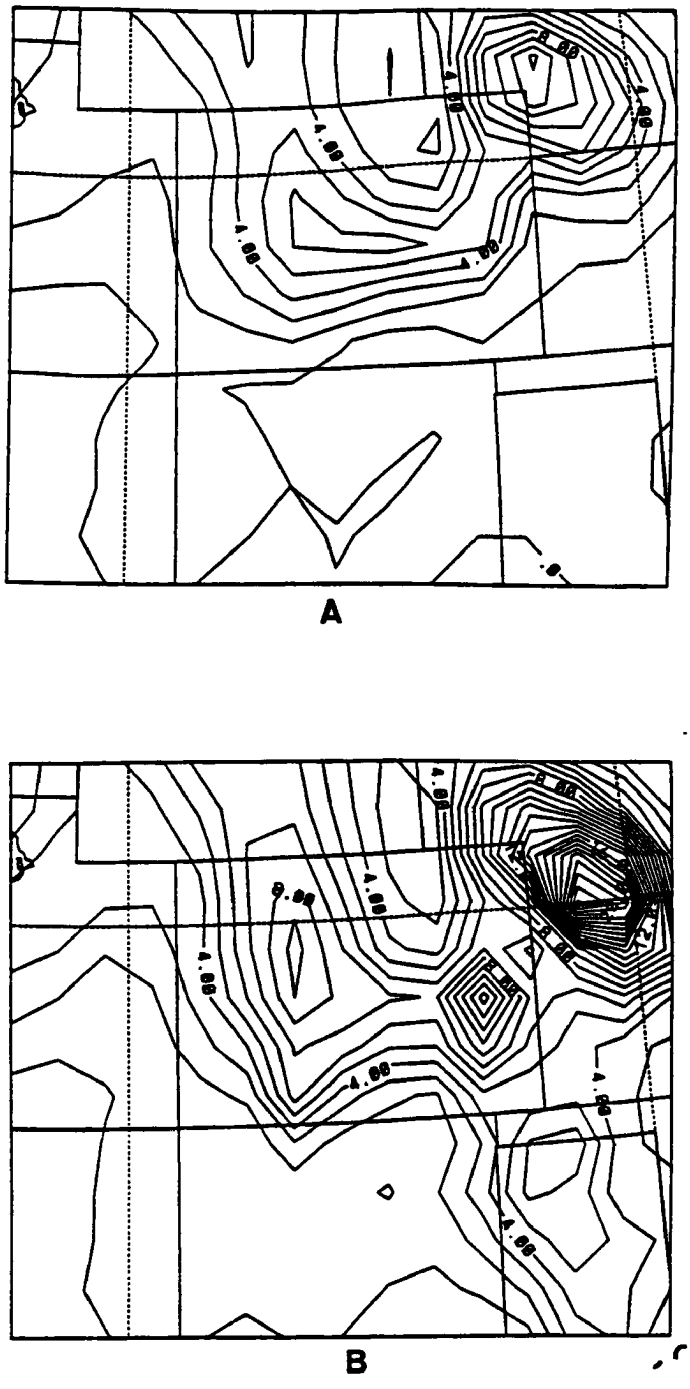


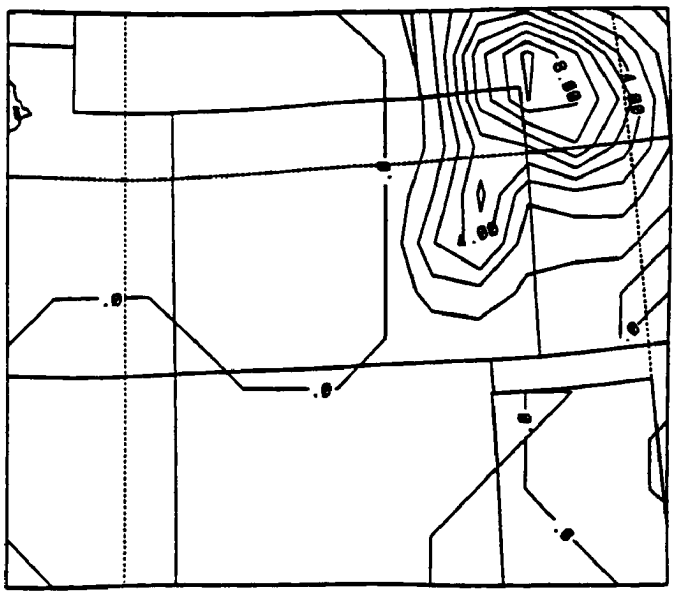
Figure 3.11: Total precipitation (mm) for (a) 0000 UTC 30 March, and (b) 1200 UTC 30 March.

Figures 3.12 and 3.13 show the breakdown between rain and snow. About 90% of the Nebraska/Kansas/Colorado area precipitation fell as rain, while about 75% fell as rain over eastern Colorado. Weather reports from this region show that the precipitation that fell over the Nebraska/Kansas/Colorado area fell mostly in the form of rain with a little light snow mixed in. In eastern Colorado, a mixture of snow and rain fell, depending on the exact location. It appears that the two areas that have the largest amounts of total precipitation, those two areas being a triangle defined between Trinidad, Walsenburg, and La Junta in addition to the area around Eads (Figure 1.4), had the majority of their precipitation in the form of snow. Table 3.2 gives snowfall totals for selected sites in the Colorado region.

Why did RAMS predict more rain than snow? It appears from the plots of potential temperature (Figure 3.8) that the model surface temperatures are too warm, which is probably a result of a large surface heat flux that is generated from a soil profile which is too warm for that time of year. Observations from Fort Collins show about a 7°C positive temperature anomaly in the model surface temperatures.

RAMS predicted less than 0.8 cm of water equivalent for portions of the Colorado Rockies. This verified quite well with most of the mountain resorts only receiving a trace of snow or no snow at all. The two mountain areas that did receive 20 cm or more of snow, Telluride and the Mt. Evans research station, were not the norm. What caused greater amounts of snowfall in these areas is probably due to local dynamics, something both models with about 300 km resolution are not able to detect. The fact that RAMS snowfall prediction (Figure 3.13) shows a large areal coverage in the mountains is because the model topography is far too simplistic to reveal individual ranges within what is normally referred to as the Colorado Rockies. The model did produce very small amounts of pristine crystals, aggregates, and graupel, and what was produced fell out over the mountains.

The overall precipitation pattern as predicted by RAMS in Colorado and adjacent states shows a high degree of variability over short distances, which seems to indicate that 'snowband-like' features were present. Even with RAMS deficiencies, it is encouraging to



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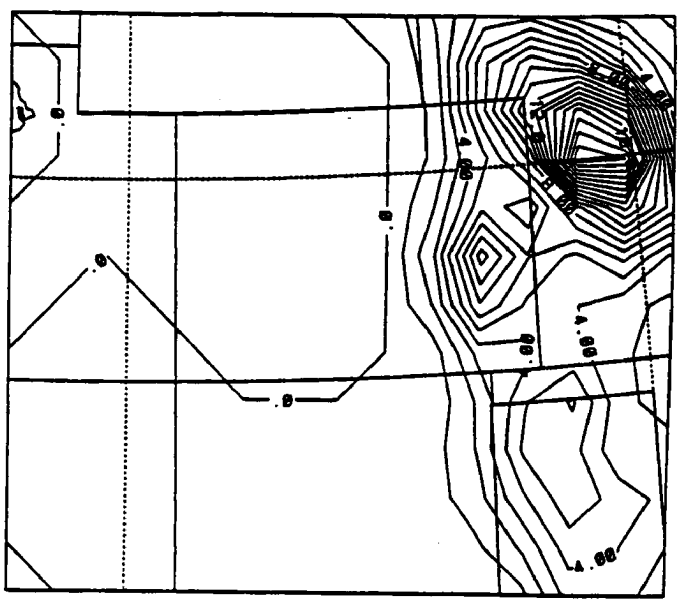
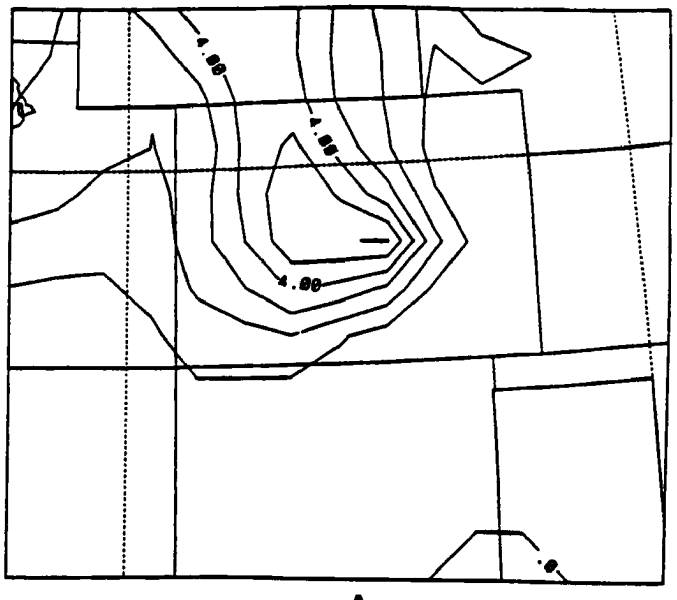


B

Figure 3.12: Total accumulation rainfall (mm) for (a) 0000 UTC 30 March, and (b) 1200 UTC 30 March (contours 1 mm).



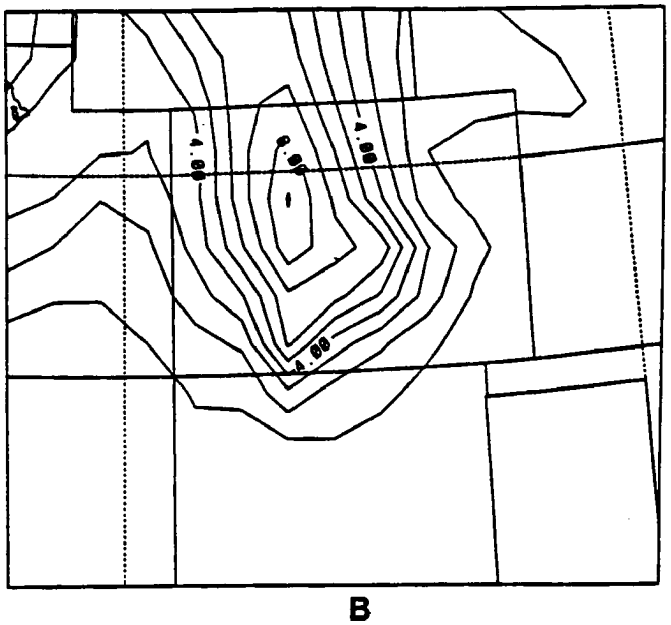


Figure 3.13: Total accumulated snowfall (mm) for (a) 0000 UTC 30 March, and (b) 1200 UTC 30 March (contours 1 mm).

Table 3.2: Storm totals of snowfall and water equivalent for 29-30 March, 1991.

Location	Snowfall (cm)	Water Equivalent (cm)
Aguliar	25	1.1
Akron	NA	0.6
Alamosa	T	T
Aspen	5	0.3
Blanca	3	0.4
Breckenridge	0	0
Canon City	T	T
Colorado Springs	T	T
Craig	3	0.2
Denver	2	0.2
Durango	T	T
Eads	NA	2.6
Fleming	10	1.3
Fraser	8	0.5
Fort Collins	T	T
Grand Junction	T	0.2
Grand Lake	5	0.3
Kim	30	3.6
La Junta	18	2.3
Leadville	T	T
Leroy	13	1.0
Limon	4	0.3
Montrose	NA	T
Ouray	21	1.5
Pueblo	T	T
Rye	30	3.0
Silverton	5	1.7
Trinidad	13	0.5
Whootton Ranch	38	2.4
Winter Park	18	1.25

see that the model is able to produce some finer structure to the areal distribution of the various forms of precipitation, and not just produce blanket coverage as the NGM does.

Since both NGM and RAMS models have about the same resolution ($\Delta x \approx 300$ km), one has to ask why RAMS did somewhat of a better job in the overall forecast? It is probably a combination of several factors, most importantly terrain representation and low-level flow field representation.

3.2 NGM and RAMS Evaluated

Both models did moderately well in predicting the movement and pressure values of the low that moved from central Wyoming to New Mexico. However, the NGM 0000 UTC 30 March forecast (24-hour forecast) has the low over southwest Colorado, about 150 km too far west and the geopotential heights about 20 m to 30 m too low. These 'errors' are not large, but do suggest an over-deepening of the low, allowing moisture from the Gulf of Mexico to be advected over the Front Range. RAMS problem with this low has already been discussed. In addition, NGM predicted vertical velocities of 9 cm s^{-1} over the Front Range while RAMS had half that value after 24 hours of simulation.

As for precipitation, NGM 24- and 36-hour forecasts called for wide-spread precipitation from southern Montana through NE New Mexico, with storm totals of about 3.25 cm water equivalent in some locations of the southern Rockies. NGM's main problem with the forecast was the overprediction of moisture. RAMS, on the other hand, did show the finer structure to the storm and realistic precipitation values for much of the region, except for the one problem area on the Nebraska, Kansas, Colorado border.

It should be mentioned that the 12-hour forecast produced by NGM for 0000 UTC or 1200 UTC 30 March had not yet corrected itself from the earlier forecast made at 0000 UTC 29 March. There was a 15% to 20% reduction in precipitation values, but the forecasts were still very inaccurate.

Chapter 4

SENSITIVITY SIMULATIONS

In order to facilitate a better understanding and confidence in the RAMS model, a number of sensitivity simulations were conducted, some aspects which have already been briefly discussed previously. With the advent of complicated multicomponent atmospheric models, it is very easy for the modeler/user to overlook the small details such as topography and surface fluxes in favor of the more prominent components such as large-scale dynamics and the actual setup of the grid. One of the main objects of these sensitivity simulations is to see how significant some of the major model components (i.e., dynamics, surface fluxes, topography) are in creating weather at any given location within the model domain.

It is my *a priori* assumption that for any given atmospheric event, one component may be more critical than other components. However this can change from one event to another, and even change over time within the same event. For example, surface fluxes may not be as important as synoptic-scale dynamics during the winter months, however in the spring and summer, those surface fluxes will play a much greater role in weather events. With that in mind, one has to remember that this study is only considering one specific storm which took place in late March 1991. One should be careful not to form too many conclusions or generalizations from such a limited study. In defense of this study, however, models need to be tested against the real atmosphere as well as other models so that its usefulness and limitations can be ascertained. In many respects, these sensitivity simulations are building a foundation for future work and improvements in the model.

4.1 Sensitivity 1

This simulation involved the adjustment of the microphysics module so that the snow species was turned off. This allowed precipitation to fall as one of the following species:

rain, pristine crystals, aggregates, or graupel. When compared to the control run, the maxima in total precipitation was reduced by 12% and accompanied by a slight shift in the areal distribution of one of the three maxima (Figure 4.1).

Concentrations of rain, aggregates, and pristine crystals towards the total precipitation were reduced from the control run but were somewhat offset by an increase in the concentration of graupel. I propose that the decrease in the total precipitation is due to differences in evaporation and condensation rates between this simulation and the control run. For example, if the average size of a snow crystal in the control run is larger than the average size of pristine crystals or graupel in ST1, then given equal fall velocities, temperatures, etc., one would then expect to find more total water equivalent precipitation in the control run due to reduced evaporation of the larger size particles.

What is interesting about this simulation is the fact that there was a decrease in maximum total rainfall from 20 mm in the control run to 18 mm in ST1. I surmise that there was a reduction in the condensation rate of rain when the snow species was turned off, or an increase in sublimation or even evaporation took place as precipitation particles fell towards the surface.

4.2 Sensitivity 2

The purpose of this simulation was to see how the model responded to a reduction in the nudging weights. It has already been noted that RAMS is nudged every 12 hours in the outermost five grid points towards the observations. The weights in the control run were .75, .45, .25, .10, and .05 (from the outer grid inward). This means that the outer set of grid points are nudged towards the observations more than the remaining four grid points in the nudging zone.

For this sensitivity simulation the nudging weights were set to .10, .10, .10, .10, and .10. In essence, the model was allowed to evolve more on its own, with reduced help from the observations. With this reduction in nudging, the wind fields and pressure patterns changed very little. In the Colorado region, there was up to 5% differences in precipitation totals, with a shift of 100 to 150 km in the precipitation maxima toward the west. These

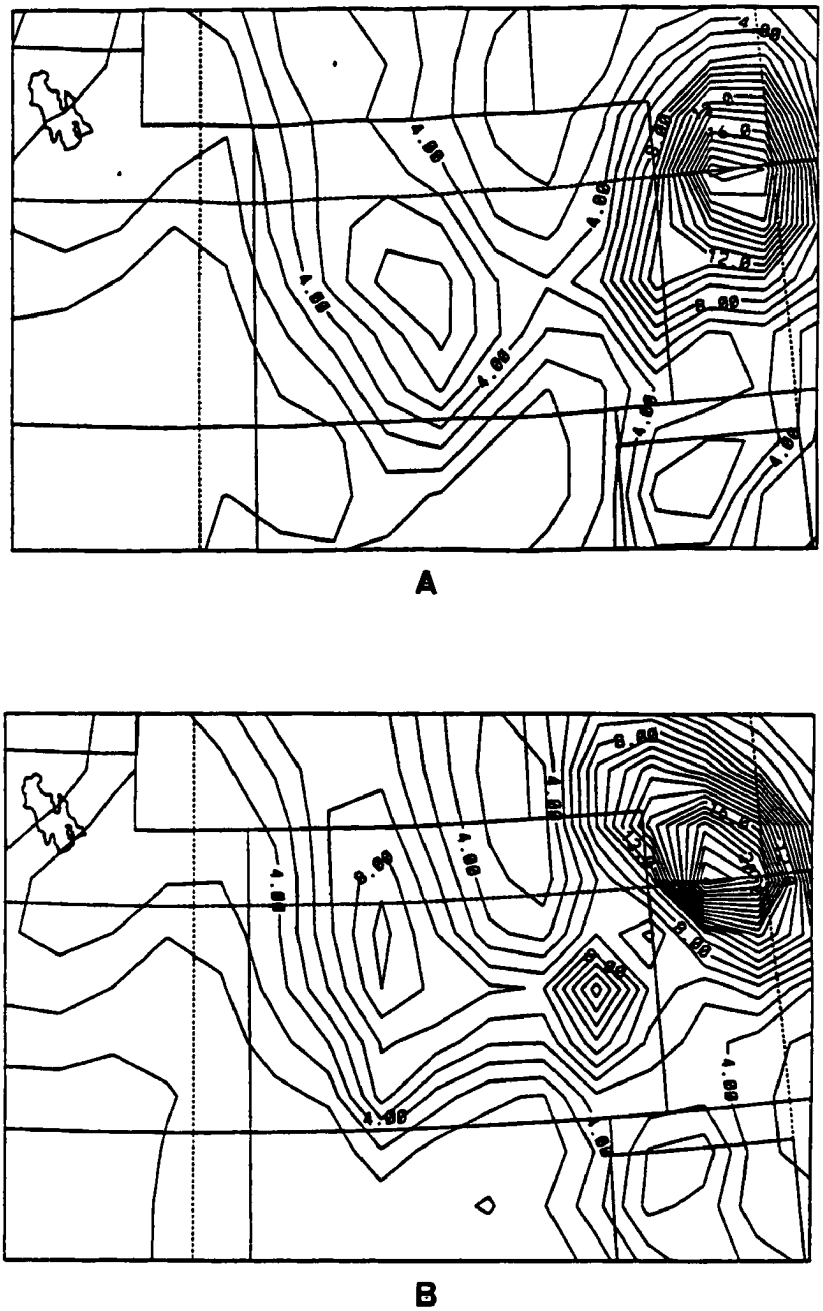


Figure 4.1: Total precipitation (mm) at 1200 UTC 30 March for (a) Sensitivity 1, and (b) control run (contours 1 mm).

results indicate that the model is doing an excellent job of calculating all fields without having to rely on the nudging for corrections.

4.3 Sensitivity 3

This simulation was conducted in order to evaluate the role in which surface fluxes played in this particular storm. There are several ways that variations in surface fluxes can be introduced, namely changing the albedo, soil moisture, or soil temperature. I choose to modify the soil temperature profile in such a way that the temperate in each of the 11 soil layers was colder than in the control run. Since this simulation only lasted some 36 hours, it is difficult to know *a priori* how this would effect the model output. I would expect, however, that the longer the simulation lasted, the more influence surface and subsurface parameters have on the final solution.

Since the domain covers some 75% of the North American continent, and since the soil module was initialized horizontally homogeneous, I chose loam to represent the typical soil type. Loam was chosen primarily because it contains sand, silt, and clay particles. Its physical properties are a blend of all three soils, with no one dominating. Hence soil thermal diffusivity, heat capacity, moisture capacity, and moisture diffusivity are an average of all three soil types. The soil moisture was maintained at 9% of total soil volume, the same as in the control run.

Soil temperatures are calculated from a soil temperature offset parameter which the user specifies in the model. This is then added or subtracted from the potential temperature in the lowest atmospheric level, depending on if the soil is to be warmer or colder than the air temperature. Since the model was initialized at 0000 UTC 29 March, the local time in the center of the domain (Colorado) was 1700 UTC 28 March. The temperature in Denver at that time was about 7°C according to the model, with a temperate offset of -5°C, the soil temperature was 2°C in that area. This is a reasonable value for that time of year, although most of the upper 20 to 25 cm of the soil was probably still frozen. Since this procedure applies to the entire domain, it is possible that in areas to the north, the soil may be warmer than air temperature, and vis versa in warmer climates. The soil

response responds much slower to thermal forcing than the air does, of course, so this type of initialization over a large domain has to be treated with some skepticism. The only other option is to make the soil temperature (and soil moisture) grid dependent. This is the ideal solution but a lack of data prevents its implementation much of the time.

The output from ST3 shows no noticeable difference in the surface wind pattern between this simulation and the control run. Looking at the total precipitation one sees almost identical patterns and amounts at 0000 UTC 30 March (24 hrs of simulation). However, 36 hours into the simulation there is a reduction in precipitation, on the order of 10 mm (42%) at the Nebraska/Kansas/Colorado maxima (Figure 4.2). In addition, the secondary precipitation maxima present in eastern Colorado in the control run is absent in ST3. Looking at plots of rainfall totals, it is apparent that after 36 hours, the precipitation maxima went from 20 mm in the control run to 11 mm in ST3 (Figure 4.3). The cooler soil temperatures in ST3 did not cause any significant difference in snowfall totals over the mountains.

These results indicate the influence that surface fluxes have on precipitation processes. Not only is the supply of moisture in many cases determined by surface fluxes, but these same fluxes can alter the dynamics of the atmosphere as well. Lower soil temperatures caused the soil skin temperature to also be lower than it was in the control run. This in turn caused reductions in the amount of moisture that could be evaporated, and reductions in the strength of the parameterized convective plumes. The net result is a reduction in precipitation totals (mainly in rainfall) in addition to some shifts in the areal distribution of that precipitation.

In the mountains where the soil temperatures were already cold in the control run, there was little evidence of change in the precipitation pattern. The implication is that one cannot simply ignore surface and soil properties as being irrelevant or minor components to model dynamics and thermodynamics in even winter storms.

4.4 Sensitivity 4

The importance of realistic terrain representation has been eluded to already in this study. This simulation is in many respects a continuation of ST3 in that the soil tem-

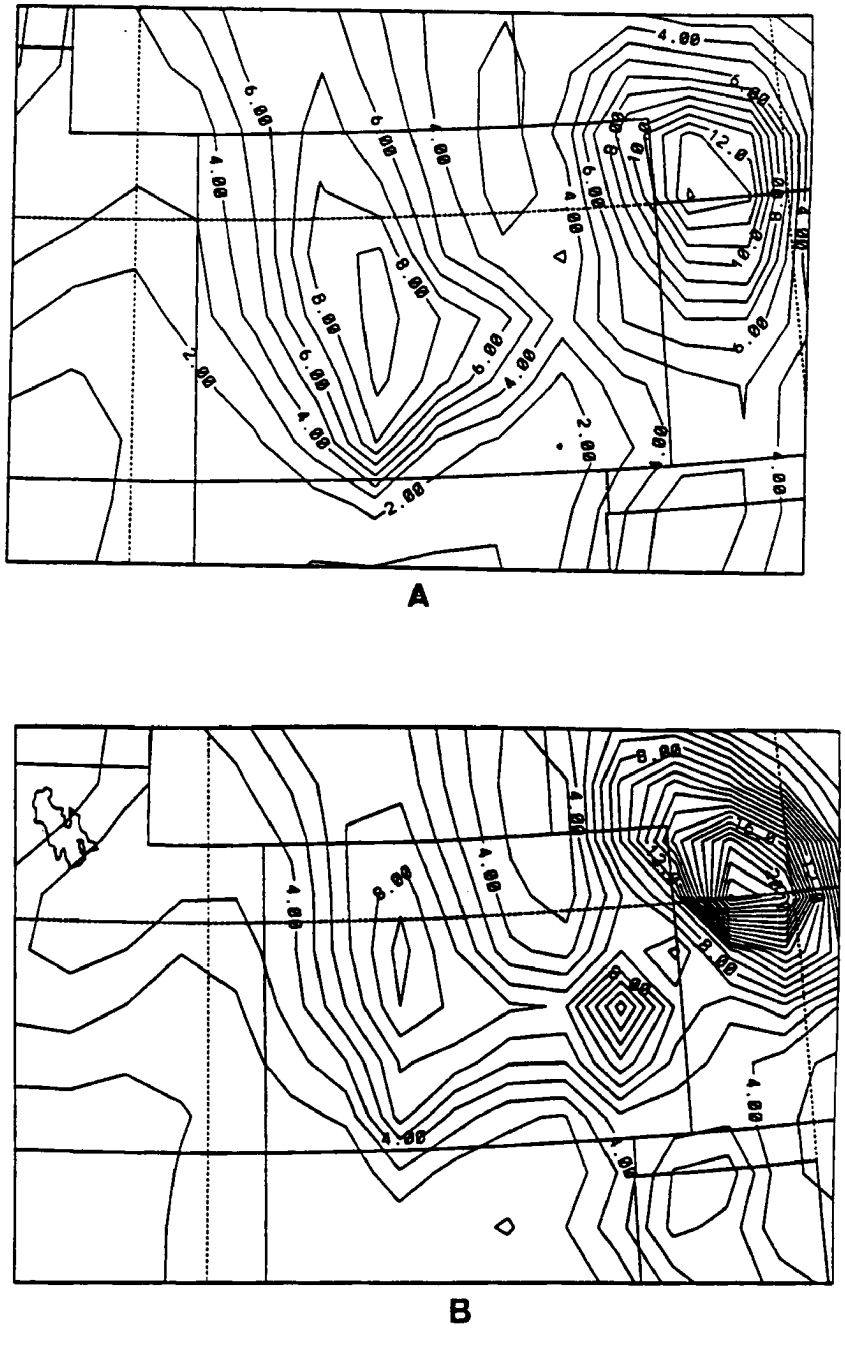
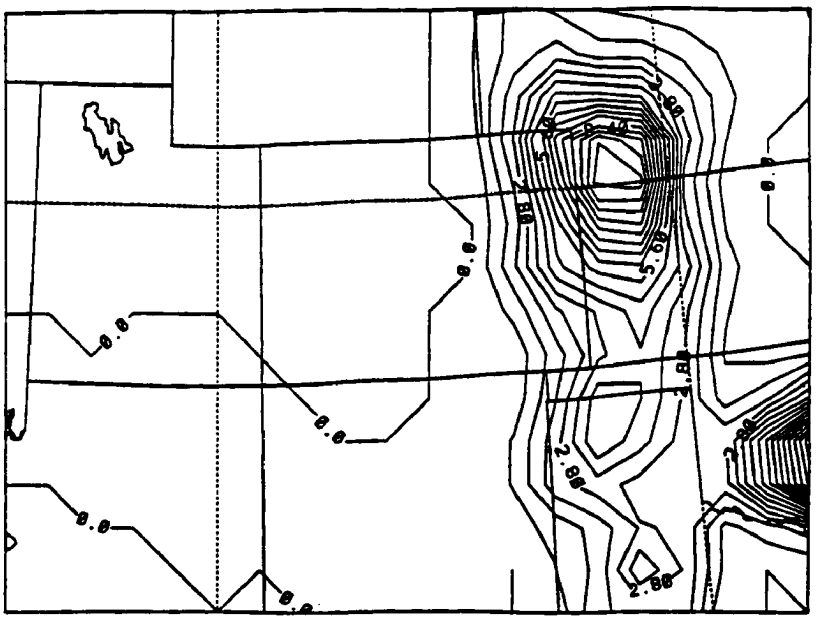


Figure 4.2: Total precipitation (mm) after 36 hrs of simulation at 1200 UTC 30 March for (a) Sensitivity 3, and (b) control run.



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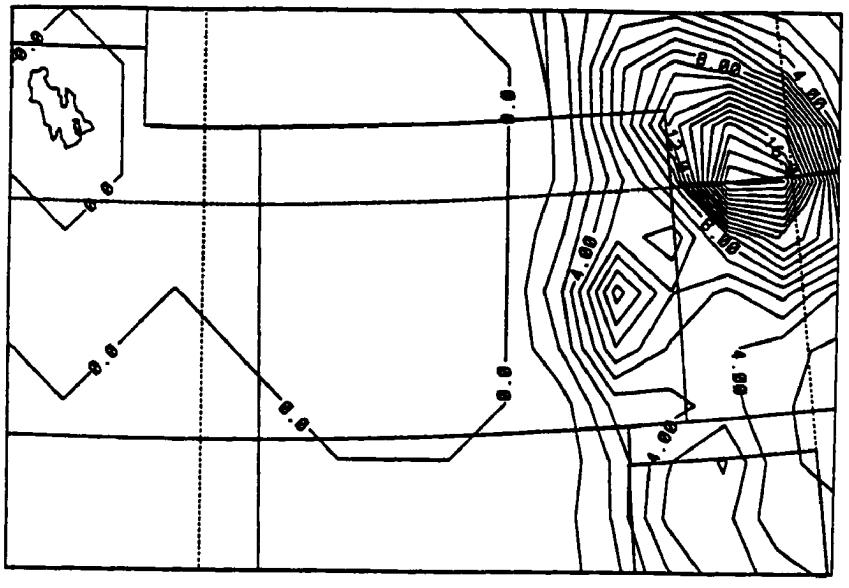

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Figure 4.3: Total rainfall (mm) after 36 hrs of simulation at 1200 UTC 30 March for (a) Sensitivity 3, and (b) control run.

perature offset was $-5°C$ for all soil levels; in addition ST4 was run with a reduction in the height of the mountains in the western half of the domain. This was accomplished by reducing the silhouette averaging from 1.0 in ST3 to 0.1 in ST4. Both sets of topography can be seen in Figure 4.4. The general pattern remains the same, the major difference being a reduction in height of the mountain barriers. The northern Rockies of British Columbia/Alberta are reduced in height from 2200 m to 1800 m, while in Colorado the barrier is lowered from 3200 m to 2600 m.

The low-level (200 m) wind vectors do show some differences between ST3 and ST4, especially at 12 hours (1200 UTC 29 March). The winds in ST4 are 10 to 15% stronger on the east side of the mountain barrier in Colorado (Figure 4.5). The winds over the barrier itself are identical despite a 600 m difference in height. The precipitation plots however do indicate that there was a 30 to 35% reduction in total precipitation over mountainous areas, with almost no change over western Nebraska and Kansas (Figures 4.6 and 4.7).

The reduction in total precipitation over the mountains (it also carried over to the Front Range) is due to a reduction in snowfall. Obviously the reduced barrier height did not allow as much condensate to condense and form snow as it did in ST3. This brings up the questions: how realistic does model topography have to be in order to get reliable output? The ideal situation is to have a near perfect terrain representation, but that is highly impractical for numerical models where some type of smoothing has to take place for model stability.

4.5 Sensitivity 5

In this simulation a second grid was added over the Colorado region. This fine grid had a 20 km grid spacing in both the x and y directions (Figure 4.8). In addition, the number of vertical levels in the atmosphere went from 16 in the control run to 29 in this simulation, all other parameters were the same as in the CR.

The most obvious difference between ST5 and CR is the topography representation which is shown in Figure 4.9 (compare with Figure 4.4). With a 20 km grid spacing over Colorado, the southern Rocky Mountains are much better resolved than the topography

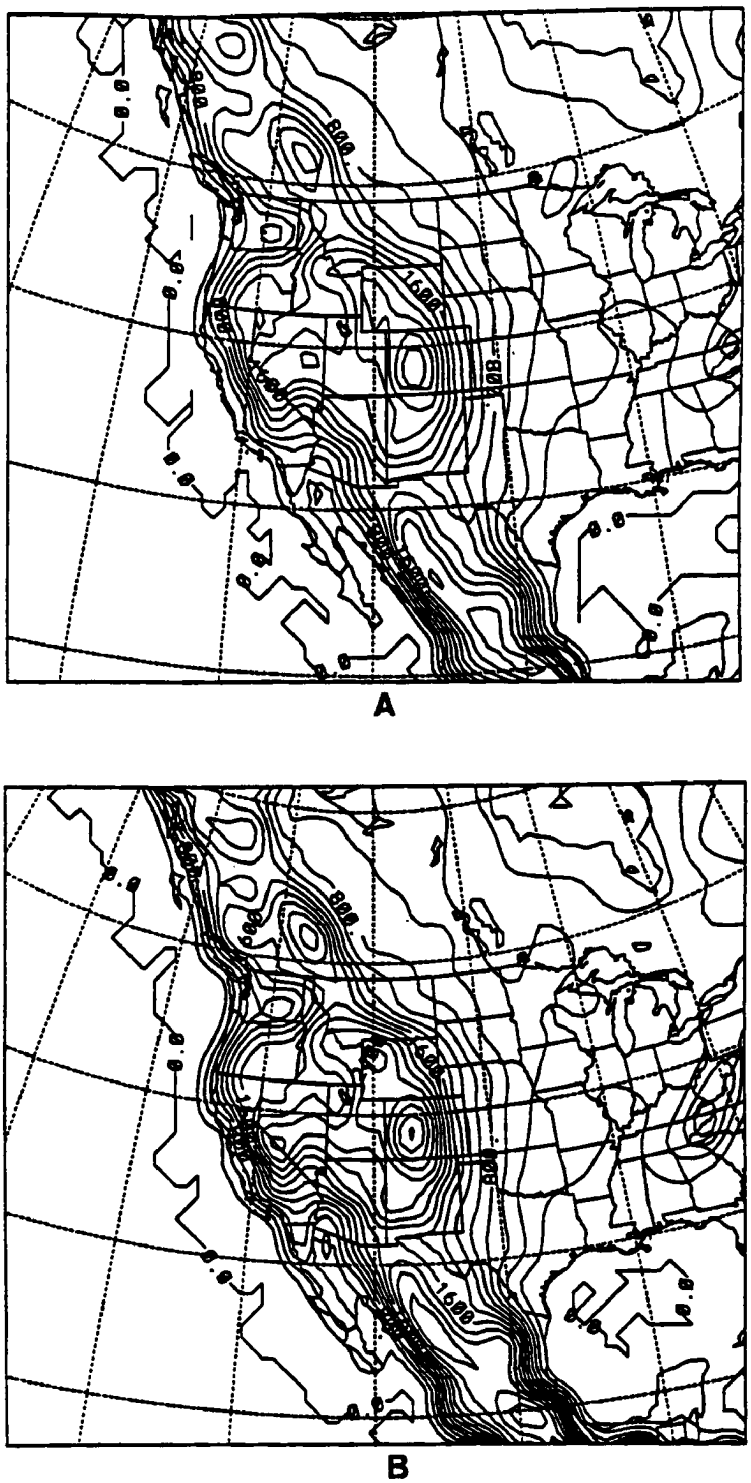


Figure 4.4: (a) ST4 topography, and (b) ST3 and CR topography (contour interval 200 m).

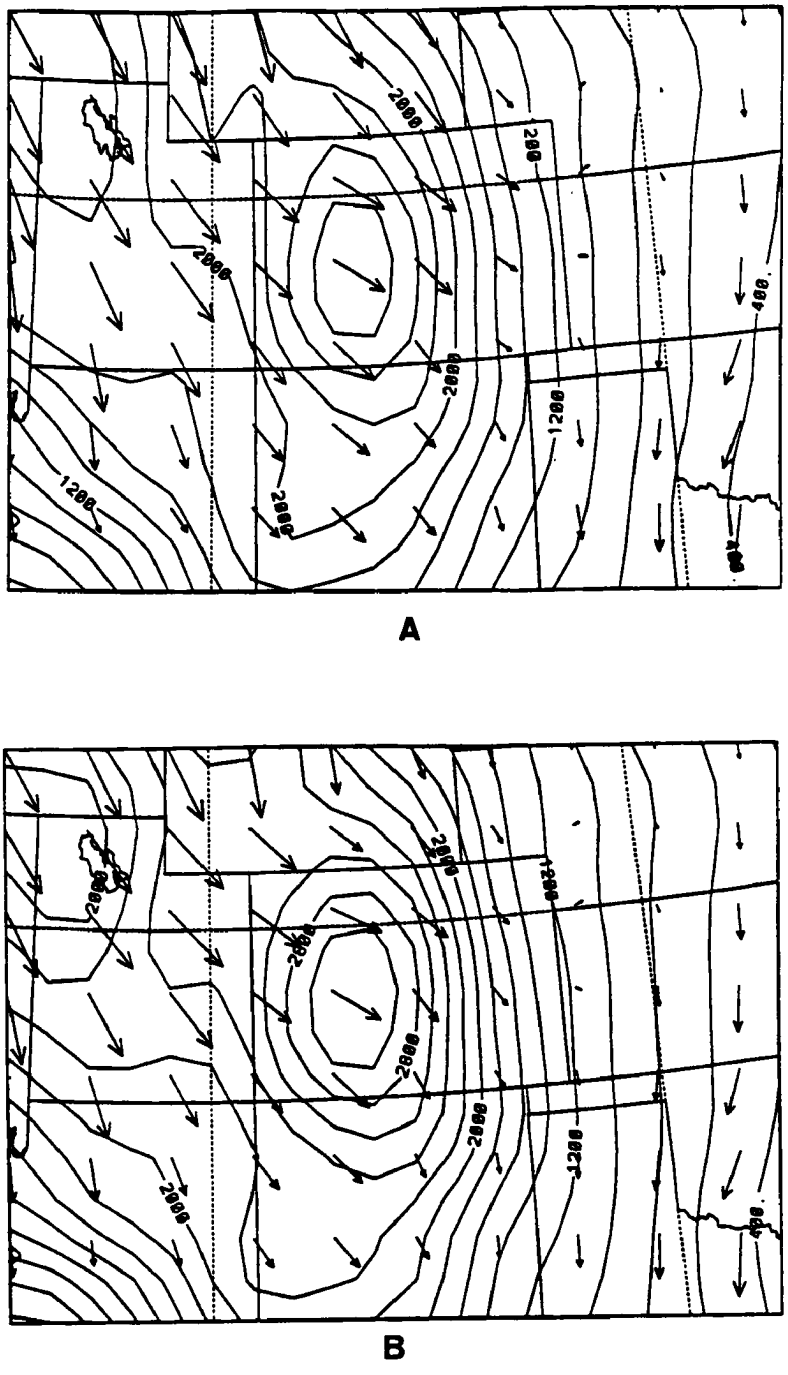


Figure 4.5: 200 m wind vectors at 1200 UTC 29 March for (a) ST4. and (b) ST3.

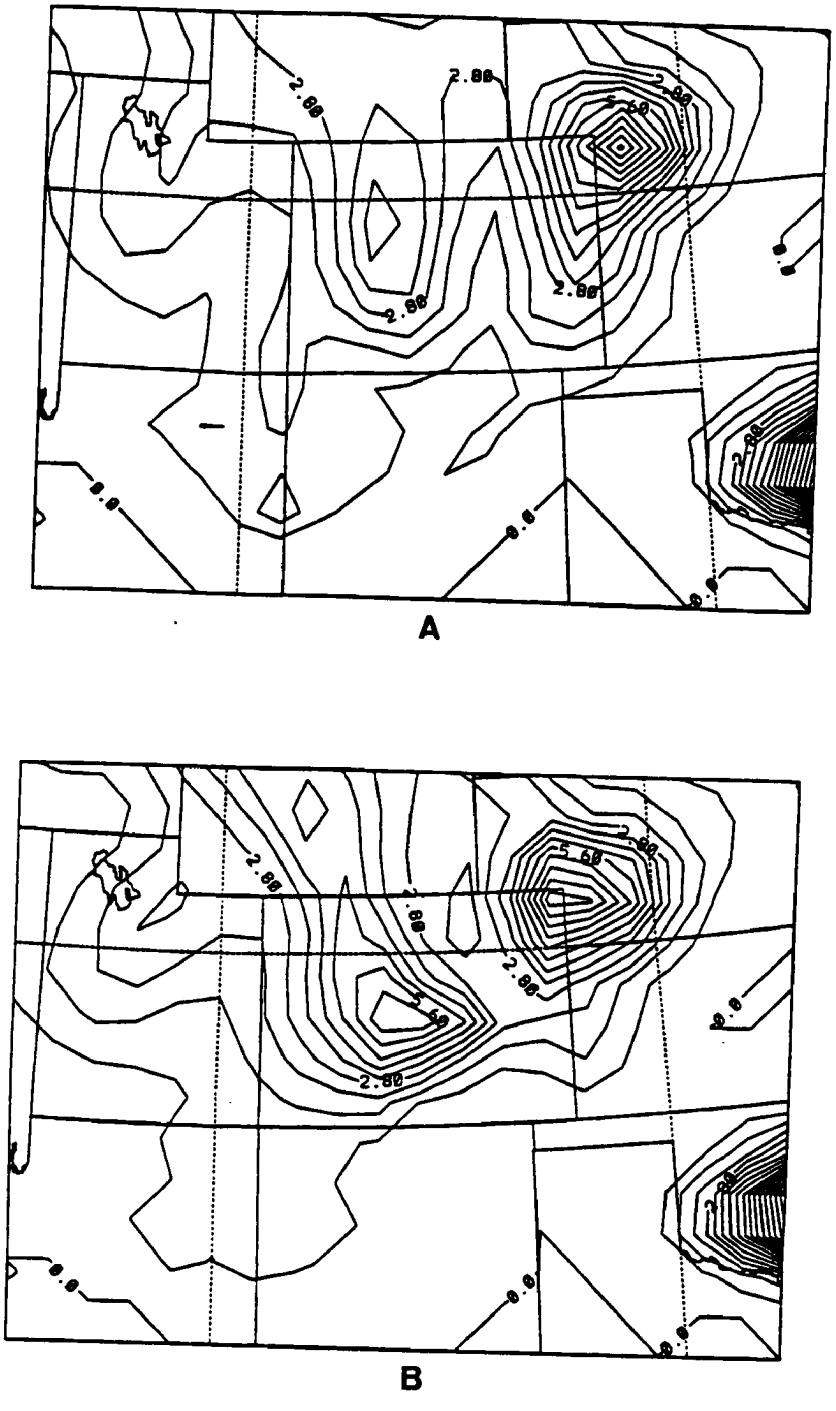
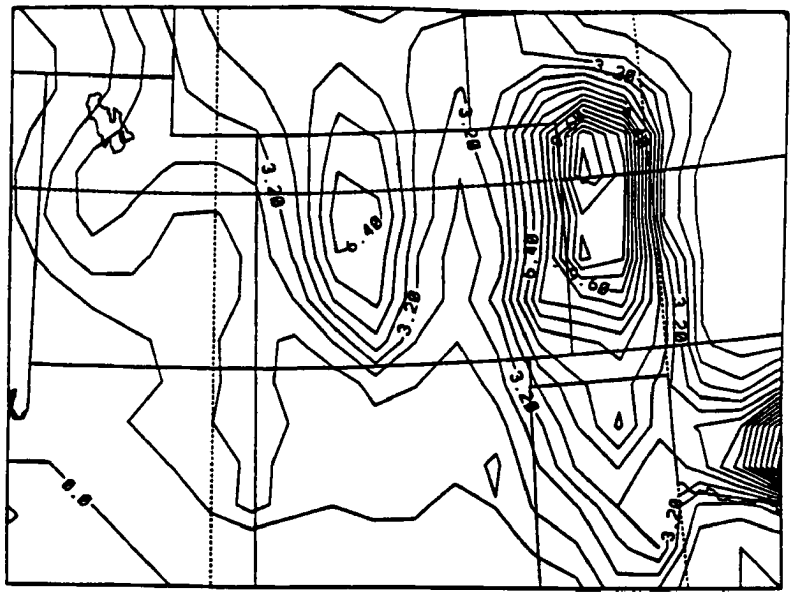


Figure 4.6: Total precipitation (mm) at 0000 UTC 30 March for (a) ST4, and (b) ST3 (contours 0.7 mm).



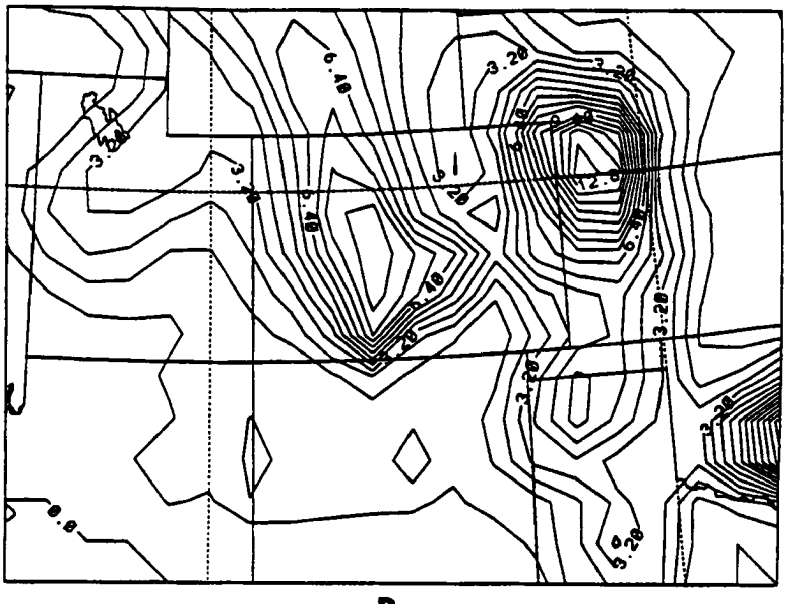


Figure 4.7: Total precipitation (mm) at 1200 UTC 30 March for (a) ST4, and (b) ST3 (contours 0.7 mm).

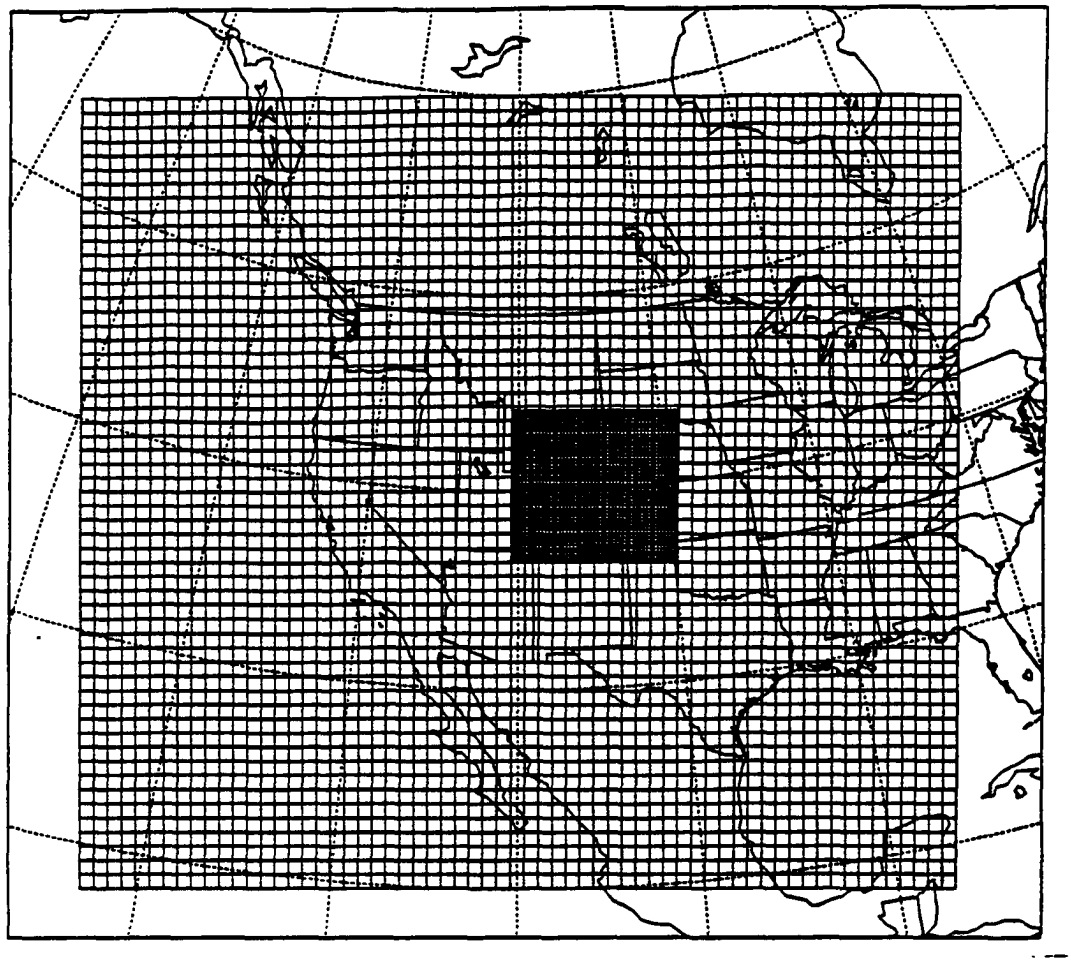


Figure 4.8: Nested grid domain used in ST5. Coarse grid 80 km × 80 km, fine grid 20 km × 20 km.

used in the CR, despite the fact that both simulations are using the same initial terrain data set.

Domain wide, the overall upper level winds are 10 to 20% weaker than in the CR. This is evident at 1200 UTC 29 March when the north/south winds (500 mb) over Memphis were 35 m s^{-1} in CR were 16 m s^{-1} in ST5, while the observations give a value around 40 m s^{-1}. Figure 4.10 shows the total precipitation for ST5 compared with CR. It is apparent that ST5 greatly underestimated precipitation amounts (mostly as rain) in the area over the central Appalachians (10% of CR values). The obvious question is why did this simulation do so poorly in this area despite an increase in the number of vertical levels? It appears from analysis of the wind fields and Exner function that in ST5 the shortwave which moved over the Appalachias at 0000 UTC 30 March was not properly resolved in this simulation. The central and southern Appalachias was an area with strong convergence of the 500 mb layer winds which was for some unknown reason better resolved in the CR.

Two other areas outside of the fine grid which received appreciable amounts of precipitation were coastal British Columbia and NE Quebec. The precipitation differences between ST5 and CR and relatively minor, totals being about 10% higher in ST5. This leads to the speculation that the fine grid upstream of the shortwave, which originated in Oklahoma, in someway affected the advection of this feature downstream. This could be due in part to the fact information is not being correctly passed on to the coarse grid from the fine grid. Although the 500 mb winds directly downwind of Colorado are roughly the same in the two simulations, this is an area of inquiry that requires further attention.

In the Colorado region (fine grid) the ST5 results are much more encouraging. Precipitation maxima which were present in the CR (Figure 3.11) over Nebraska/Kansas/Colorado border region has been eliminated in ST5 (Figure 4.11. The area of heaviest precipitation is in the Canon City, Walsenberg, Alamosa triangle, very close to an area of an observed maximum (Figure 1.4). The model tried to develop a band of heavy precipitation in the La Junta area of eastern Colorado, but the amounts are greatly underestimated. In the mountains, RAMS did very well in depicting precipitation patterns, as would be expected with better terrain representation. The Telluride

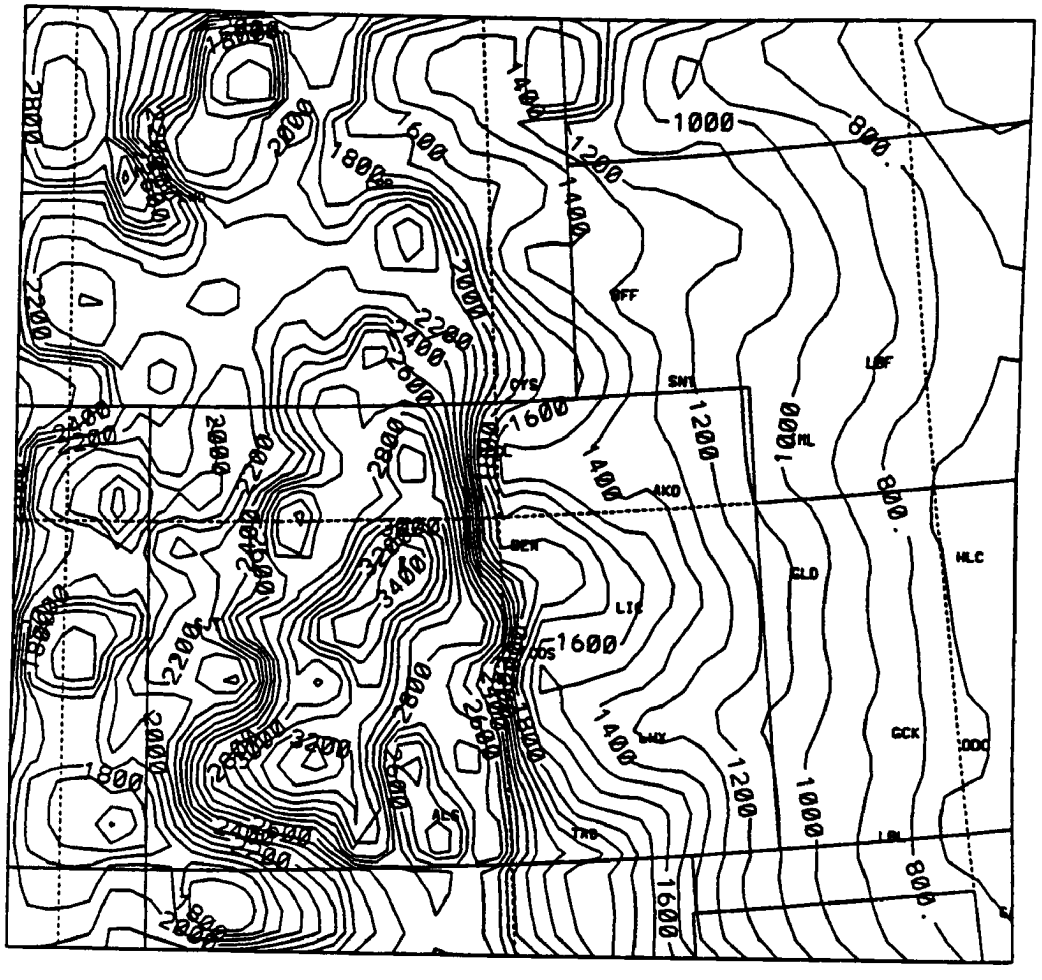


Figure 4.9: Topography within fine grid.

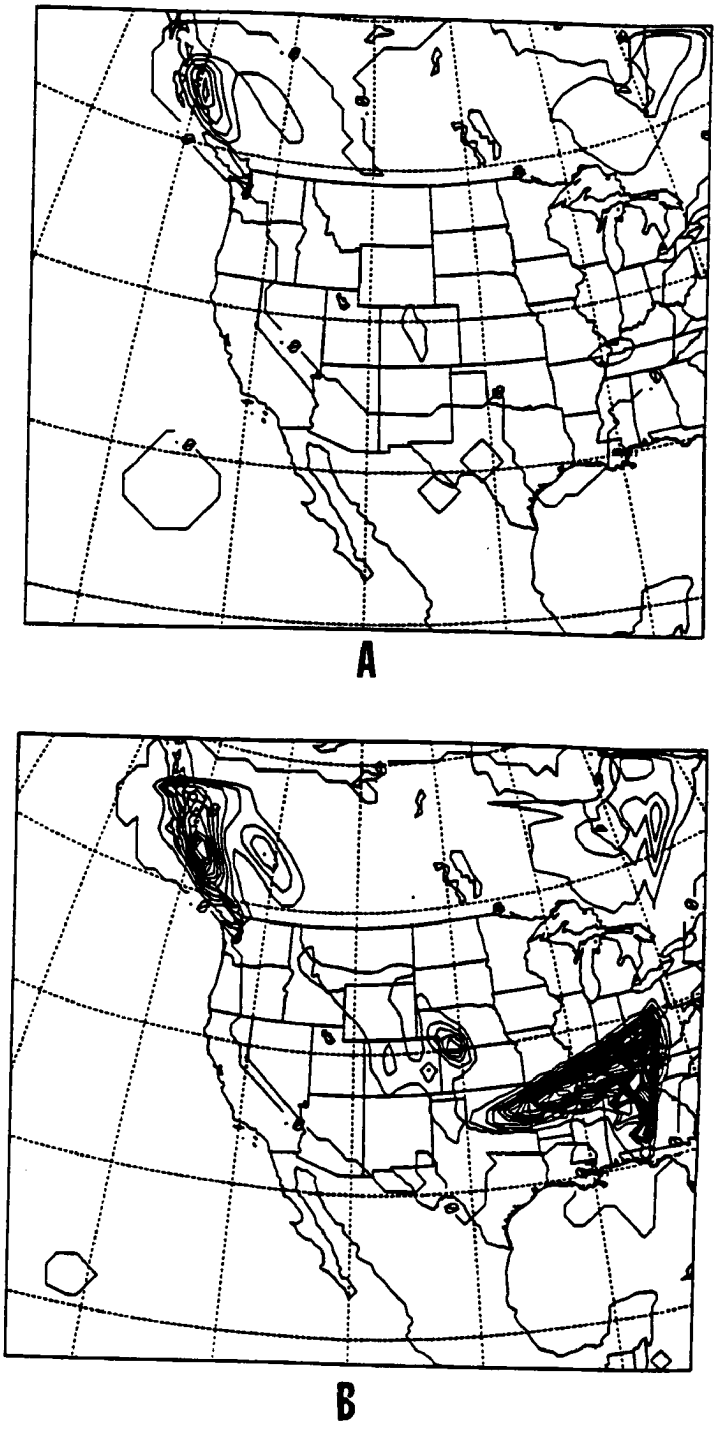


Figure 4.10: Total precipitation (mm) for (a) ST5 and (b) CR (contours 5 mm).

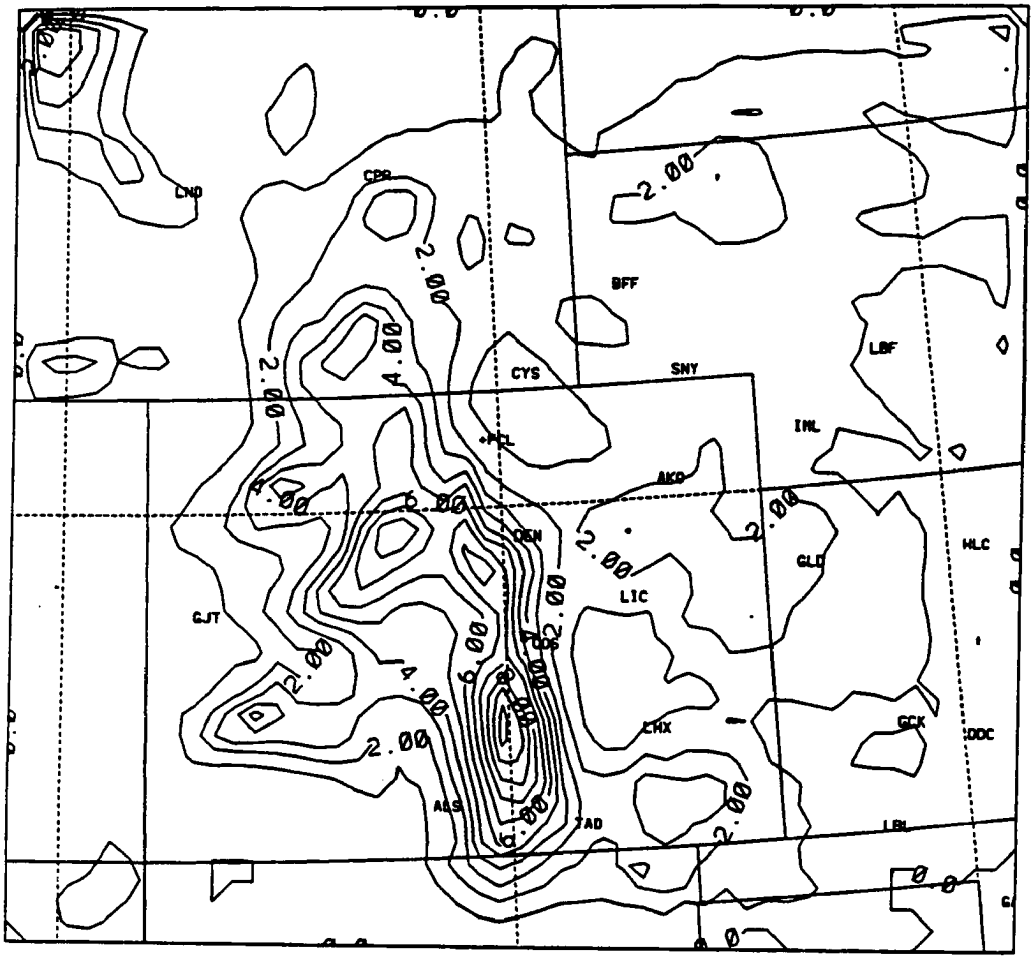


Figure 4.11: Total precipitation (mm) in the Colorado region (contours 1 mm).

and Mt. Evans/Winter Park maximas are clearly represented, although model snowfall is about 50% of the observations from these areas.

It is apparent from Figure 4.11 that the precipitation maxima west of Canon City was a result of the abrupt eastern edge of the Front Range. The 2000 meters of relief in this area caused the northeast winds (around 0000 UTC 30 March) to be forced upslope and help create fairly strong vertical ascents of 7 cm s^{-1}. This was the only area along the Front Range with persistent upslope flow. In this case there was no cold air damming; the low-level flow was decoupled from the upper-level flow and helped generate the vertical motion. This northeast flow also explains why there was a small area south of Trinidad which observed as much as 30 cm of snow during this storm. The flow was forced up the northern side of Raton Mesa where moisture condensed and fell as snow. In ST5 the vertical velocities were not large enough to cause large amounts of snowfall, probably as a result of the model not moving the low pressure into central New Mexico where it was located by the end of the simulation time.

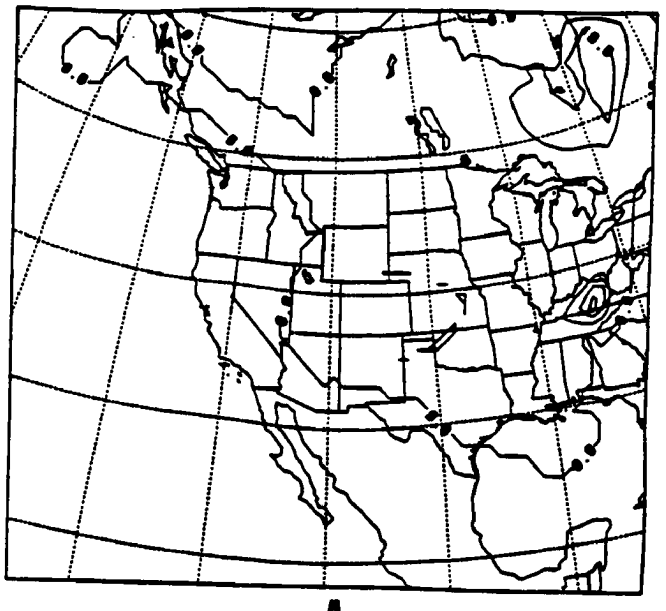
4.6 Sensitivity 6

In ST6 the height of the domain-wide topography was set equal to zero, in effect, the domain was entirely flat. At 5800 m (about 500 mb), the wind vectors have a much larger westerly component (more zonal) then in the CR. Closer examination of the east/west and north/south winds show flow patterns that are highly generalized, especially over the mountainous regions of the western USA. For example, in ST6 at 1200 UTC 30 March (36 hrs into the simulation) the winds at 5800 m over Colorado are westerly about 12 m s^{-1}. In the CR, the winds over Colorado are easterly in the western part of the state, and westerly over the eastern half. The perturbation in the Exner function indicates that a low never develops over Wyoming and Colorado like it does in the CR.

The result of a domain without any terrain is most apparent in plots of precipitation as shown in Figure 4.12. Over the southern Appalachias, the amount of total precipitation is about 10% of the CR. In addition, the areal coverage in ST6 is about half of the CR coverage. This is due to the fact that without the presence of the Appalachias, the low and

mid-level winds do not experience any convergence, with the result that vertical velocities are weak to non-existent, and heavy precipitation is never able to develop.

Surprisingly, in Quebec Province, ST6 produced a relatively equal amount of total precipitation as in the CR. In western British Columbia, where the Coast Range blocks storms moving in from the northern Pacific, precipitation decreased from 35 mm in the CR to 4 mm in ST6. IN Colorado there was only trace amounts of precipitation in ST6, the secondary maximum over the Nebraska/Kansas/Colorado border region which was present in the CR, was absent in this simulation. The role of the mountains on atmospheric flow and precipitation is obvious. The problem areas in ST6 were over the Appalachias, Coast Range, and the southern Rockies. In central Quebec, where there are no mountains, ST6 performed well. This reinforces the belief that accurate portrayal of terrain is crucial to these types of simulations.



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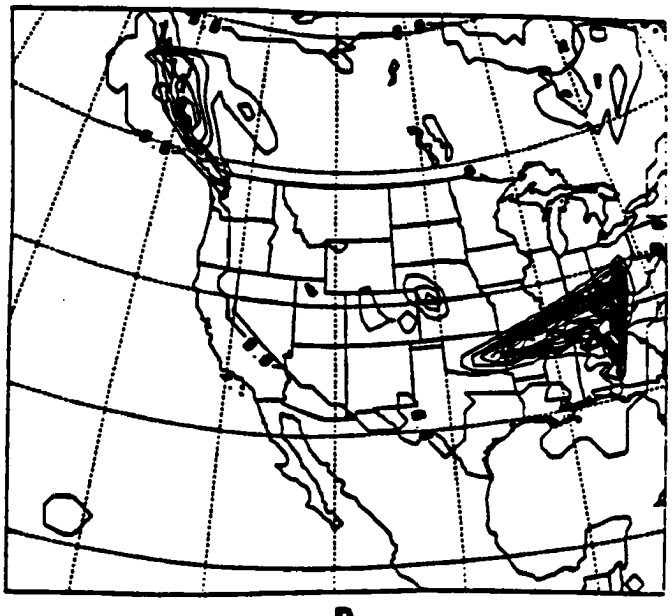

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Figure 4.12: Total precipitation at 1200 UTC 30 March for (a) ST6 and (b) CR (contours 5 mm).

Chapter 5

DISCUSSION AND CONCLUSIONS

This study has utilized the Regional Atmospheric Modeling System in order to better understand winter storms along the Colorado Front Range as well as investigate the model's response to various changes in its initialization.

In comparison with the Nested Grid Model for the 29-30 March 1991 storm, RAMS produced a relatively better 'forecast', with roughly similar grid structures. The differences between the two simulations include a difference in parameterization schemes, topography, and the 12-hour nudging in RAMS. The RAMS model, in this study as well as in Wesley's (1991) work, has proven that it can successfully simulate winter storms in Colorado's complex terrain, as well as in situations where low-level decoupled flow is pronounced.

In essence, RAMS was able to capture the areal distribution of precipitation patterns across the domain, although the amounts of precipitation were consistently less than observed amounts. One advantage of RAMS is the availability of model output at a variety of spatial scales. For example, the model user can specify the plotting routine in order to enhance the resolution, in effect, zeroing in on an area of interest. NGM's plotting is too crude to show much in the way of detail, hence an observer's perception of its output is downgraded.

Variations in RAMS initialization was carried out in six sensitivity studies. In the first one, ST1, it was found that precipitation amounts were sensitive to what species in the microphysics module are turned on or off. This is probably due to a change in the size distribution, hence evaporation and sublimation changes. It should be mentioned that since this simulation was conducted, modifications to the microphysics module have been made. Whether changes in precipitation totals still occur when one of the ice/water species is turned off, still remains to be tested.

In ST2 the nudging was reduced in the outer five grid zones. The model results were changed only slightly by this operation, confirming the expectation that RAMS is able to simulate the events of 29-30 March without the help of continued external input.

ST3 revealed the importance of proper initialization of soil temperatures. The results indicated substantial influences on precipitation totals. I would generalize this result to include soil moisture and surface albedo (due to changes in the vegetative cover) as well. The timescale of a simulation becomes important in regards to these surface fluxes. Longer simulations (>72 hours), require more accurate surface initialization than shorter simulations since the response time for some of these processes (i.e., movement of soil water, thermal waves in the soil), are on the order of several days to weeks.

In ST4 it became apparent that accurate representation of terrain features is crucial to simulations which take place in complex terrain. In order for any model to properly simulate low-level decoupled flow and terrain-forced ascent, the terrain, of course, has to be as realistic as possible.

With the addition of a fine grid over Colorado in ST5, RAMS was able to produce a simulation that was in better agreement with the observations than in the Control Run which lacked this grid. This is due in some degree to better representation of the mountainous regions within the state as well as to the inclusion of 13 more vertical levels. However, this simulation was not without its problems. Downwind of the fine grid, there was a drastic reduction in precipitation totals. What role the fine grid had in all this is not know at this time.

As would be expected, the influence of mountain barriers is of extreme importance to atmospheric flow and precipitation processes. ST6 was conducted over a domain that consisted of a perfectly flat land surface.

5.1 Suggestions for Future Research

In the context of what has been presented in this study, further modeling studies could attempt to study/explain the following questions.

1. How do surface fluxes change when there is a thin snow cover present?

2. Is RAMS able to resolve elongated snowbands that propagate through the Front Range?

3. At very high resolutions (grid spacing of several kilometers), can RAMS accurately produce the spatial variability of snowfall seen so often in Front Range winter storms?

4. What influence, if any, does a nested grid produce on the downstream portions of the domain?

Some of these suggestions are already being investigated by interested parties. The influence of a nested grid should be investigated with a variety of weather scenarios, as well as at different spatial scales. To the author's knowledge, no one has yet to determine at what point in a simulation surface fluxes become crucial. This will of course depend on the length of the simulation as well as its spatial extent. At present, soil temperature profiles, soil moisture contents, and surface albedo, and snow cover depth, are in many simulations chosen as best guess scenarios. In simulations that, for example, forecast the release some toxic substance, the types of surface properties mentioned above could be critical for accurate simulations.

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