THE INFLUENCE OF OROGRAPHY ON AN EAST COAST STORM, AS SIMULATED BY THE NCAR GCM

by

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Submitted to the Department of Meteorology on May 12, 1978, in partial fulfillment of the requirements for the degree of Master of Science.

ABSTRACT

A high resolution, limited area version of the NCAR GCM is used to simulate the March 14, 1975 east coast storm. Based on climatological data, synoptic experience, and an examination of numerous cases, this cyclone was chosen because it seemed to have been strongly influenced by the Appalachian Mountains. The storm was investigated in detail and is briefly described.

The NCAR GCM is also described briefly. Two simulations are made with the 1.25° resolution, 24 layer version, the first simulation including orography and the second incorporating a uniform land elevation of 10 meters above sea level.

Neither of the 36 hour forecasts exhibited exceptional skill in this situation. It is doubtful that the simulations could be of much use in analyzing the more important aspects of the observed storm. However, the model did appear to be sensitive to the inclusion of orography, producing a ridge-trough couplet east and west of the mountains, respectively, when orography was included. Although this perturbation "improved" the forecast, the observed couplet is due more to the passage of the primary center west of the mountains, whereas the model perturbation is believed due to an upslope/downslope divergence/convergence effect.

A cold perturbation in the temperature field also resulted from the incorporation of orography. Lying east of the mountains at low levels, this perturbation is believed due to cooling accompanying orographically forced upward motion.

Precipitation differences between the two cases were, in general fairly small. Direct orographic forcing, while perceptible, was judged to be of significantly lesser importance than the indirect effect of the mountains in altering the storm track and larger scale circulations.
The method of initializing the moisture field (zonal climatological mean) is believed to be the major reason for many of the faults in the forecasts, particularly with regard to overall precipitation and deepening. This aspect of the model probably deserves at least as much attention as the orography in any future sensitivity experiments.

Lastly, correlations amongst several model variables are made and examined versus time and versus height. Changes between the two cases are minimal. One of the more interesting aspects of these results is the tendency, in both cases, for precipitation to be more highly correlated with condensational heating at upper levels early in the storm and become more evenly distributed as time progresses. This is likely another symptom of the poor moisture initialization.

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I. INTRODUCTION

Of considerable concern to East Coast forecasters are the frequent coastal storms and their associated precipitation that plague this region during the winter and early spring. It is often very difficult to predict not only the initial development, but the motion and/or redevelopment of the storm. Forecasting a future pressure pattern is not particularly useful in and of itself, but information about associated winds and precipitation, both of which are frequently of damaging intensity, is of obvious value. And while the Appalachians are, by no means, a major mountain range, it has been recognized for some time that they exert a strong effect on these storms (e.g., Petterssen, 1941 and Miller, 1946).

Several studies have been made in determining the climatology of North American cyclones, such as Hosler and Gamage (1956), Reitan (1974), and Colucci (1976). But, while such studies can be of interest in analyzing the general features of these storms, they are of little predictive value on a case-by-case basis. Statistical predictive value models such as the one developed by Veigas and Ostby (1963), while displaying some skill, yield no insights into the dynamics of the cyclones. Today's large numerical models, however, with their ever greater resolution and capability for handling a vast amount of data on a large number of variables, offer a possibility of investigating the factors influencing the behavior of these storms. But, in this case, one must be constantly aware that the results represent a model storm and not the real atmosphere.

This study was prompted by a combination of personal interest in coastal storms, introduction and access to the NCAR General Circulation
Model, and the results of Colucci's (1976) study. His statistical analysis of ten years of coastal storms was done on much finer time and space scales than previous works. A domain covering the eastern U.S. and adjacent western Atlantic was divided into 375 1° squares and surface analyses made at 3-hourly intervals were examined. The results show a distinct minimum in cyclone frequency along the Appalachian Mountains and the eastern foothills, with a weak maximum to the west, a belt of very high frequency from South Carolina northeastward to the Gulf of Maine, and a secondary belt of high frequency east-northeastward from Cape Hatteras out over the northern boundary of the Gulf Stream. The detail displayed by these frequency distributions was heretofore unresolvable in larger scale studies (e.g.; Reitan, 1974). An even more interesting aspect of these storms appeared in an analysis of the percentage of storms that deepened over a particular area. Colucci finds that an axis of maximum percentage of deepening storms occurs on or near the axis of lowest cyclone frequency. He concludes that the few cyclones that do track along or through the mountainous areas almost always deepen, i.e., the upper level support must be not only strong enough to maintain the surface cyclone as it passes through the mountains; but also strong enough to cause a deepening of the storm during the process. A troublesome aspect of this interpretation is the lack of a middle ground between the weakly supported cyclones which cannot be maintained in passing through the mountains and the strongly supported storms which deepen as they move over the topography. In actuality, one would expect to find cyclones intermediate between these two extremes, which could maintain their
identity while passing over the mountains without necessarily deepening. This experiment is both an attempt to simulate a storm crossing the Appalachian area and a sensitivity test of the relatively new 24-layer, 1.25° horizontal resolution (in both the latitudinal and longitudinal direction) limited area version of the NCAR General Circulation Model. The model itself is described in greater detail below. Two forecasts were made from the same initial data, one forecast incorporating orography (the "mountain" case, designated M) and the second with a uniform land elevation of 10 m (the "no-mountain" case, designated NM). The specific objectives were (1) to test the model's ability to make a real-data forecast with and without orography, (2) to determine to what degree the model is sensitive to the inclusion of orography, and (3) to make inferences about the effect of orography on the real atmosphere.
II. THE MODEL

The model used for this study is the Limited Area Version of the NCAR GCM (LAM). The LAM is basically a high resolution, refined version of the original NCAR model described by Kasahara and Washington (1967) and updated and summarized by Oliger et al (1970). The reader is referred to these papers concerning details of the model. Only the more general and/or more pertinent features will be described here.

Table 1 lists the general characteristics of the second generation NCAR GCM as they apply to the LAM used in this experiment (taken partly from a report of the GCM Steering Committee, 1975). Some of these require a fuller description and are elaborated on below.

Figure 1 illustrates the vertical grid structure and the placement of variables. The lower boundary level, s, is placed at anemometer height, approximately 10 m above the ground surface, H. The n levels are always horizontal and are placed at integral numbers of 750 m. The m-levels are placed halfway between n-levels except in the case where that layer is intruded upon by orography. In that case, the m-level is defined to be the level halfway between the s-level and the next higher n-level. More complications arise when the orography completely pierces a layer, resulting in a "blocked" point. The m-level in that layer is assumed to end when it meets the s-level. Variables are not explicitly carried at the blocked points. Values are rudimentarily retrieved by simple extrapolation into the mountain if required by a particular calculation in the model. However, in the analyses conducted for this paper, the points were considered blocked and no values were assigned to or assumed at those points.
The model orography for the area of interest in this study is shown in figure 2. This is obviously quite smoothed, partly due to resolution restrictions and partly for the sake of numerical stability.

The convective adjustment procedure used in this experiment was of the same type used in most previous experiments with the NCAR GCM. An adjustment is made subject to the following constraints:

1. The internal energy remains constant between layers:
\[ \int \rho T \, dz = \text{constant} \]

2. The new lapse rate equals the appropriate critical value if the computed lapse rate is in excess of that value:
\[ \frac{\partial T}{\partial z} = -\gamma_d \text{ if } w < 0 \quad \text{or} \quad \frac{\partial T}{\partial z} = -\gamma_m \text{ if } w > 0, \]
where \( \gamma_d \) and \( \gamma_m \) are the dry and moist adiabatic lapse rates, respectively, and \( w \) is the vertical component of motion.

3. A mean temperature is defined (constant throughout the adjustment process),
\[ \bar{T} = \frac{\int \rho T \, dz}{\int \rho \, dz} \]
which is used to compute \( \gamma_w \).

Working upward from the lowest two layers, the process essentially cools the lower layer of each pair and warms the upper layer until a pair of layers is reached which is stable. Pressures at the \( n \) levels are then recalculated hydrostatically assuming the surface pressure is unchanged. "Convective" precipitation is not calculated explicitly in the model. However, if a net cooling occurs in a layer as a result of the convective adjustment, this will be reflected as an increase in the relative humidity of the layer. If, during the subsequent time step, the humidity exceeds the specified maximum allowable relative
humidity (95%), the excess is rained out as "stable" precipitation.

The most obvious and serious shortcoming of this convective adjust-
ment procedure is the absence of any saturation criterion in determi-
ning whether or not to made an adjustment. The implications of this
defect for forecast skill are not at all obvious. But one can ima-
gine, for example, the abrupt change possible in the model when descent
gives way to ascent, no matter how minor the motions.

Perhaps the major fault of the model in this simulation was the
initialization of the moisture field. Relative humidities in the
surface to 6 km layer were rather uniform, varying from about 45% at
25°N to about 60% at 50°N, with the gradient lying almost entirely
in the north-south direction. As will be seen later, this does not
represent the actual humidity field very well at all. At present,
several other schemes for initializing moisture are being pursued by
the NWP group at NCAR.

The initial data used for this experiment were acquired from NMC
analyses stored on magnetic tape and archived at NCAR (Jenne 1975).
These analyses included wind and geopotential height data at the ten
mandatory pressure levels, analyzed using the Flattery technique on
the NMC octagonal grid (Schuman and Hovermale, 1968), which has a
horizontal resolution of 381 km at 60°N. These data were, in turn,
interpolated to the NCAR six layer grid. Next, temperatures at the
midpoints of each of the six layers were computed hydrostatically.
A polynomial curve was fit to these six values to obtain values at
the 24 equally spaced points in the vertical corresponding to the
mid points of the 24 layers. Pressures at the remaining levels could
then be computed hydrostatically using these temperatures. To eliminate any gravity waves induced by the above procedure, a kind of filter was applied which removed the column mean divergence from the data (Washington and Baumhefner, 1975). (Actually, very little divergence was introduced in the interpolation, but its removal constituted a reasonable precaution.)

Lastly, the criteria for precipitation are as follows:

(1) relative humidity in excess of 95%, and

(2) \( w > 0 \) (upward motion).
III. SYNOPTIC SITUATION

Several years of maps and numerous storms were examined in search of the storm which best exhibited an influence from the Appalachian Mountains. The storm of March 14, 1975 was eventually chosen.

Teletype data archived at MIT were used to construct maps of sea-level pressure and surface temperature at 3 hour intervals, as well as upper air maps at 12 hour intervals. Both are discussed in detail later in the text. In addition, hourly precipitation data for approximately 800 stations east of the Mississippi River, obtained from the National Climatic Center in Asheville, N.C., were examined and plotted as three hour amounts corresponding to the intervals of surface plots.

While the original intention was to make a 24 hour forecast from 0000 GMT on the 14th, the NMC data for this period was missing from the NCAR archives. It was rather fortuitous that we were forced to begin the simulation 12 hours earlier, at 1200 GMT on the 13th, and make a 36 hour forecast. Any precipitation forecasts during the 24 hour period beginning at 0000 GMT the 14th, the most interesting and important period, would have severely underestimated the amounts had the forecast itself begun at 0000 GMT the 14th. This is because the moisture field, initialized at climatological values, required several hours to become quasi-realistic. The actual moisture distribution at 1200 GMT the 13th, shown in Figure 3 (taken from NAFAX chart No. 87) can be seen to deviate significantly from the monthly zonal mean described above.

Figures 4a and 4b show the initial distributions of surface pressure and temperature as analyzed by hand and as initialized in the model,
respectively. A trough runs from a fairly deep storm in Quebec southwestward over the mountains to a developing low center off the Louisiana coast. A cold high is centered over southwest Minnesota and a warm high is centered southeast of Bermuda. The most significant differences between the actual and model-initial fields are (1) an insufficient depth of the Gulf low, (2) an insufficiently strong temperature gradient in Mississippi, Louisiana, and eastern Texas, and (3) an insufficient rate of decrease of temperature to the north, becoming particularly evident in the upper midwest and southern Canada. Temperatures in general are too warm, largely a consequence of the extrapolation method of initializing temperature at the s-level.

Ocean surface temperatures are the same as the s-level temperatures over water as described in table 1.

Surface analyses for succeeding time periods are shown in figures 5a-10a, as mentioned earlier. While analyses were made every three hours, only those periods exhibiting synoptically significant features or change will be presented, for the sake of conciseness.

By 0000 GMT on the 14th (figure 5a), the low center in the Gulf region has moved northeastward to central Alabama and deepened by 3 mb. An extension of a trough toward the northeast has become more evident, but the temperature gradient there has not become very strong yet. The high over the upper midwest has moved east and become more elongated along a curved line east of Wisconsin, southwest through Iowa and southward through Oklahoma and eastern Texas. The very cold air over northern Minnesota has disappeared, indicating that it was probably a result of radiational cooling the previous night. An
apparent cold front lies southwest from the low center through south-
easternmost Louisiana. Showers and thunderstorms are occurring in
a belt from southeastern Tennessee to the Florida panhandle. Hourly
reports indicate that precipitation of a more stable nature is taking
place in a band about 200 km wide from southern New Jersey and Dela-
ward westward through northern Virginia and into eastern Kentucky
and Tennessee. More showery precipitation is occurring in southeastern
and south central Virginia and northeastern and north central North
Carolina. At this time, none of the precipitation is frozen.

Figure 6a shows the surface conditions for 0600 GMT on the 14th.
The low center has moved into eastern Tennessee and filled by 1 mb.
The high pressure center to the north is now positioned just north of
Lake Ontario. But it is the smaller scale, almost mesoscale, varia-
tions on these larger systems that are the outstanding features at
this time. An "inverted trough" extends from the low center north-
northeastward along the western slopes of the mountains. This, coupled
with an "inverted ridge" developing east of the mountains, is creating
a very strong pressure gradient over the mountains. The trough
previously noted over North Carolina is intensifying on what is becoming
a very strong temperature constast there - on the order of 20°F
over a horizontal distance of only 120 km. This contrast has appeared
primarily because surface temperatures have dropped 10°F over Virginia
and northern North Carolina, due to cold advection there, while southerly
winds blowing in from over the Gulf Stream have maintained, and even
enhanced, the higher temperatures over south Carolina and southern
North Carolina.
To the west an active cold front is still present from the low center south through the Florida panhandle. Showers and thundershowers associated with this front are occurring in a belt from Columbia, South Carolina, through Macon, Georgia, to Panama City, Florida. Showers are also associated with the front over North Carolina, reaching moderate intensity in the area of Greensboro, North Carolina. Light precipitation is falling in the central and western Ohio Valley and over Delaware, Maryland, and southern New Jersey. Somewhat heavier amounts are falling in the Virginia foothills. A map of the observed 3 hour precipitation totals ending 0600 GMT is shown in figure 11a. (The contours are not labeled in succeeding maps because of the intricacy and steep gradients in the observed patterns.) The heavy cold frontal precipitation is evident over Georgia, with a large band of amounts in excess of 0.64 inches and some areas receiving over 1.28 inches in three hours. Amounts in excess of 0.32 inches were received over the western three quarters of Virginia, with heavier amounts falling in the foothills of Virginia and North Carolina, reaching intensities of over one inch per hour at some points. Evidently, moist air from the east and southeast is being lifted in this region, in part by (1) orographic forcing (particularly evident in the west central portion of Virginia), (2) by being lifted over the (shallow) cold dome forming in this region, and (3) by synoptic scale ascent ahead of the storm.

The 1200 GMT map (figure 7a) shows the center to have moved only slightly northeast and deepened by only 1 mb. There are, however, some notable developments. First, pressures have fallen to the north
of the low center and a cyclonic circulation has become more pronounced in the entire southern Great Lakes area. Second, the trough over central North Carolina has deepened considerably while pressures over Alabama and southwestern Georgia have risen just as dramatically. The low center now has an extension of low pressure around the southern tip of the mountains, joining with the trough extending east-north-eastward over Cape Hatteras. The front in North Carolina is now at its greatest intensity, having strengthened to about 20°F over a horizontal distance of only 50 km. The strengthening now is primarily a result of cold air spreading southward east of the mountains.

The cold front to the south shows signs of weakening as both the temperature gradient and trough are decreasing in intensity. The high pressure to the north is now centered about 200 km north of Vermont but has a long curving axis to the west through Wisconsin, Iowa, and south to eastern Texas. The inverted ridge over the middle Atlantic states is not an obvious perturbation of the high to the north. This shortwave feature is absent north of Allentown, Pennsylvania, but is very evident to the southwest. It is not reflected at all in the 850 mb analysis, but its strength at the surface is not entirely supported by the surface temperature field. Some possible explanations for this feature will be discussed later.

Precipitation totals for the three hours ending 1200 GMT are shown in figure 12a. Precipitation along the cold front has subsided between 0900 and 1200 GMT (although 0600 to 0900 GMT amounts were still running over 0.64 inches in parts of Georgia and South Carolina). The heaviest precipitation is now occurring in two bands. The first is associated
with the trough developing rapidly to the east in northern South Carolina and south central North Carolina. It is likely that latent heat release is a contributing factor in the development of the trough in this area. More heavy amounts are falling in a second band in the eastern foothills of North Carolina and southern Virginia. The lighter precipitation in western Tennessee noted previously has spread north and east, with heavier amounts, mostly frozen, falling over a larger area. A band of snow, sleet, and freezing rain has broken out over the southern half of Pennsylvania, separate from the areas of heavy precipitation to the south.

The surface analysis for 1500 GMT is presented in figure 8a. This marks the first appearance of a second closed circulation, identifiable in both the wind and pressure fields, now located over south central North Carolina. Its central pressure is only 1005 mb and that of the primary is now 1003 mb. The secondary owes its existence to rapidly rising pressures over western North Carolina and northwestern South Carolina rather than local deepening. The persistent front runs through and is strongest at the center of the secondary, now amounting to 30°F over about 100 km. The heaviest three hour precipitation totals at this time (not shown) are falling along this front, with a narrow band of over 0.64 inches centered right along the front. The cold frontal activity seems to have been rejuvenated somewhat in Georgia and Florida with isolated "bulls-eyes" of heavy showers totaling up to 0.3 inches lying on a line from central South Carolina to central Florida. The bulk of the storm's precipitation, however, is falling in the area of the inverted ridge, evidently enhanced by flow
over the cold dome and the orography. A nearly separate area of rain, sleet, and snow is falling over West Virginia and southeast Ohio in association with the primary center which is tracking north northeastward along the western slopes of the Appalachians, consistent with quasigeostrophic theory which shows that cyclones in general will favor such a track. Also, eastward progress of the cyclone is discouraged by the dissipative effect of the orography and by the higher stability of the lower layers east of the mountains. The latter will become more evident in a later section displaying vertical cross sections.

A third area of snow and freezing rain persists over southern Indiana and western Ohio, slightly northeast of the 1200 GMT position.

High pressure has become nearly stationary over northern New England, merging on its southern flank with the ridge over the middle Atlantic states. Pressures are rising at a moderate pace over the lower midwest and Gulf states.

By 1800 GMT (figure 9a), the secondary is as deep as the primary at 1004 mb, but shows a preference for northeastward extension rather than deepening. This extension is occurring along the same strong front which runs through North Carolina and shows no sign of weakening. The heaviest precipitation (figure 13a), now generally less than one-half inch in three hours, lies in and north of this developing trough, while the upslope precipitation has tapered off considerably. Significant three hour amounts continue to fall in eastern Ohio and western Pennsylvania as the primary, held "blocked" by the mountains, begins to fill. Light snow is also continuing over western Ohio. A new center of high pressure is evolving over eastern Kansas and Oklahoma.
Development of this high is primarily toward the southeast with the large area of cyclonic circulation over the Ohio Valley being affected very little from without, although filling at the center is reducing the intensity from within.

In the succeeding six hours, the primary storm has filled another 5 mb. (see figure 10a) and has lost its closed circulation. In fact, pressures have risen throughout the entire Ohio Valley although the character of the pattern is still cyclonic. Most of the precipitation at this time is in the form of light snow or rain. Much of it is very spotty, especially in and around the mountains (see figure 14a).

The sprawling, rather bland low center in North Carolina at 1800 GMT is now better organized about 200 km offshore, having deepened 2 mb. The front in North Carolina, stationary until 1800 GMT, has now weakened and started to move southeastward, merging with the remnants of the old cold front now off the coasts of Florida and Georgia. The high pressure center north of New England has moved east of Maine and weakened slightly. High pressure in the midwest is spreading eastward through Tennessee, Kentucky, and the northern Gulf states, but is yielding in eastern Texas and over the Great Lakes.

The initial 500 mb height and temperature fields for 1200 GMT on the 13th are shown in figure 15. The hand analysis and the model initial field are virtually indistinguishable and hence only the analysis is shown. Strong west-southwesterly flow is the main feature over the east. Confluence is present over the midwest, a ridge dominates the northern plains states, and a shortwave trough is moving east into Texas. Figures 16a through 18a show the observed 500 mb pattern for
the subsequent 12 hour periods, 0000 GMT and 1200 GMT on the 14th and 0000 GMT on the 15th. In general the trough of interest deepens (primarily between 0000 GMT and 1200 GMT) and moves northeast through the Ohio Valley to be centered over eastern Ohio and western Pennsylvania at 0000 GMT the 15th. More will be said about this upper air flow in later sections.

In summary, this storm seemed a very good choice for study, exhibiting several characteristics indicative of orographic influence and consistent with synoptic experience (e.g., Austin, 1941 and US Weather Bureau, 1941) and with Colucci's climatology:

1. A primary cyclone which tracks northeast along the western slopes of the Appalachians.
2. A shallow cold dome east of the mountains accompanied by a shallow inverted ridge over the middle Atlantic states.
3. A cold anticyclone centered over New England and a warm anticyclone in the western Atlantic.
4. Frontogenesis in the Carolinas or Virginia followed by the formation of a secondary low center.
5. Heavy precipitation in the middle Atlantic states, particularly in upslope flow along the eastern slopes of the Appalachians.
6. Lingering precipitation in a cyclonic circulation west of the mountains as the primary storm stagnates and slowly fills.

While it is unreasonable to expect that any model with finite grid size could predict some of the smaller scale features described above, 1.25° resolution should be sufficient to resolve the primary storm, inverted ridge and trough, secondary formation, and some
semblance of the strong front. (Of course, features with a scale of $2\Delta x$ or less are not to be taken too seriously.) It will still be up to the forecast skill of the model, however, to produce meaningful features as well as any accompanying precipitation.
IV. RESULTS FROM THE MOUNTAIN CASE

Surface analyses for M are shown in figures 5b-10b, corresponding to the observed surface patterns in figures 5a-10a.

A 12 hour forecast to 0000 GMT the 14th brings us to the point from which we had previously hoped to begin. While these 12 hours have given the moisture field time to become more realistically suited to the synoptic situation, some very important discrepancies have evolved. The largest error by far involves the low center in the Gulf states. The position is fairly accurate, but the model storm has filled by 6 mb, now underforecasting the depth of the storm by 12 mb.

The trough to the northeast of this center was likewise forecast too shallow, the error decreasing in magnitude to the northeast to about 8 mb in eastern North Carolina and 4 mb on the eastern Maine coast. The high pressure system in the southeast has become quite fragmented and disorganized. This is believed due primarily to a reorganization of the surface high to fit a slightly different upper air flow which, in turn, is due to the inevitable inaccuracies accompanying the lack of data over the Atlantic Ocean. The decrease in the strength of the southerly flow as a result of this weakened high partially explains the inability of the model to maintain the warm temperatures over the southeast coast. Another major problem is the continued lack of development of a cold front to the rear of the cyclone. In actuality this cold front accounted for much of the heavy precipitation observed, but the model shows no sign of such a development, partly because of the underforecast of the intensity of the low center and its attendant circulation.
Another curious and disturbing development is the quite localized region of falling temperatures in eastern Kentucky and Tennessee. This trend is evident throughout the forecast up to this point and, as shall be seen, continues for yet another 18 hours. Several possible explanations were investigated in an attempt to explain the formation of this anomalous cold pool in these first 12 hours. Cold advection was present early in the forecast period but its magnitude was small and limited to areas north of Tennessee. In fact, as time progressed, warm advection became rather significant.

Upward motion due to orographic forcing could cool the lowest layer, but this was an area of orographic descent during much of the time this anomalous pool was intensifying.

Evaporational cooling by falling precipitation, while a plausible physical process, is not incorporated into the model.

Radiation effects could, under certain specific conditions, cause a local cooling. However, both the long and short wave radiation were assumed to be zero in these simulations.

This behavior has, in fact, been recognized in other experiments with different versions of the model. The error has been associated to numerical instability which arises under a certain combination of conditions of ground temperature, s-level temperature, snow cover, and particularly layer thickness. Recall that in a z-coordinate model, it is possible for the layer next to the ground to have any thickness less than or equal to the full thickness of an upper layer. It is this behavior which is believed to be responsible for the odd behavior of the temperature field in Kentucky.
Moving on now to the 18 hour prediction for 0600 GMT (figure 6b), the low center can be seen to have moved almost due east. Pressures north of about 39°N between the Mississippi River and the East Coast are virtually unchanged with the exception of the western slopes in West Virginia. Here there is a slight hint of an inverted trough as far north as northern Ohio. However, it does not begin to resemble the observed circulation pattern induced by the northward movement of the surface low.

The model's anticyclone over the Great Lakes is somewhat too strong and slightly too far north, but is an overall good forecast, particularly with respect to its extension into the southern plains states. While there is some hint of an inverted ridge east of the Appalachians, its amplitude is much too small. It appears to be more a part of the anticyclone proper, unlike the sharp features in the observed pattern. But this may only be a matter of magnitude, the model pattern being a grossly smoothed version of the observed distribution.

Also of note at this time is the southward shift of features in the Atlantic. The surface trough has moved south to about 37°N, while the observed trough remains more to the north. The forecast position of the warm anticyclone, now reorganized, is to the south and west of its observed position. The combination of these two features signals a decreased blocking effect, allowing easier progression of the cyclone to the east. Indeed, while the center itself fills slightly, pressures have fallen about 4 mb over coastal North Carolina, whereas observed pressures in this area have fallen only
about 2 mb.

The forecast temperature field must be considered poor, even aside from the anomalous cold pool. The two major zones of observed contrast are, at present, almost wholly absent in the simulation. There is a small increase in the temperature gradient over North Carolina. The significance of its magnitude may be questionable, but the model probably can do no better given the 1.25° resolution. Closer examination of a detailed observed analysis shows a significant filtering of cold air southward at the surface. This process is present in the model to a much lesser degree, also accounting for much of the error in frontal intensity. The cooling effect of melting and/or evaporation of falling hydrometeors, an effect not incorporated in the model, could also account for some of the cooling north of the observed front.

The cold front trailing the observed cyclone is totally absent in the model forecast. This has a very obvious effect on the precipitation in this region, completely missing the heavy convective rains in the Gulf states. This is shown in figure 11b. (Note the change in contour interval from the observed case. Though still quasi-exponential, the smaller intervals were necessary because of the severe under-forecast in the amount of rainfall.) The lack of that cold frontal band is clearly the major fault in the forecast precipitation. The stable precipitation was forecast with somewhat greater skill, although its axis was too far south, particularly over the coastal areas. Also, there is little evidence from this forecast alone of any orographic effect; there is no axis of enhanced rainfall aligned along the windward slopes and precipitation continues to fall on the lee side.
The 24 hour forecast for 1200 GMT (figure 7b) finally indicates that the model simulation should be considered a failure, at least for the purposes of this experiment. The "primary" (and only) storm center has not tracked west of the mountains, but east, confirming suspicions based on the 0600 GMT patterns mentioned above. The cyclonic circulation west of the mountains cannot be construed as the model's attempt to locate a separate center there. It is, rather, a lee trough effect which will be discussed in greater detail later. Some possible reasons for the general failure of the forecast will also be discussed.

The anticyclone to the north of the storm is forecast too strong over the Great Lakes and southeast of New England, but is otherwise fairly accurate, particularly if one allows for the fact that the cyclone has moved too far east. Again, that portion of the high which was located over the middle Atlantic states tends to form an inverted ridge there, and now seems to be developing some character of its own. It is accompanied by a wedge of cold air, evident in the surface temperature analysis. Cross sections to be discussed in detail later show that this perturbation in temperature is a maximum at the surface, and virtually non-existent above 900 mb. This cold wedge is, of course, hydrostatically necessary, and is in agreement with the fact that the phases of the temperature and pressure perturbations are roughly coincident in the simulation. They are not so clearly coincident in reality only because the observed cold wedge is not strong at the surface, but at around 910 mb. (This will also become clearer in the cross sections.) A simple example involving
a purely hydrostatic calculation shows that replacing a 400 m thick slab of air with an equal depth of air which is 5°C cooler results in a pressure increase of about 1 mb. While certainly too simple a scenario, the general idea is certainly applicable and consistent with the model forecast patterns.

Other aspects of the storm at this time include the growing cold pool in Kentucky and the lack of development of either the stationary front in North Carolina or the cold front now in Georgia. The precipitation pattern (figure 12b) shows two maxima. The smaller maximum is presumably due to the eastward movement of the storm out over the warm water. The second maximum, if not accurate, is at least more realistic. There is an apparent enhancement in the upslope flow and a decrease of the lee side compared with the 0600 GMT pattern. The observed precipitation west of the mountains, however, has increased, although this may be largely attributable to the difference in storm tracks. Lastly, the anticyclone to the southeast has almost completely collapsed, leaving little or no blocking effect ahead of the advancing model cyclone.

At 1500 GMT (figure 8b), the forecast position of the low center has shifted dramatically, however insignificantly, west, but still lies east of the mountains. Pressures have risen in general, with the largest increases occurring southwest of the low and resulting in a northeastward shift of the geometric center of the storm. The cold anticyclone to the north is still reasonably well forecast except for its intensity and its extension east-southeastward which pushes the Atlantic trough too far south. The ridge-trough couplet east and
west of the mountains is still quite prominent. This is particularly true of the trough in Ohio, since pressures there have remained steady for the most part while surrounding areas are experiencing pressure increases.

The temperature field may, at first glance, appear more accurate, largely because of an apparent "cold front" to the rear of the storm. Such an illusion is dispelled with the realization that this "front" is solely a product of the anomalous cold pool in Kentucky, now covering one-third of Kentucky with temperatures below 20°F. The temperature gradient through eastern Virginia might possibly be construed as the model's attempt to create a front, but it is neither very strong nor accurate, nor does it accompany the vorticity axis associated with the pressure trough to the east as in the observed case.

In the next three hours, the model low undergoes its greatest deepening. From 1500 GMT to 1800 GMT, the central pressure falls 7 mb to 1008 mb (see figure 9b). Unfortunately, the central pressure of the real storm was not observed to fall this drastically at any time during its entire evolution. Of some consolation is the placement of the low very close to the position of the deepening secondary. While the amplitude of the wave-like streamlines in the ridge-trough couplet over the mountains appears to have decreased, the corresponding anticyclonic-cyclonic couplet has not really decreased in intensity since the winds themselves are stronger. Vorticity patterns will be displayed and discussed in a later section.

The character of the high pressure system to the north is now beginning to deviate more seriously from the observed character.
Pressures are held too high over the Great Lakes and the second center is missed altogether. Aside from the generation of temperatures colder than 10°F in the cold pool, the temperature field remains essentially unchanged.

The forecast 1800 GMT precipitation pattern (figure 13b) appears somewhat better, although much of its success is admittedly fortuitous. The observed pattern has become less influenced by smaller scale features and more influenced by synoptic scale forcing, looking more and more like the model results rather than vice-versa. However, the general region experiencing some kind of precipitation was fairly well forecast, particularly the latitudinal extent. Smaller scale features as well as total amounts were not forecast nearly as well. It may be worth noting that the general upward trend of apparent skill in forecasting precipitation may be in part due to the model recovering from its poor moisture initialization (climatology). If this is indeed the case, it points up the need for further experimentation concerning the sensitivity of the model to the initial moisture analysis.

The 36 hour forecast for 0000 GMT on the 15th (figure 10b) bears as close a resemblance to reality as we have seen since the earliest part of the simulation. While the centers and intensities of high pressure are not perfectly forecast, the general regions are picked up fairly well. A trough of sorts is forecast for western Pennsylvania and Ohio coupled with a small ridge in northeast Pennsylvania. Even the temperature field looks more realistic. But, again, this may be more a result of reality developing so as to resemble the simulation rather than the model skillfully simulating the actual
development, in a manner analogous to a stopped clock giving the correct time twice a day.

It is difficult to evaluate the final precipitation forecast because most of the precipitation is falling over the water. But the forecast of small amounts lingering over land was fairly good despite some error in placement.

The 500 mb forecast series is shown in figures 16b-18b. While the height field at 0000 GMT on the 14th lacks some intensity and negative tilt at the trough line, the forecast can, in general, be considered quite good. The weakness of the cold advection into the trough, however, may foretell future insufficient deepening. This does, in fact, happen. At 1200 GMT the position of the trough is quite good, but lacks the depth and amplitude of the observed trough. Another discrepancy which may have significant bearing on the track of the surface cyclone, is the insufficiently strong ridge along the east coast. This is probably an important reason why the model storm took a more easterly course than observed. Similar comparisons can be drawn from the 36 hour forecast for 0000 GMT on the 15th. Also notable at this final period is another trough moving in over the high plains, contrary to observation. The Rocky Mountains may be having too much downstream effect in this area (i.e., a lee trough such as described earlier).

The observations above are consistent with the findings of Bettge et al. (1976). Working with the 2.5° latitude-longitude, 6 layer global GCM, they found a 15% loss of kinetic energy in the atmosphere as a result of the initialization alone. A 48 hour forecast
resulted in a further decrease of 15%, particularly in the mobile short waves. While a direct, quantitative comparison with this simulation is not possible, a visual inspection strongly suggests that the same problem likely exists in the 1.25° LAM since the meridional height gradient is significantly weaker and the low level patterns are much less intense.

As mentioned before, the simulation must be considered a failure, at least for the original purposes of this study. The model storm, while it represented a synoptically realistic situation, did not adequately simulate the characteristics of the type of storm chosen for study and described in detail earlier. In particular the model failed to simulate:

(1) a primary cyclone which tracks west of the mountains and fills,

(2) a secondary storm which forms on the coastal plains of the Carolinas,

(3) a strong quasi-stationary front on the southern edge of a shallow cold dome of high pressure east of the mountains (though this may be resolution dependent),

(4) heavy precipitation in association with a cold front moving through the Gulf states, and

(5) heavy upslope precipitation in easterly winds forced over the mountains.

Other shortcomings include both an average error in central pressure of about 10 mb and insufficient rainfall throughout the 36 hour period.

Several possible reasons for the insufficient strength of the cyclone and its failure to follow a more northerly track are as follows:
(1) insufficient strength of the mid-tropospheric short wave trough resulting in
   (i) insufficient vorticity advection aloft in turn resulting in decreased vertical motion and low level convergence,
   (ii) decreased southerly component discouraging northward progression;

(2) poor initialization of the moisture distribution resulting in a delay and an underestimation of precipitation and consequent latent heat release and its contribution to cyclone development. This may indeed be a major factor. Danard (1964) documents a case in which "the low level convergence associated with the release of latent heat is about three times as large as that arising from dry adiabatic processes." The consistently higher forecast central pressures may, then, be related to the consistently lower values of precipitation produced by the model. (Experiments are currently being conducted at NCAR to determine the sensitivity of the model to various initial moisture distributions);

(3) excess smoothing and/or numerical diffusion within the model;

(4) the necessity of using smooth topography may result in a much less obvious enhancement of precipitation in upslope flow.

Conceding the model's inability to accurately simulate the storm that was so carefully chosen and analyzed, it may still be possible to examine the model's sensitivity to the orography by comparing this run to a simulation excluding mountains. Did the inclusion of orography improve a forecast that otherwise would have been even worse? Was it to a large degree, indicating that the mountains play a crucial
role, even in their highly smoothed form? Or was the improvement very minor, indicating that other, more important factors dominate and are primarily responsible for the poor forecast? If the former, this indicates that much more work should be done on the inclusion of realistic topography and its effects. Worst of all, could the topography have been detrimental to the forecast? If so, this presents a whole new class of more serious problems.

With these questions in mind, we proceed now to an analysis of the no-mountain (NM) case.
V. RESULTS FROM THE NO-MOUNTAIN CASE - COMPARISON WITH THE MOUNTAIN CASE

Surface plots from the NM run are shown in figures 19-24, corresponding to the time periods shown for M. Since the NM precipitation forecast was quite similar to the M forecast, the differences between the two runs, M-NM, are shown in figures 25a-28a for the 3 hour periods ending 0600 GMT, 1200 GMT, and 1800 GMT the 14th and 0000 GMT the 15th, respectively. This also facilitates visual comparisons with the distributions of orographically forced vertical motions from M. This quantity, designated OROWS, is calculated simply as

\[ \text{OROWS} = w_s \cdot VH \]

where \( w_s \) is the vector wind at the s-level and \( H \) is the height above sea level of the ground surface. Use of the first and/or second layer winds in the computation of orographic forcing was considered, but the s-level winds were chosen so as to be consistent with the model formulation which also used the s-level values. The plots of OROWS at 0300, 0900, 1500, and 2100 GMT are shown in figures 25b-28b, respectively, depicting the conditions at the beginning of the 3 hour periods of precipitation in figures 25a-28a. Although they were also calculated, areas of orographic forcing not associated with the cyclone under study are not shown in the figures.

The NM simulation was, in general, very similar to the M run. A detailed description of the features such as was given for M will be omitted in the case of NM in favor of stressing the differences between the two cases.

Starting with the 12 hour forecast for 0000 GMT on the 14th and
comparing figure 19 with figure 5, the similarity is plainly evident. Only three differences truly stand out: (1) a westward shift in the position of the cold high, (2) the lack of perturbation in the pressure field around the mountains, and (3) the cold pool in M is missing in NM, strongly suggesting that the anomalous behavior in M was indeed precipitated by the inclusion of mountains.

Similar comments apply also to the 0600 GMT patterns. (figure 20 vs. figure 6). The pressure perturbation in M, as compared to NM, is becoming even more evident. The isotherms in NM are virtually straight lines over the "mountains." This indicates that perhaps the perturbation in the M temperature field really did reflect model skill in predicting the cold dome east of the mountains, though this conclusion must be tempered with the knowledge that scale of this perturbation is dangerously close to the scale of computational noise (Note: from this point on, the term "temperature perturbation" shall refer to that part of the perturbation not caused by the anomalous cold pool. This cold pool did not appear at any time in the NM case and, although it represents a major difference, shall not be referred to again.) In addition to a more westward shift from the M case, the high to the north is also stronger on its southern side. For example, pressures average 1–2 mb higher at 40°N in NM, even in the inverted ridge on the east coast in M.

Before examining the precipitation difference maps, it is important to realize that although it is tempting to treat the two cases as the same storm with varying topography, the two simulations actually represent two different storms. It is true that they greatly resemble
each other, but to say that one particular point got rain in M, but not in NM, as a direct result of orographic forcing is unfair. It may be that the M storm merely tracked further northward, shifting the northern boundary of its precipitation shield. This is, in fact, one of the major reasons why it was decided to visually, rather than quantitatively, correlate the precipitation difference fields with the OROWS fields. Furthermore, it was decided that no correlations should be computed involving fields of difference between the two cases, primarily because of this difficulty in ascribing cause to the difference.

A second reason for not calculating a correlation between the precipitation difference and OROWS fields is the nature of the hypothesized relationship. An example will illustrate this point. Suppose that at both point A and point B OROWS is negative (downward). Suppose further that there is precipitation at point A in both M and NM, but M has 0.05 inches less than NM. This results in a positive correlation, lending support to the hypothesis that downward (upward) topographic forcing decreases (enhances) precipitation. Now, suppose that at a point B, there is no precipitation in either case and thus no difference. While we would recognize this situation as not applicable to the hypothesis, a calculation of correlation including that point results in a lower correlation. Limiting the area of correlation to the areas receiving some arbitrary minimum amount of precipitation, x, in both cases presumes a particular relationship (depending on x) and severely limits the sample size, resulting in a lower significance of the correlation obtained. (A detailed description of the correlation...
procedure and its problems follows in a later section.) It was con-
cluded then that visual comparison was the easier, more flexible, and
generally preferable method.

With these points in mind, refer now to figures 25a and 25b,
depicting the precipitation difference and OROWS fields, respectively.
Up to this point, the latitudinal difference in storm track has not
become very evident, at least in terms of precipitation. The area
of enhancement over western Virginia and North Carolina and the area
of lessened precipitation to the west show very good agreement with the
distribution of OROWS. The correlation appears even better after
referring back to the plot of 0300-0600 GMT M precipitation (figure 11b).
We can now mentally discard the large area of strongly positive OROWS
in Georgia and Alabama because little, if any, precipitation fell there.
The couplet of positive and negative anomalies east and west of the
mountains now match the OROWS pattern well, being aligned with the
main axis of precipitation. The area of strong descent in New England
can also be ignored, because there was no precipitation there in either
case. Likewise, the area of negative precipitation difference off
the Virginia Coast should be ignored, being more likely due to a
somewhat slower advance of the M storm and not a direct orographic
effect. (The deficit is, obviously, an indirect effect of the topo-
graphy, since the two simulations would be identical were it not for
the mountains.)

However, prior to concluding that the mountain forcing could have
played a role in the redistribution (aside from the caution used when
ascribing cause to correlation), it is logical to question how large
a percentage of the total vertical motion OROWS comprises. The graph in figure 29 may help answer this. The lower curve charts the value of the maximum upward OROWS versus time. The upper curves, solid for M and dashed for NM, represent the value of vertical velocity at the 3 km level at the point of maximum OROWS. In general, the values of OROWS, compared to the total vertical motion, is not insignificant, averaging perhaps 25% of the total. One should remember, of course, that OROWS is a maximum at the ground surface and its effect above the surface damps with height, particularly under stable conditions. However, such an effect could be reflected in the precipitation field, and it does seem to be at this time.

The NM forecast for 1200 GMT is shown in figure 21 (corresponding to figure 7). The greater strength and westward shift of the high is the dominant difference. Although the central pressure is only 4 mb higher in NM, the differences throughout the majority of the area also average higher. The largest differences are found in the area of the inverted trough in M, where pressures are about 6 mb lower than NM. The cyclones themselves are comparable in both the location and in the depth of their (geometric) centers. Wind speeds are slightly higher northwest of the NM low, owing to a somewhat steeper pressure gradient produced by the stronger high. The temperature perturbation in M can now be seen to have shifted the orientation of the isotherms over the middle Atlantic States from predominantly east-west in NM to northeast-southwest in M. The magnitude of the temperature gradient is slightly higher in M but is located east of the cold "dome" and north of the trough, unlike the structure of the observed gradient (see figure 7a).
The precipitation difference map in figure 26a is now showing signs of the more northerly progression of the M precipitation, suggested by the fact that the positive differences are generally north of the negative differences. For this reason, correlation with OROWS becomes more difficult and less meaningful. As before, we can ignore the area of positive OROWS over Georgia and Alabama, as well as the areas of negative OROWS over New England and southern Canada. Even allowing for this and the fact that the areas of positive OROWS move northward, the correlation is not impressive. Many of the positive and negative anomalies occur in areas of upward and downward forcing, respectively, but the peaks in these anomalies fall closer to areas of no forcing than the OROWS extrema. Isolating a particular reason for this curious behavior is difficult. It may be that the cold perturbation east of the mountains represents a shallow dome of cold air. This could represent an added "obstacle" over which warm, moist air is forced. Although this seemed like a plausible explanation, it did not seem sufficient, so a more extensive examination was initiated.

Figures 26c and 26d are plots of the difference between the three kilometer vertical motions, M minus NM, for 0900 and 1200 GMT, respectively; i.e., the beginning and the end of the period of precipitation depicted in figure 26a. Units are cm s^{-1}. Note, again, the northward shift of the M case relative to the NM case. The correlation of these fields with the precipitation difference field (figure 26a) is remarkably good. The only area which does not appear to correlate as well lies south of the nearly coincident maxima, through central Virginia. The precipitation differences remain highly positive here, but the
three kilometer vertical velocity differences decrease considerably, becoming negative in southern Virginia. This region is precisely where the low level orographic forcing is taking place (see figure 26b). It seems highly likely, then, that these added precipitation differences are a result of the orographic lifting. But although the magnitude of OROWS approaches the magnitude of the three kilometer vertical velocity differences, its effect on the precipitation is apparently much smaller. This suggests that its influence is, indeed, restricted to a much shallower layer, while the three kilometer vertical velocity differences are representative of a much deeper layer. This conclusion might be expected. On the other hand, an axis of local maximum in the three kilometer vertical velocity differences from central Virginia to eastern South Carolina also suggests that the influence of the orographic forcing is felt, at least to some degree, as high as three kilometers. But it must be concluded that the principal effect of the mountains on the precipitation in these simulations was not a direct one involving very localized enhancement through orographic lifting. Rather, the mountains acted in a very indirect way by altering the track and circulation patterns of the storm and this eventually had a much more profound effect on the precipitation pattern.

At 1500 GMT (figure 22, corresponding to figure 8) the pressure systems in M and NM have virtually the same central pressures. But there is a decided shift in high pressure eastward and low pressure westward in M. Pressures are 4-6 mb lower in M over the mid-Mississippi Valley, and up to 10 mb lower over the northern plains. While slightly cyclonic, the NM flow over Ohio shows little sign of the distinct
inverted trough present in M. And, despite the eastward shift in the position of the M high, pressures are still lower in M virtually everywhere in the east also. This results in the overall impression that NM is dominated to a significantly greater degree by high pressure, incorrectly it would seem, since the M anticyclone bears a much stronger resemblance to the observed high.

Notice should again be taken that the M temperature perturbation has resulted in a slightly amplified "frontal" zone not present in NM. However, it is strongest east of the cold perturbation and is very much weaker near the low itself. In contrast, the observed front is strongest at the low center, occurs south of the cold perturbation, and relaxes substantially to the northeast. Furthermore, the observed low center appeared to form on the observed front, whereas the model front seems to have developed relatively independently from the model cyclone.

Most of the comments made about the 1500 GMT forecasts apply also to the 1800 GMT (figure 23, corresponding to figure 9). The pressure differences over the Mississippi Valley and plains states mentioned earlier remain virtually unchanged, although pressures over the former have dropped about 4 mb in M and NM. Unfortunately, these falls are occurring in an area where the observed pressures are rising. In fact, a new high center is observed to be forming nearby. Overall, the M forecast for this ridge is probably better, primarily because of its character there and despite the pressures being too low. The low centers themselves deepened by 7 mb in both cases, perhaps pointing out the importance of forcing on scales much larger than the area in study and the relative unimportance (at least for short period forecasts near low orography) of the inclusion of moun-
tains. The inverted trough west of the mountains in M is the primary
difference in character between the two cyclones and the principal
reason why the M storm appears to be a better forecast. Some of this
apparent skill may be fortuitous, since the M simulation, as well
as the NM simulation, failed to forecast the movement of the primary
storm west of the mountains. This movement was clearly the principal
cause of the cyclonic character there in the observed case. That
the lee trough effect (mentioned briefly earlier and discussed in
greater detail later) could represent a small contribution to the
observed trough west of the mountains may be one conclusion drawn
from the M-NM difference.

As we continue to follow the "front" resulting from the M tempera-
ture perturbation, we still notice no corresponding feature in NM.
Optimism concerning the skill of the M simulation is, however, unwar-
ranted. Strikingly unlike the observed cyclone, which expands and
depens along the observed front, the model storm proceeds eastward,
seemingly oblivious to the model "front". Up to this point, conclu-
sions concerning the ability of the model to recreate this aspect
of the storm have been withheld. But it now seems safe to conclude
that the model did not adequately simulate the observed front or its
role in the development of the storm, partly because of scale limita-
tions.

The precipitation difference for the 1500-1800 GMT period
(figure 27a) has a maximum of about 0.1 inches in northern Virginia,
amounting to approximately one-half the M precipitation. The M storm
appears to be producing more overall precipitation now, contrary to
the previous two periods. The important aspects of the OROWS pattern have not changed significantly in the past six hours. The larger lobe of positive values remains in the same location and is about 0.4 cm s\(^{-1}\) less intense. The principal area of downward motion is also somewhat less intense. Located in West Virginia, it now extends more strongly into western New York and southern Ontario. The smaller lobes of upward and downward motion have shifted to Tennessee and South Carolina, respectively. Any other comments that need be made concerning these two plots have already been made regarding the 1200 GMT distributions, including the general northward shift of precipitation in M.

The NM surface analysis for 0000 GMT on the 15th (figure 24, corresponding to figure 10) shows considerably greater change from 1800 GMT on the 14th than does M. Specifically, the NM anticyclone has built to the west-southwest, now centering over Lake Michigan and weakening significantly over northern New England. In this regard, M is clearly a superior forecast. It holds the high over northern New England, builds higher pressure over the southern Mississippi Valley, and reduces pressures over the Great Lakes, very much as observed. Though perhaps right for the wrong reason, the M forecast for the cyclone appears very good also, with a trough lingering over western Pennsylvania and eastern Ohio. The NM storm, though centered at the same point, lacks all traces of this feature, with the streamlines unperturbed from New England, through the Ohio Valley, and south to the Gulf of Mexico.

The major precipitation differences at this time (figure 28a)
occur as a result of somewhat greater lingering precipitation over Maryland and southern Pennsylvania. The area of difference there of over 0.03 inches represents up to about half of the M precipitation. The peak upward values of OROWS correlate better now, with less latitudinal shift which was noted earlier. The lobe of positive precipitation difference seems somewhat set off from the principal precipitation areas, also strongly suggesting an orographic influence. It also makes the M forecast appear slightly more skilled in light of the observed precipitation distribution (figure 14a). The effect of the three kilometer vertical velocity difference (not shown) is more difficult to isolate now. The positive precipitation anomalies lie somewhere between these vertical velocity differences and OROWS, indicating that both are important.

The 500 mb series for NM are shown in figures 16c-18c. The only really significant differences from M occur toward the end of the simulation. The apparent dissimilarity in the trough due to the northward shift in NM of the 552 dm contour is of no great consequence because of the very flat character of the 500 mb surface there. It does imply a slightly weaker southerly component of the flow east of the trough, compatible with the slightly northward shift of M precipitation and, as will be seen, the low level vorticity associated with the surface feature. The most notable difference in the 36 hour forecast is the greater acceleration in M of the ridge-trough system over the western half of the study area. This is likely a consequence of a lee trough produced in the westerly flow over the Rocky Mountains. Such a feature would tend to shorter the wavelength of the ridge in
the plains states, causing it to move faster and propagating this
effect downstream, albeit weaker.

Thus, we have seen that neither simulation produced a forecast
suitable for examining the details of the storm chosen for study.
The simulation incorporating orography did, however, produce somewhat
more accurate features:

(1) a trough on the western slopes of the Appalachian Mountains,
(2) some enhancement of rain in upslope flow, and
(3) a more accurate distribution of high pressure systems adja-
cent to the cyclone.

In lieu of investigating the real storm via model simulations, the
remainder of this study will concentrate more heavily on the differences
between the two cases. Recognizing the impossibility of investigating
every aspect of the simulations, the remaining discussions will focus
on:

(1) the three dimensional structure of the two cases,
(2) an analysis of the distribution and production of low level
vorticity in each case, and
(3) correlations amongst several variables within each simulation.
VI. CROSS SECTIONS

Cross sections provide an excellent method of viewing the two dimensional structure of a storm through any particular plane. By using several planes, a quasi-three dimensional picture evolves. Among the characteristics that become evident are the temperature, moisture, and stability distributions as well as the placement, intensity, and extent of frontal zones. In this case, the cross sections are meant to display the observed and simulated storm characteristics mentioned in previous sections, such as the cold dome east of the mountains. It will be instructive to note the degree to which the orography and its direct and indirect effects are reflected in the temperature and/or moisture distributions.

The placement of the three cross sections chosen for this study are shown in figure 30 and will be designated A-A', B-B', and C-C'. The "observed" cross sections are taken from soundings made at the indicated stations, transmitted via teletype, and archived at MIT. The radiosonde stations used in constructing the sections are those closest to the lines, with the station data assigned to a point on the line closest to the station. JFK was used for both the A-A' and B-B' sections. All of the lines lie along a diagonal of grid points and, hence, grid point data can be used for the model cross sections. Although details of the calculations are omitted here, temperature was calculated hydrostatically as a layer mean, relative humidity was calculated following a procedure given by Oliger et al (1970), and the values were assigned to a point in the vertical corresponding to the pressure at the midpoint of the layer.
Figures 31-33 are the cross sections at 0000 GMT on the 14th for A-A', B-B', and C-C', respectively. Figures 34-36 correspond to each of the sections at 1200 GMT, and figures 37-39 are for 0000 GMT on the 15th.

The distributions for 0000 GMT on the 14th show no truly remarkable features. There is a front in the northern regions of A-A' at about 800 mb, not reflected at the surface, with an inversion above it extending to around 700 mb. M reproduced this feature much better than NM, albeit 50 mb too high. The M and NM horizontal temperature contrast is comparable here, but the ground to 650 mb layer is much too close to isothermal in NM. A front is observed at the surface near Cape Hatteras (not evident in the surface analysis merely due to the choice of isotherms and contour intervals in that display). Neither M nor NM recreated this front (again, largely because of the scale question), but both produced one (erroneously) further north near Chatham. The forecast distribution of moisture throughout A-A' does not even resemble the observed pattern for either M or NM. The two are quite like each other however, with the only regions near saturation lying above 600 mb. This is undoubtedly due to the poor moisture initialization, requiring the air to be lifted higher and longer) to reach saturation.

Proceeding westward, the observed layer of cold air is found progressively lower and penetrates further south. In B-B' (figure 32) the axis of cold air is found at about 830 mb in the north to 930 mb over Washington, D.C. From Greensboro, N.C. south, the temperature decreases monotonically with height. Both M and NM capture the lowering
of the cold intrusion, M again faring better at the magnitude and creating greater stability further south. M even produces a small, low level inversion between Greensboro, N.C. and Athens, GA, perhaps hinting at some skill in simulating the cold dome at the surface there. This feature is relatively important to this study and shall be followed closely, but with scale considerations always in mind. The moisture forecasts are slightly better here for the large scale. It is interesting to note (1) the more northward progression of moisture at the surface in the middle latitude region of NM, (2) the greater humidity on the southern (upslope) side of the orography in M, and (3) the strong gradient of moisture at the "crest" of the orography. The orography and the edge of the cold air appear to be blocking the northward advance of the moist air and simultaneously increasing the humidity on the upwind slopes (see figure 25b).

Further west, at C-C' (see figure 33), the observed cold layer is again lower, but more continuous and well defined. An inversion is found at the surface just north of the warm low center. It begins somewhat higher aloft (60-100 mb) further to the north. The M forecast is clearly superior in this case. Though a bit disorganized in some other areas, the surface inversion is present. In NM, only the intrusion of cold air aloft is evident, with no hint of the cold layer near the surface toward the south. Before crediting the M forecast with greater skill in this regard, it should be realized that much of this apparent improvement is a result of the anomalous cold pool discussed earlier (see figure 5b). The model relative humidity patterns are better in this cross section, though both severely
underforecast the low level moisture. M fares somewhat better, with 90% humidity extending to the surface and generally greater low level humidity throughout its southern half. Orographic uplift and colder temperatures probably combine to account for this.

At 1200 GMT (figures 34-36) the observed cross section through A-A' has undergone some rather large changes (see figure 34). The cold air at about 780 mb in northern regions at 0000 GMT has descended to below 900 mb and spread southward through the entire northern half of the cross section. The previously mentioned surface front near Cape Hatteras has become somewhat more intense and a second zone of surface temperature contrast has developed near Chatham, Mass. The temperature and moisture distributions near the latter reflect the fact that there is (or has been) colder, drier air moving from land areas southeastward out over the water. Chatham is a coastal station very much affected by surrounding waters and the temperature contrast and higher humidity at low levels seem to reflect this modifying effect. The remainder of the observed A-A' moisture distribution shows a moderate increase in saturated area, particularly in the vicinity of the front, but also at higher levels. There is also a northward shift in the saturated "column", which has a northward tilt with height. Aside from the highly variable humidity patterns, the regions south of the front remain largely unchanged.

In M, the most striking development in A-A' is the greatly expanded area of 90% saturated air. Though it, too, has moved northward since 0000 GMT, it is curious that the axis of moist air not associated with any front, unlike the strong feature observed. The axis of cold
The observed patch of moist air was neither strong nor extensive
and is believed to be merely a consequence of having Chatham (CHH)
as an observation point, it possessing a more nearly "oceanic"
character near the surface. In the case of the model, the local
maximum is believed due to a small maximum in s-level wind speed
which passed over that point, since the model moisture flux calcula-
tion is proportional to $v_s$. But, in any event, this particular feature
is not of major consequence to the forecast and is of a scale which, again, must be suspect.

The observed cross section through B-B' (figure 35) is very similar to that through A-A' except for a weaker intensity of the features and more widespread "saturated" air. The change in B-B' from 0000 to 1200 GMT is not as great as that in A-A'. The cold intrusion, already well south at 0000 GMT, has moved farther south to a point somewhere between Greensboro, N.C. and Athens, Ga. The southern boundary lies near the front which appears so strong in the surface analysis. The front does not appear as strong in the cross section because (1) the line joining Greensboro and Athens does not cross the front in such a way as to really emphasize the intensity of the temperature contrast and (2) the resolution afforded by the radiosonde station network is not as great as that of the hourly, surface only reporting stations.

In the southern two-thirds of the cross section, M seems to be the superior forecast. Warm air is overriding cold air to a greater extent (though still insufficient) and saturation is found all the way to the surface. It is tempting to tout this as evidence of orographic enhancement, except for the fact that even though NM did not produce saturation all the way to its surface (sea level + 10 m), it did produce it as low as the M surface.

Again, note the cold air right at the surface near the middle of the cross section. Though the actual low level lapse rate is of the opposite sign, the "advance" of this cold air south and the larger scale inversion it creates there are good reasons to conclude that M
is the better simulation. In the northern one-third of the cross section, though not really the principal area of interest in this experiment, NM appears to have done a better job. M produces a spurious double inversion pattern that simply is not observed.

The most obvious feature in C-C' at 1200 GMT (figure 36) is the warmth near the low center and its northward progression since 0000 GMT. To the north is the cold intrusion/inversion noted previously. To the south is a new inversion between 900 and 800 mb associated with cold advection southwest of the low center. Approximately half of the observed cross section possesses relative humidity greater than 90%, this area centered roughly on the cyclone position.

The M simulation is quite poor here near the surface, principally because of the anomalous cold pool discussed in earlier sections. Very cold air at the surface is located precisely at the point where the observed surface temperatures are the warmest. This cold air results in the placement of a "front" south of the cold pool and a widespread inversion up to about 880 mb, both of which are totally wrong. The extent of saturated air is also highly underforecast, reaching the surface largely because of the development of spuriously low temperatures there without the moisture field being significantly altered.

While not a terribly skillful forecast either, NM does manage to create the basics of (1) a warm "sector", consisting of a local maximum of temperature, albeit below 900 mb, and (2) high stability well to the north of the warm sector, although the layer of warm air is centered about 100 mb too high at the northern boundary. The
humidity pattern, like M, is a rather sizable underforecast of the area of near saturation (over 90%). Again, the high surface humidity in M, while it may, in fact, be partially due to orographic lifting, could have resulted purely from the colder temperatures there given the NM mixing ratio distribution.

At 0000 GMT on the 15th, the observed front in A-A' (figure 37a) has relaxed, particularly at the surface, and moved north slightly, along with a somewhat expanded area of greater than 90% relative humidity. The layer of high stability to the north is both stronger and more extensive. The remnants of the warm sector over the southern coastal states is also evident. Little else has changed. At this time, M shows an intensifying front off the coast of New Jersey and a stable layer that is of somewhat greater extent at lower levels, though slightly less intense aloft. The axis of high humidity in M has moved north, particularly near the surface. This is due to a trend toward trajectories which have longer fetches over water where humidities are higher (not shown). Such a development is also present in the observed pattern, although evaporation of falling precipitation is also a possibility in the real case.

NM's comparison with M is virtually the same as at 1200 GMT. The layer of cold air in the north is centered nearly 100 mb lower than that in M or that observed at the northern border. A small, low level front and its associated inversion extends further south and looks more like the observed distribution than M, but M is superior above the lowest 40 mb and to the north. The only major differences in the humidity patterns are that NM does not produce
saturation near the surface and that NM has a curious extension of high humidity to the south at about 970 mb. No obvious reason for the latter could be found.

B-β' has become much more complex since 1200 GMT (see figure 38). The warm, moist air overriding the shallow, cold air near the surface has started to become somewhat cut off in the observed case. The warm perturbation in the temperature field in the Chesapeake Bay region (see figure 10a) is also evident. A second inversion, noted previously in C-C' at 1200 GMT, is now present in the southern portion of B-B'. The area of moist air is observed to have become limited to the region below 700 mb for the most part. M's cross section B-B' (figure 38b) shows only minor changes in the temperature field in most regions. General cooling is taking place, especially at low levels in the southern three-fourths of the area. This results in a generally weaker surface temperature gradient, except for a "front" in Virginia near the crest of the orography. This resulted from stronger cooling (up to 5°C) north of the front and below 900 mb, the cause of which is discussed below. The general cooling south of the front is in response to cold advection taking place in the southeastern states. The moist region of M has moved north since 1200 GMT, particularly at the surface, but does not show any sign of lowering as in the observed case. The NM moisture distribution is similar in shape but located further south, again, particularly at the surface (e.g., the humidity is much higher below 900 mb throughout the southern half of the NM cross section). The NM temperature field differs from that in M in two respects. First, the highly stable layer is centered
about 80 mb lower, near 940 mb, largely as a result of the axis of cold air lowering more in NM than in M. Secondly, the lowest 50 mb are significantly colder near Washington, DC in M. The latter is related to the temperature perturbation in M east of the mountains which was described earlier. In light of (1) this large M-NM difference, (in excess of 5°C), (2) the greater degree of change here in M from 1200 GMT to 0000 GMT, and (3) the importance attached to the "shallow cold dome" in relation to the type of storm being studied, the cause of that temperature drop was investigated in greater detail.

The local rate of change of temperature at a point is given by

\[ \frac{\partial T}{\partial t} = - \mathbf{w} \cdot \nabla T - \mathbf{w} \left( \frac{\partial T}{\partial z} - \left. \frac{dT}{dz} \right|_{\text{ad}} \right) \]

where \( \mathbf{w} \) is the horizontal vector wind, \( \frac{\partial T}{\partial z} \) is the environmental lapse rate, and \( \left. \frac{dT}{dz} \right|_{\text{ad}} \) is the adiabatic lapse rate. In this instance, the moist adiabatic lapse rate is used because of the high humidity in the area. As we shall see, this results in, if anything, a more conservative estimate of the magnitude of the mechanism to be proposed.

For a sample calculation, a point is selected from B-B' lying very close to Washington (DCA; see figure 30) at 1800 GMT. The winds and temperatures to be used are those in the first layer, \( m=1 \). \( \mathbf{w} \) is assumed to be an orographically forced upward motion similar to OROWS, but calculated using winds at the first level, i.e.,

\[ \text{OROW} = \mathbf{w}_1 \cdot \nabla H. \]

This OROW is assigned to the \( n=1 \) level. \( \frac{\partial T}{\partial z} \) is calculated using the
s-level and second layer (m=2) temperatures and is assigned to the m=1 level.

The terms in equation 1, then, become

\[- w \cdot \nabla T = +19.4 \times 10^{-5} \text{C}^\circ \text{s}^{-1} = +8.4 \text{C}^\circ /12 \text{hrs}\]

\[w = \text{OROW} = 2.12 \text{ cm s}^{-1}\]

\[\frac{\partial T}{\partial z} = 7.9 \times 10^{-5} \text{C}^\circ \text{ cm}^{-1}\]

The value of the moist adiabatic lapse rate is taken to be -6.5 C^\circ cm^{-1}, valid for a temperature of 0°C, as given by Hess (1959). Applying the values to equation 1, we obtain

\[\frac{\partial T}{\partial t} = -11.1 \times 10^{-5} \text{C}^\circ \text{s}^{-1} = -4.8 \text{C}^\circ /12 \text{hrs}.\]

This value is quite close to that observed. Though its accuracy is undoubtedly partly due to chance, this represents a fairly good agreement with the M changes.

This result suggests, then, a possible mechanism for the development in the model of the M perturbation of temperature east of the mountains. Orographically induced upward motion, a maximum at the surface, has its greatest cooling effect at the surface (particularly if the motion is dry adiabatic). This, combined with southerly flow and warm advection aloft acts to stabilize the lower layer. There is a positive feedback of sorts through \(\frac{\partial T}{\partial z}\) in equation 1, resulting in ever cooler m=1 layer temperatures. Though at some point this argument must break down (possibly through the advection term becoming more dominant as the horizontal temperature gradient increases), the
process is plausible and seems to fit the model results. By this reasoning, one might expect the cold perturbation to extend up to the topographical ridge line. This is pretty much the case. In those instances where the ridge is warmer than points just to the east (despite cold advection instead of warm advection), two explanations come to mind:

1. the ridge itself is above the cold layer and into the warm air and warm advection aloft, and
2. the effect of finite differencing is manifesting itself by yielding only a small value of the height gradient when there may, in fact, still be a large value.

In actuality the process described may be of much lesser importance. The topography is actually quite level within 200-300 km of the coast. What we may have then is another case of the model results looking "right-mostly-for-the-wrong-reason". The observed advance of cold air is more likely due to cold advection, evidenced in analyses more detailed than those in figures 4-10. Winds in these analyses show some greater ageostrophic component east of the mountains, particularly in the eastern foothills of the southern Appalachians. What may be happening in the real case is that there is a buildup of mass east of the mountains (witness the pressure gradient at 0600 GMT in figure 6a), enhanced by the low level stability, leading to a subsequent shift in wind direction southward parallel to the mountains and resulting in cold advection. Some of this cold air which is advected may, in fact, be produced by the method described above in reference to M's cold perturbation.
To complete the picture in cross sections, then, refer now to C-C' at 0000 GMT (figure 39). As might be expected because of the anomalous cold pool, the agreement between M and the observed distributions is poor at low levels. M creates a highly stable layer below 900 mb, while there is actually a layer of instability there. NM does little better in this respect, and is worse than M up to about 700 mb since it extends the warm inversion much too far north. This is due less to excessive warm advection above 850 mb than to excessive cold advection below 850 mb. Here, several factors combine to indicate that M is a better forecast and that the mountains do play a role in shaping at least part of the character of the storm. For whatever reason, M produced a trough west of the mountains (discussed in detail later). The circulation around this trough prevented cold advection into the Ohio region such as that present in the NM forecast. This resulted in the M cross section of temperature being a much better rendition of the observed pattern in the north. (Were it not for the anomalous cold pool in M, the middle latitude, low level pattern would probably also be superior.) This lesser cold advection, in turn, resulted in lower stability in M, more consistent with that observed. The showery precipitation observed in the upper Ohio Valley is partly a reflection of this instability. There is still a great deal of nearly saturated air through C-C' in the observed cross section, with layer of moist air becoming deeper to the north. Both of the simulations have dried out, with the only significantly saturated air lying above 600 mb. The obvious reason for this discrepancy is that the model storms failed to track far enough north along the western
slope of the mountains. Relative humidities in the northern third of the region and at levels below 600 mb, however, are 20-30% higher in M. The differences above about 900 mb can be ascribed partly to the difference in storm track and associated regions of upward motion which were described earlier. (Note the differences in three kilometer vertical motion in figures 26c and 26d.) The higher humidities at lower levels, however, are curious, since they occur in a region of orographically forced downward motion. Downward vertical motion would preclude model precipitation and moist convective adjustments, both of which would lead to a greater retention of moisture. But this explanation would only apply if the relative humidity were close to 95% anyway, which it is not. We can conclude, then, that this extra moisture is a reflection of the altered flow patterns advecting moister air from the east and southeast into and around the trough west of the mountains (e.g., compare figures 9a and 23). This added moisture evidently increases the relative humidity to a greater extent than the warming during descent decreases it.

As a final note, the warm layer aloft in both cases is more than 100 mb too low. Also, the humidities are greatly underforecast, though M is somewhat better in that it retains higher values in the north at low levels.
VII. LOW LEVEL VORTICITY

Up to this point, it has been noted that the two simulations differed very little in the important aspects of the storm. They were both very much more like each other than the real storm. There were three differences, however, that could be considered significant. Two have already been discussed in previous sections. The first involved the moderate enhancement of precipitation on the windward side of the topography. Differences of up to 50% were evident, although it would be difficult to say with certainty how much was actually due to orographic enhancement. Comparison of the differences with orographically induced vertical motions suggest some degree of correspondence, especially when combined with a realization that the two storms were distinct and some differences are to be expected on that basis alone. The second difference involved the temperature perturbation east of the mountains. The surface temperature analyses, in conjunction with the vertical profiles displayed in the cross sections and a simple prognostic equation for temperature, suggested a fairly simple explanation for the evolution of such a feature through orographic forcing.

The third difference is perhaps the greatest. The M simulation produced a secondary, lingering trough west of the mountains that was completely absent in the NM case. In an effort to quantify the degree of difference and perhaps suggest a reason for the existence of such a feature, the vorticity near the surface was investigated. The hypothesis here is that some kind of lee trough is forming in the low level easterly flow.
The details of such behavior are discussed in various texts, such as Holton (1972). Basically, the assumption of homogeneous, incompressible, inviscid flow reduces the theorem of potential vorticity conservation to a very simple form:

\[ \frac{\zeta + f}{\Delta z} = \text{constant}. \]

This equation is most commonly applied to deep westerly flow (on the order of 10 km) over large mountain barriers. Compression of a vortex tube on the upwind side of the mountains results in a decrease of vorticity, and the subsequent stretching of the vortex tube on the lee side results in an increase of vorticity. The variations of \( f \) during these changes require that a lee wave on the horizontal plane be set up downstream from the mountains (see Holton for details).

The case of easterly flow is somewhat different, though the conservation law must still hold. In easterly flow, the resultant trajectory pattern is strongly dependent on the incident angle of the flow because of the variations of \( f \). Our particular case is further differentiated by its much smaller horizontal and vertical extent than is usually the case in applying this theory. The horizontal scale is on the order of 1700 km and the easterly flow is present only below about 2.5 km. The consequences of including the planetary boundary layer are uncertain. Nevertheless, as we shall see, the patterns of vorticity are strongly indicative of such "lee trough" behavior.

To perhaps aid in the determination of what factors are contributing to any vorticity change, the various terms in the vorticity
equation were calculated. The equation for the local time rate of change of vorticity in cartesian coordinates is

\[
\frac{\partial \zeta}{\partial t} = - \mathbf{w} \cdot \nabla \zeta - w \frac{\partial \zeta}{\partial z} - \nabla (\zeta + f) \cdot \mathbf{w} - \left( \frac{\partial w}{\partial x} \frac{\partial v}{\partial z} - \frac{\partial w}{\partial y} \frac{\partial u}{\partial z} \right) + \frac{1}{\rho^2} \left( \frac{\partial p}{\partial x} \frac{\partial p}{\partial y} - \frac{\partial p}{\partial y} \frac{\partial p}{\partial x} \right)
\]

where \( \mathbf{w} \) is the horizontal wind velocity and \( \beta = \frac{df}{dy} \). The last three terms on the right hand side of the equation represent the divergence term, the tilting or twisting term, and the solenoid term, respectively.

Now the problem arises as to where and how to calculate the vorticity and these various terms. In order to capture the character of the trough to the west of the mountains (as well as any pseudo-ridge east of the mountains), it is necessary to examine the vorticity at as low a level as possible since these features disappear quickly with height. In order to evaluate the terms containing vertical derivatives, this meant calculating the terms at 750 meters, intermediate between the lowest two layers. This did not present any particular problems in the case of NM, where, when necessary, values could be computed as simple means of the two lowest layers. Horizontal advections are computed as an advection of a mean vorticity by the mean wind and the values of divergence and density used were the average values of the two layers. The M case, however, presented some more difficult problems. When the lower layer was completely blocked (\( H \geq 750 \text{ m} \)), values from the second layer only were used.
The vertical advection term was not calculated for lack of a vertical velocity. The solenoid and tilting terms were omitted at those points and at the adjacent grid points (due to horizontal finite differencing) for lack of both a vertical velocity and a pressure at the blocked point. A pressure could have been estimated hydrostatically, but the solenoid term is consistently at least one order of magnitude smaller than, say, the divergence term. No such estimate for w could be made, however. When the lowest layer was not completely blocked, a depth-weighted average of wind, vorticity, etc. was used. This description is a gross simplification of the procedure, since the actual computations become quite involved. But the effective level, then, at which the various terms are calculated is that level which lies midway between the ground surface and the n=2 level (1500 meters).

Having described the basics of the hypothesis, the terms, and the methods of calculation, we now turn to a look at the fields themselves. To avoid the spate of figures involved in displaying all the terms for each time period, the following selections were made. The vorticity field (analyzed in units of $10^{-5} \text{s}^{-1}$) at six hour intervals are shown, beginning with 0000 GMT on the 14th of March in figure 40. As expected, the $\beta$ term and the solenoid term are, almost without exception, an order of magnitude smaller than the divergence term, which is the largest term in almost every case, and, hence, are not shown. The tilting term frequently approaches the divergence term in magnitude and cannot be neglected. For the most part, the u, v, and w advections tend to cancel but do, on occasion,
become important. The fields of the divergence term, the tilting term and the total of all the terms for 1200 GMT and 1800 GMT were chosen to be displayed, since this is the period of greatest activity. Units are $10^{-9} \text{s}^{-2}$. The latter two fields have blank areas near the mountains because of the blocking effect mentioned above. Also, they do not include areas west of 90°W or north of 45°N as they are not essential to this study.

It is interesting to watch in the vorticity patterns the development of the features which were noted in previous sections. At 0000 GMT on the 14th (figure 40), the differences between the two cases are not yet very strong. M has a slight northward shift and a small positive perturbation in West Virginia. This and the appearance of a separate center of negative vorticity in western Pennsylvania seem to indicate the beginning of a couplet which might be expected from the vortex tube compression and expansion hypothesis. The terms in the vorticity equation show only minor differences at this time.

At 0600 GMT (figure 41) the trend becomes more evident. The secondary pool of negative vorticity is now somewhat stronger and is located in central Pennsylvania. The perturbation of positive vorticity is evident in Ohio, but is certainly not an outstanding feature. The only other major difference is the northward shift of positive vorticity in M. The divergence terms (not shown) do possess some differences at this time. In addition to a 350 km east-northeastward shift of the overall pattern in M, a separate maximum has evolved in western Virginia, over the topographic ridge. Since there is no separate vorticity maximum there, it is obvious that there must be
greater convergence there. The problems with this and some possible explanations will be discussed below. There is also a very small positive perturbation in Ohio, perhaps a first indication of lee trough formation there but also likely a reflection of the coincident perturbation of positive vorticity. The patterns of the tilting term contributions are quite different between the two cases. The maximum magnitudes are much smaller than the maximum divergence term values, but they occur, in some cases, where the divergence term is very small, making their contribution important. One important such region is in the coastal area of Virginia and North Carolina, where values in M are twice as great as those in NM and comprise up to 50% of the net total rate of change.

All these features are present in a more definite form at 1200 GMT (figures 42, 45, 47 and 49). The couplet of negative and positive vorticity in M east and west of the mountains is quite evident now, in sharp contrast to NM (even allowing for latitudinal displacement). There is still, however, only one maximum, located in southern West Virginia, 200 km north of the NM maximum. The patterns of the divergence effect (figure 45) show some complex differences. The NM forecast possesses a smooth pattern with a broad maximum over southern Virginia. M is virtually identical over the ocean and immediate coast, but possesses an axis of decreased positive convergence contribution on a line from eastern Pennsylvania to central North Carolina. This is the axis of positive values of OROWS and is evidently a reflection of the upslope divergence effect modifying the synoptic scale convergence. The M maximum in positive vorticity
tendency due to the divergence term is located in western Virginia and is skewed to the northwest in general and into West Virginia and southeastern Ohio in particular. Figure 26b shows this to be a region of moderate orographically forced downward motion, consistent with the vortex tube stretching argument presented earlier. The areas of downward motion over New England and southern Canada, however, are accompanied by only weak positive vorticity tendencies (though they are slightly higher than the values in NM). The apparent contradiction is easily explained by the fact that the absolute vorticity, a multiplicative factor in the divergence term, is lower by about a factor of two.

The maximum positive divergence effect in M is notably larger than that in NM. By dividing out the corresponding absolute vorticities, one finds that the convergence over the Virginia-West Virginia border in M is about one-third again as great as that in NM. Evidently this added convergence is not orographically forced via the vortex stretching mechanism, since OROWS is weakly positive here (see figure 26b). If it were due to greater synoptic scale forcing, the upward motion should be enhanced throughout a fairly large depth and there should be enhanced rainfall at that point. Figure 26a shows that there is no such enhancement, at least at the point of maximum M divergence effect. Equivalently, it shows that the convergence is shallow in nature. What it appears we are left with, then, is an estimate of the added frictional convergence due to the rougher topography. (Of course, such convergence cannot act to increase the vorticity to a greater extent than the friction reduces it.)
This is the first time, to the author’s knowledge, that such an estimate has ever been obtained, or even considered, for the NCAR model.

Figure 47 shows that the contributions due to the tilting term are not negligible and that there are some sizeable differences between the two cases. The NM case is much less difficult to interpret. Its outstanding feature is the broad band of negative values north of the cyclone. This is due simply to the tilting, by a synoptic scale $w$, of a vortex tube which results from strong easterly flow at the surface giving way to more southerly and westerly flow aloft. M is a good deal more complex. No explanation can be as clear cut as that in NM. The maximum over eastern Virginia and North Carolina appears to be a result of the increased gradient of vertical motion in the $x$-direction due to OROWS (see figure 26b) acting on the positive vertical gradient of meridional motion there. Negative values do appear immediately adjacent to the mountains, where the gradient of OROWS in the $x$-direction becomes very small and its gradient in the $y$-direction begins to contribute toward the production of anticyclonic vorticity in the manner mentioned above concerning NM. (It is interesting to note that this area of negative values and the negative values south of New England might form a band similar to NM were it not for the north-south band of positive values caused by OROWS.) The positive contribution in Ohio may also be explained on the basis of orographically induced tipping. (Of course, there may be other factors at work, but it would be very difficult, if not impossible, to separate them out.) A vertical gradient of horizontal
wind is created as strong east-northeasterly flow in the first layer gives way to weaker east-southeasterly flow in the second layer. This creates a vortex tube with its axis pointed toward the northwest. The tipping action near the mountains provided by negative OROWS results in the production of positive vorticity, in contrast with the negative production in NM at this point.

The advections at this time (not shown) are negligible in NM. M, however, has some areas of significant vorticity accumulation. Strong easterly flow at about 41°N is resulting in a net \(-1.2\times10^{-9}\text{s}^{-2}\) advection in eastern Ohio. (Similar values should be expected in central to western Pennsylvania, but the lack of a vertical advection term there prohibits the calculation of any net advection.) There is also a small accumulation of positive vorticity in southwest Ohio, northern Kentucky, and southeastern Indiana as the flow is away from the positive vorticity perturbation.

The local rate of change of vorticity, given by the sum of all the various terms, for 1200 GMT is shown in figure 49. The areas of higher positive vorticity production in Ohio and over the coast in M are quite evident. The more rapid eastward and northward advance of the M storm as a whole is also evident. The upslope divergence effect does not actually produce negative vorticity, but it does seem to result in a minimum of positive values along the windward side of the mountains.

There is no need to be as verbose concerning the 1800 GMT patterns of vorticity and vorticity production (figure 55). The features are qualitatively the same as those at 1200 GMT for the most
part. There is, however, at least one small change in the vorticity distribution which may nevertheless represent a significant difference. While the primary maxima of both M and NM have moved slightly north and east, M has developed a second, separate maximum over northeastern-most North Carolina. In some sense then, the M simulation does create two separate centers, albeit in vorticity and not in pressure, and only briefly. The NM maximum clearly remains a single entity. The perturbation in the vorticity field on either side of the mountains in M still remains the greatest distinction between the two cases, however. The feature is even more prominent now than at 1200 GMT. There is also a small but perceptible shift in the couplet in the upstream direction from the topography, more visible now than at 1200 GMT. From experience, this shift is the rule rather than the exception. Though the theoretical reasons for this are quite complicated, a common qualitative explanation is that friction acts in a way such that the air "feels" the presence of the orographic changes before it reaches the topography itself.

A significant increase in vorticity has taken place over Ohio and Lake Erie. This is somewhat contrary to the tendencies indicated at 1200 GMT, but by averaging in the tendencies at 1800 GMT to get an estimate of the six hour tendency, the change is in good agreement. Also noticeable is the increase in the actual peak vorticity in M from 8x10^{-5} s^{-1} to 10x10^{-5} s^{-1}, while the peak in NM has dropped from 9x10^{-5} s^{-1} to 8x10^{-5} s^{-1}. The anticyclone to the west is again observed to be advancing faster in M and cyclonic vorticity is appearing in the lee of the Rocky Mountains.
The 1800 GMT divergence term (figure 46) is also showing two distinct maxima in M. The smaller maximum off the Virginia coast is in good agreement with the NM maximum there, although somewhat stronger. The axis of a local minimum in the M divergence term is still present east of the mountains. The primary M maximum has not shifted its position since 1200 GMT, but is now extremely intense, almost $8 \times 10^{-9} \text{s}^{-2}$ or an increase of more than $17 \times 10^{-5} \text{s}^{-1}$ is sustained over six hours. (Some of this convergence is undoubtedly frictional in nature and therefore really represents a net dissipation.) This maximum, again, shows good agreement with the positive precipitation anomaly (see figure 27a) and especially the areas of negative orows (see figures 27b and 28b). An example of the more direct divergence effects induced by flow over topography can be found at this time over the southern Appalachians, where there is no precipitation (see figure 13a). Upward forcing over northern Georgia and Alabama and southern Tennessee (see figure 27b) is resulting in somewhat greater production of negative vorticity. The effects on the downslope side are not as dramatic, but there is a perceptible decrease in negative vorticity production over South Carolina as compared to NM. It is also interesting to note how small the differences are between M and NM where the added forcing is negligible, i.e., over flat areas such as ocean or the Great Lakes region.

The patterns of the tilting effect are displayed in figure 48 and show little change from the 1200 GMT distribution. The M maximum over Ohio has disappeared, however, and the positive values over the east coast have intensified, with a maximum now in excess of
1.5x10^{-9} s^{-2} over northeastern North Carolina.

The advections at this time are fairly significant in M but limited in extent. Over southwestern and south central New York, northeastern Ohio, and northwestern and north central Pennsylvania, the various advections sum to, at most, -2.5x10^{-9} s^{-2}, primarily due to u-advection from the area of negative vorticity in eastern Pennsylvania to the perturbation of positive vorticity in Ohio. To varying degrees, depending on position, this offsets the large positive tendencies generated by the divergence term. The only other area of significant advections is over eastern and south-eastern North Carolina amounting to -1.5x10^{-9} s^{-2} in M and about -1x10^{-9} s^{-2} in NM.

The sum total of all the terms at 1800 GMT is shown in figure 50. The positive vorticity production over Ohio is evident, as well as the areas of net negative advection over western New York and central Pennsylvania. Evidently, the tilting term contribution was important in M, since a minor axis of positive net tendencies along the coast in M has replaced negative values there in NM. In both cases, the greatest net tendencies are located in the Chesapeake Bay region, but the M maximum is approximately 50% greater. This is due, in part, to a greater divergence effect there, but also to a greater tilting term contribution (on the order of 1.5x10^{-9} s^{-2}).

Figure 44 illustrates the vorticity pattern for 0000 GMT on the 15th of March, after the storm center itself has moved out over the water. Both the M and NM vorticity maxima remain inland. The terrain induced perturbation of vorticity is still present and quite strong.
There is now less difference in the peak magnitudes of the two cases, but some local differences are quite large, particularly over Ohio, where differences exceed $5 \times 10^{-5} \text{s}^{-1}$. The southern portion of the M anticyclone has advanced further east, and the entire anticyclone appears stronger despite its lower central pressure. And, again, it is interesting to note how similar the two cases are over open water and away from any terrain. This is perhaps the best evidence we have that the mountains did indeed have some impact. And as far as a qualitative inspection can tell, they improved the forecast.

This tendency for only minor differences over the ocean is also evident in the distribution of values of the divergence term (not shown). All the notable differences occur over land. In particular, there is an axis of higher M values from Washington, DC, through western Pennsylvania, and north into southern Ontario. A reference to figure 28 suggests that both the precipitation enhancement and purely orographic mechanisms are important, the former over Maryland and Pennsylvania and the latter north through Buffalo, where the precipitation differences are negligible. Vorticity changes due to purely orographic forcing have become somewhat greater east of the mountains also, with higher magnitudes over New England, eastern New York, and eastern Pennsylvania. The tilting term values remain qualitatively unchanged from 1800 GMT. The magnitudes of the differences over the coast are somewhat higher, particularly over the New York City area. The differences in the area of the lingering trough are, again, quite small. The patterns of total local rate of change of vorticity also display little change from 1800 GMT. Both
M and NM have maxima which have moved east but M, again, has a second maximum of equal strength over eastern Virginia, NM does have a lobe of positive values in virtually the same position, but it is not strong enough to constitute a second maximum. An area of negative tendency persists over central and western New York and Pennsylvania in M, a feature which is totally absent from NM. This is primarily a result of negative vorticity advection in easterly flow over this region and is the biggest difference between the two cases. There is some lesser net positive advection vorticity advection over southeastern New York adding to the positive increase due to the tilting term there. To the west, the large area of positive tendencies at 1800 GMT is much smaller and weaker and is located over eastern Lake Huron.

Thus, by looking at the low level vorticity, it has been possible to quantitatively track a particular aspect of the storms rather than note, in a purely qualitative manner, the presence of that secondary trough upon inclusion of the mountains. One can watch it evolve in time and also say something about why it develops the way it does. Contrary to most traditional scaling arguments dealing with larger scale circulations, it was found that the tilting term did contribute significantly, at least on smaller scales near the mountains. Changes in the distribution of latent heating due to variations in storm track and, to a lesser degree, orographic lifting acted in conjunction with downslope motion to produce low level convergence and positive vorticity over the mountains and to the northwest.
The upslope divergence effect was not quite so obvious in either the pressure or the vorticity fields. The sharp ridge east of the mountains in the observed analyses was not simulated, but the nature of the model storm was too different from the observed storm to make any direct comparisons. A comparison between the two simulations seems much more justifiable, although of questionable application to the real atmosphere. However, some physically reasonable explanations for the differences between the two simulations can be put forward. Those given here represent the most detailed examination to date of output from this version of the LAM.
VIII. CORRELATIONS

Correlations between several pairs of variables were computed and examined versus time and versus height. With a null hypothesis of "no correlation", the correlation coefficients were tested for significance which, in turn, would tell us if indeed there is a relationship. While no shattering revelations ensued, some interesting relationships did appear that had not been investigated before. Of course one must keep in mind that the correlations themselves imply no causal relationships. Furthermore, all the quantities used here are derived from model data because of their much greater accessibility, and therefore any insights they provide are strictly applicable only to an examination of the model. The degree to which one accepts these correlations as indicative of the real atmosphere's behavior is purely a subjective matter, but surely they represent a reasonable and readily available first guess.

Following Mack (1966), the correlation coefficient, $r$, between two sets of numbers, $x_i$ and $y_i$, is given by

$$r = \frac{\sum_{i=1}^{N} (x_i - \bar{x})(y_i - \bar{y})}{\left(\sum_{i=1}^{N} (x_i - \bar{x})^2 \sum_{i=1}^{N} (y_i - \bar{y})^2\right)^{\frac{1}{2}}}$$

where $N$ is the number of pairs, $(x_i, y_i)$ and $\bar{x}$ and $\bar{y}$ are the means of the $x_i$ and $y_i$, respectively.
For a normal bivariate distribution of pairs of numbers the test function for the significance of a correlation, \( r \), is

\[
    c = r \sqrt{N-1}
\]

If \( c \) exceeds 1.96, 2.58, or 3.29, the hypothesis that \( r=0 \) can be rejected at about the 5%, 1%, or 0.1% probability level, respectively. Furthermore, one can test the hypothesis that the population correlation coefficient equals \( \rho \) by computing

\[
    z = \frac{1}{2} \ln \left( \frac{1+r}{1-r} \right) \quad \text{and} \quad z_0 = \frac{1}{2} \ln \left( \frac{1+\rho}{1-\rho} \right)
\]

and applying the test function

\[
    c = (z-z_0) \sqrt{N-3}
\]

for the same confidence limits. These formulae can also be used to construct confidence intervals for a given correlation. This will be the case here, since we are not hypothesizing any particular values.

The problem with these tests is that their utility rests on two assumptions that may or may not be satisfied:

1. The members of the sets from which values for the pairs are drawn are independent. It is questionable whether we can satisfy this requirement strictly due to the continuous nature of the fields involved.
(2) The values for each of the two variables in each pair comprise a normal population. Precipitation, in particular, is a problem here because of the large number of points with no precipitation in addition to points with, conceivably, large values.

The latter was dealt with as best possible following some advice from Professor E. N. Lorenz. Fields suspected of having a high degree of skewness (such as precipitation) were transformed into a more "normal" sample by taking the fourth root of the values. In some cases, however, the proportion of zeroes was so large that any attempt at normalizing the distribution would be fruitless. The clearly unmanageable fields were omitted, and those that are merely questionable are commented on in the text.

The problem of independence remains. The area chosen for the correlations was centered on the mountains and covered the area from 28.75°N to 43.75°N and 68.76°W to 95°W. This encompassed 286 grid points (in the horizontal plane) and therefore 286 pairs for each correlation. With 285 degrees of freedom and a confidence limit of 95%, a correlation of +0.116 is necessary to reject the hypothesis of no correlation. These pairs, however, are clearly not independent, particularly on a fine mesh grid, i.e., 1.25°. Suppose, however, that we select 32 points at random within the region. We can probably be reasonably sure that these points are independent. We have cut the number of degrees of freedom by a factor of nine and now require a correlation coefficient of +0.348 to reject the null hypothesis (r=0) at the 5% probability level. As shall be seen, several of the corre-

\footnote{Personal communication}
lations will fall below this level. But we may wish to accept them anyway, because many are consistently at the same value throughout the period and/or in nearby layers. The only remaining problem encountered in this approach is that of importance. While we may accept a particular correlation as being real (as opposed to chance), we may not regard it as truly meaningful if the correlation is too weak.

One last note involves the dependence of the correlation on the area chosen. By appropriate choice of area, one could, of course, get a very good correlation. But the questions of dependence and the number of degrees of freedom arise again as the area becomes more limited. And it may actually be preferable to obtain the relationships on the basis of the entire storm anyway.

With these problems in mind, we proceed to an examination of the correlations themselves. Figure 51 plots the correlation between precipitation and the three hour mean condensational heating rate, $\bar{Q}_c$, at the levels indicated (i.e., the height of the midpoint of each layer in meters). The lower five levels are shown in figure 51a and the next higher six levels are shown in figure 51b. The trend that immediately meets the eye is the tendency for correlations to increase with time below 5625 m and to decrease with time above 5625 m. The development of the storm during the 24 hour period ending at 0000 GMT on the 15th of March has been described in detail in a previous section. The storm lies mostly in the study area throughout this period.

What is probably happening is that, because of the moisture
initialization, a great deal of lifting is required to achieve the necessary 95% saturation and therefore "rain" falls from higher levels first. This is evident to some degree in the cross sections. The trends are a consequence of the lower levels becoming increasingly saturated. The effect is amplified if the higher actual density of water vapor at lower levels is considered. There are a number of other comments that should be made in reference to figure 51. First, the tendencies at upper and lower levels should be expected to be opposite, particularly if there is a tilt to the saturated "column". As the precipitation pattern becomes increasingly affected by low level moisture (i.e., the low level correlation increases), and unless conditions change drastically enough at upper levels to overshadow it, the correlation at upper levels must decrease. Secondly, in relation to this, the contrasting tendencies at upper and lower levels are made more visible by plotting them alongside one another than if the tendency at any particular level was presented alone, particularly since the tendencies are fairly small. That is, the tendencies are more likely to be accepted as tendencies and less likely to be discounted as too weak to be significant. This leads to the third comment. Although almost all of the correlations here may be considered significant by the test mentioned earlier, their consistency in time and in the vertical definitely lend support to this conclusion. This principle is applicable to most of the rest of the correlations. It has been explained in more detail here at the outset to save repeating in each instance.

Figure 52 presents the same data, but plotted versus height for
each time period as indicated (in GMT time). The general trend
toward uniform correlation with height is somewhat clearer. The
behavior at the lowest two levels should not be taken too seriously.
The regions of non-zero $\bar{Q}_C$ are quite limited in some cases, particularly
during the early stages of the forecast when the lower layers are
not yet saturated (see figures 31-39).

Figure 53 is a plot versus time of the correlation between pre-
cipitation and the three hour total number of moist convective adjust-
ments in a particular layer. Despite the poor formulation of the
convective adjustment procedure (particularly the absence of a satu-
ration criterion) and the lack of knowledge about the magnitude of
the adjustment, this correlation was computed in hopes of determin-
ing the degree to which convection might affect the precipitation
distribution. Layers above the $m=9$ level are not included because
there are a very small number of adjustments above 7 km. Even within
the levels examined, the reliability of the correlations seems to
decrease with height, based on their greater fluctuation in time.
Otherwise the general trend seems to be toward increasing correlation
with time and with height. Both are fairly slight, however. This
might be expected in the vertical since the temperature, and its
alteration, affects the stability of two layers, and hence creates
a type of dependence in the vertical. While no explanation could
be found for the 0300 GMT minimum, the following interpretation is
offered regarding the slight dip between 1200 and 1800 GMT. Some of
the variation in the precipitation pattern is due to an orographic
lifting, detracting from its correlation with the moist convective
adjustments. Also, the storm is located over land and the stratification is more stable without an oceanic heat source at the surface.

Figure 54 plots the same correlation coefficients versus height. Aside from the comments made already, only the large increase in correlation at upper levels at 1800 GMT is very obvious. This results in a shift in peak correlation upward, from about 3.6 km to 5.6 km. This may be related in some way to the rapid deepening between 1500 and 1800 GMT. There was also a band of numerous adjustments (not shown) which is not producing any precipitation at the old cold front to the east. This area stopped recording convective adjustments in this period and, hence, ceased to contribute negatively to the correlation. Although most of the values exceed the critical (significant) level according to the test, this actually tells us little about the role of convection in producing precipitation. Some degree of positive correlation should be expected, since a region of upward motion is favorable for both precipitation and, in the model, moist convective adjustments. A saturation criterion and an indication of the quantity of additional precipitation induced by the adjustment could yield much more meaningful results.

Figure 55 is an effort to determine the degree and trend of orographically forced precipitation, with due consideration taken concerning causality. Synoptic scale ascent is occurring over much of the area. There are, however, regions where no condensation is taking place, yet there are orographically induced vertical motions. These points contribute negatively to the correlation and this must be kept in mind. One encouraging factor is, however, that OROWS is a very nearly normally distributed quantity, far better than almost
any other field.

As might have been expected, there are higher correlations present in lower levels. Although the correlations for the levels above 4125 m are probably not to be trusted (not significant), it is interesting to note that they continue to be positive for all the levels and decrease with increasing height, at least in the latter half of the period. Part of the general positive correlation is obviously the coincidental placement of the maximum in synoptic scale ascent and precipitation over the region of upward forcing, and this should also be kept in mind. One noticeable trend is in the time of maximum correlation for a particular level. Although in toto, the highest correlations appear during the time period when the storm is passing over the mountains, it seems that the lower levels experience earlier peaks. One possible explanation may be the decreasing stability of the environment as the storm both intensifies and moves into the study region and becomes centered on the mountains. The forcing could then be felt to a greater degree at higher levels. But this tendency to peak later at higher levels actually argues against a purely coincidental correlation due to concurrent synoptic and orographic lifting. Since the saturated "column" (actually more like a slab) tilts northward with height, and the storm has a northward component of motion, one might expect, coincidentally, an earlier peak in correlation at upper levels as the "column" moves northward over the OROWS maximum. This suggests that the orography is having at least some direct effect on the precipitation.

The general decrease in correlation after 1500 GMT is probably
a result of the bulk of the precipitation moving out to sea, but the earlier decrease at the first three levels remains unexplained.

Figure 56 is the corresponding plot versus height for the times indicated. It shows, as well, the upward progression of the peak correlation. The most consistently significant correlations appear to be at about the 2km level.

The correlation between OROWS and the three hour total number of moist convective adjustments is plotted versus height in figure 57. None of the correlations exceed the more stringent significance level mentioned earlier (r=0.348). Even if significant, the correlations are very weak.

Until 1800 GMT, the tendency seems toward increasing correlations aloft and decreasing values at lower levels, reversing the "tilt" of the plots. It is very curious that the highest correlations are found, on average, at about the 6 km level. One might expect the extra orographic uplift to add latent heat at low levels and reduce the stability there. It is hard to imagine how weak orographic forcing could affect processes at 6 km. Some coincidental explanation could be found, such as the following. Referring to the 1200 GMT 500 mb field in figure 16b, significant cold advection is seen from Alabama west and Mississippi south and also over eastern Virginia and North Carolina. If these advections increase with height, the destabilization could set off convective adjustments in approximately the same regions where OROWS is upward (see figure 26b or 27b) however indirectly related.

Figure 58 is very simple, plotting versus time the correlation
between OROWS and precipitation. The dashed lines delineate the more liberal estimate of the 95% confidence interval ($\approx r^{+} .12$), which assumes independence of neighboring grid points, obviously a questionable assumption. In fact, the correlations are quite low throughout the period, in agreement with the earlier conclusion that direct orographic effects are fairly small compared to synoptic scale forcing. But the persistence and continuity of the correlations are strongly indicative of, at least, a weakly positive correlation. Figure 58 is, in a sense, a composite of the various levels in figure 55. The low correlation early in the period should be largely explicable by the non-saturation at low levels, displayed well in the cross sections and indicative, again, of the poor moisture initialization. And since OROWS is a maximum at the ground, the physical mechanism one might think of as leading to a sizable positive correlation does not apply. Interestingly enough, the correlation does not drop off drastically late in the period. Evidently, the orographic motions are associated with variations in precipitation in the wake of the storm such as were noted earlier. And this is despite the fact that the bulk of the precipitation, by far, is falling out over the ocean.

The remaining figures are the NM analogs to the preceding figures. Figure 59 is a plot versus time of the correlation between $Q_c$ and precipitation for various levels. Aside from the lowest three levels during the first half of the period, there are no differences one might consider significant, since the confidence interval is at least $0.112$ and probably larger. The correlations in the bottom two layers may be significantly lower than in M, and may indi-
cate some degree of orographic enhancement of humidity. But the early
values of humidity in these layers were too low to amount to either
a substantial or a normally distributed sample. Figure 60 is the plot
versus height, also showing no significant differences.

Figures 61 and 62 correspond to figures 53 and 54, dealing with
the correlation between moist convective adjustments and precipitation.
Considering the higher variability of these correlations, the two
cases are remarkably similar. The only conspicuous difference, though
perhaps not significant, is that the M values in the very earliest
time periods are almost uniformly higher. Considering also the parti-
cular difficulties with the convective adjustments mentioned above,
it is perhaps best not to try to explain this. One comment that can
be made concerning the plots is that the dip in the correlation
around 1200 and 1500 GMT is present to some degree in NM as well as M.
This indicated, then, that perhaps the dip is likely not due to oro-
graphic influence on the precipitation pattern, but to some factor more
or less common to the two areas. Lastly, referring to figure 62,
the only notable difference is the lack of a strong maximum correla-
tion at upper levels late in the NM storm which, again, will not
be commented on.
IX. SUMMARY

It has been recognized for some time that the Appalachian Mountains must play a role in the development of east coast cyclones. Colucci's (1976) results are very suggestive in this regard and represent the greatest detail with which the climatology of these storms has been examined. To date, numerical prediction models have typically had a great deal of difficulty in simulating these complex situations. A new, very fine mesh, limited area version of the NCAR General Circulation Model has been applied to a storm of this type. The model has been described in general here, and details may be found in Oliger et al (1970).

A storm was chosen and analyzed which appeared to be strongly affected by the Appalachian Mountains. The cyclone was typical of winter east coast cyclones, according to Colucci's climatology, and exhibited many features shown by synoptic experience to be characteristic of such storms. Heavy precipitation accompanied the storm, particularly in association with orographic and cold frontal lifting. A strong stationary frontal band evolved near the southern tip of the mountain range followed by the formation of a secondary cyclone. This secondary eventually became the dominant center as the primary filled west of the mountains. Throughout the entire period, an inverted ridge was present east of the mountains over the middle Atlantic states.

Unfortunately, the model did not accurately simulate this storm in either the M or the NM case. Specifically, the model failed to simulate:
(1) a primary cyclone which tracks west of the mountains and fills,
(2) a secondary storm which forms on the plains of the Carolinas,
(3) a strong stationary front on the southern edge of a shallow cold dome of high pressure east of the mountains,
(4) heavy precipitation in association with a cold front moving through the Gulf States, and,
(5) heavy upslope precipitation in easterly flow forced over the mountains.

Several reasons for these shortfalls have been offered, but one of the most important was probably the poor moisture initialization. The initialization at climatological values, while reasonable for a large number of 2.5° hemispheric forecasts\(^1\), was very unrealistic in this instance. It also delayed the onset of significant precipitation and its very important effect on low level convergence (Danard, 1964). Insufficient strength of the mid-tropospheric trough was also noted, resulting in less forcing due to differential vorticity advection and warm advection.

Of the two simulations, the one incorporating orography was superior. It reproduced, to some degree,

(1) a trough on the western slopes of the Appalachian Mountains,
(2) some enhancement of precipitation in upslope flow,
(3) a more accurate distribution of high pressure systems adjacent to the cyclone, and
(4) a semblance of the cold dome near the surface east of the mountains.

\(^1\)Baumhefner, D.P.; personal communication
Because the forecasts were not of sufficient quality for use in a detailed analysis of the actual storm, the emphasis in the study was shifted toward an investigation of the differences between the two cases, i.e., the sensitivity of the model to the inclusion of orography. When justified, conclusions concerning the model behavior were extended to the real atmosphere.

The primary difference between the two cases was the presence in M of a trough northwest of the storm center and west of the mountains. An analysis of the low level vorticity made this even more obvious. Superposed on a general northward translation of the system in M is a perturbation in the vorticity field, with enhanced anticyclonic vorticity east of the mountains and enhanced cyclonic vorticity west of the mountains. The principal mechanism believed to be responsible for this configuration is the conservation of potential vorticity in easterly flow over the topographic barrier. The tilting term contribution in this process, often neglected on the basis of scaling for larger scale circulations, underwent changes that could also be considered important in the production of relative vorticity about the vertical axis. Application of these findings to the actual storm must be considered quite limited. Such a potential vorticity conservation effect may have added to the observed pattern, but the actual distributions of pressure and vorticity, though similar to the model patterns, are clearly of different origin.

Examination of the distributions of OROWS, the differences in precipitation, and differences in the three kilometer vertical motions indicated that orographic lifting did result in a noticeable enhance-
ment of precipitation, though of much smaller magnitude than the changes induced by altering the synoptic scale vertical motion pattern.

The cold perturbation east of the mountains in M developed in the face of warm advection. Since OROWS, and hence adiabatic cooling, are a maximum at the surface, and the low level inversion in this region was intensifying (as shown by cross sections), it was concluded that adiabatic cooling due to orographic lifting was probably the major factor. While such a mechanism may be present in the real atmosphere, the observed perturbation was much stronger and must be a product of other influences as well.

Overall, this experiment showed that the NCAR LAM is not yet at a level of sophistication and/or resolution which will enable it to forecast accurately the evolution of this type of storm and its associated smaller scale features. While in both cases the model storm tracked too far south, the model did seem sensitive to even low topography, and the inclusion of this topography resulted in a slightly superior forecast. The results also seem to suggest a greater sensitivity of the model to the moisture initialization, and more work should probably be directed toward that area.
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Figure

44a. Same as figure 40a, except for 0000 GMT, March 15, 1975.

44b. Same as figure 40b, except for 0000 GMT, March 15, 1975.

45a. (top) Contribution toward vorticity tendency due to the divergence term (see text), 1200 GMT, March 14, 1975, M case. Contour interval of $0.5 \times 10^{-9}$ s$^{-2}$. Negative values dashed.

45b. (bottom) Same as figure 45a, except for NM case.

46a. Same as figure 45a, except for 1800 GMT, March 14, 1975.

46b. Same as figure 45b, except for 1800 GMT, March 14, 1975.

47a. (top) Contribution toward vorticity tendency due to the tilting term (see text), 1200 GMT, March 14, 1975, M case. Contour interval of $0.25 \times 10^{-9}$ s$^{-2}$. Negative values dashed. Blocked areas omitted.

47b. (bottom) Same as figure 47a, except for NM case.

48a. Same as figure 47a, except for 1800 GMT, March 14, 1975.

48b. Same as figure 47b, except for 1800 GMT, March 14, 1975.

49a. (top) Local rate of change of vorticity (sum of terms; see text), 1200 GMT, March 14, 1975, M case. Contour interval of $0.5 \times 10^{-9}$ s$^{-2}$. Negative areas dashed. Blank areas are where any term is blocked by topography.

49b. (bottom) Same as figure 49a, except for NM case.

50a. Same as figure 49a, except for 1800 GMT, March 14, 1975.

50b. Same as figure 49b, except for 1800 GMT, March 14, 1975.

51a. Plot versus time of the correlation between $Q_{ci}$ (height in meters) and precipitation, M case. Five lowest layers ($m=1-5$).

51b. Same as figure 51a, except for layers $m=6-11$.

52. Same as figure 51 except correlation is plotted versus height at times indicated (GMT).

53a. Plot versus time of the correlation between total three hour precipitation and the three hour total number of moist convective adjustments at a point in a layer, M case. Lowest five layers ($m=1-5$).
Figure

53b. Same as figure 53a, except for layers m=6-9.

54. Same as figure 53, except correlation is plotted versus height at times indicated (GMT).

55a. Plot versus time of the correlation between OROWS and $Q_{ci}$, M case. Lowest five layers (m=1-5).

55b. Same as figure 55a, except for layers m=6-11.

56. Same as figure 55, except correlation is plotted versus height at times indicated (GMT).

57. Plot versus height of the correlation between OROWS and the three hour total number of moist convective adjustments at a point in a layer, M case.

58. Plot versus time of the correlation between OROWS and total three hour precipitation, M case.

59a. Same as figure 51a, except for NM case.

59b. Same as figure 51b, except for NM case.

60. Same as figure 52, except for NM case.

61a. Same as figure 53a, except for NM case.

61b. Same as figure 53b, except for NM case.

62. Same as figure 54, except for NM case.
Table 1
Characteristics of the Second Generation NCAR GCM
Limited Area Version (LAM)

<table>
<thead>
<tr>
<th>1) Vertical coordinate</th>
<th>geometric height</th>
</tr>
</thead>
<tbody>
<tr>
<td>2) Height of top</td>
<td>18 km</td>
</tr>
<tr>
<td>3) Vertical resolution</td>
<td>24 layers</td>
</tr>
<tr>
<td>4) Horizontal domain</td>
<td>10°N–75°N, 50°W–140°W</td>
</tr>
<tr>
<td>5) Horizontal resolution</td>
<td>1.25°</td>
</tr>
<tr>
<td>6) Horizontal diffusion</td>
<td>variable eddy viscosity</td>
</tr>
<tr>
<td>7) Vertical diffusion</td>
<td>none in this experiment</td>
</tr>
<tr>
<td>8) Horizontal approximation of derivatives</td>
<td>second-order</td>
</tr>
<tr>
<td>9) Moisture processes</td>
<td>explicit hydrological cycle, convective adjustment</td>
</tr>
<tr>
<td>11) Surface hydrology</td>
<td>explicit prediction of snow cover and soil moisture</td>
</tr>
<tr>
<td>12) Surface temperature calculation</td>
<td>calculated as a function of s-level and ground temperatures (see Oliger et. al., 1970)</td>
</tr>
<tr>
<td>13) Cloudiness</td>
<td>diagnostically determined from relative humidity and vertical motion at 3 km and 9 km</td>
</tr>
<tr>
<td>14) Boundary layer</td>
<td>bulk transfer coefficient and Deardorff formulation</td>
</tr>
<tr>
<td>15) Orography</td>
<td>realistic consistent with resolution</td>
</tr>
<tr>
<td>16) Ocean assumption</td>
<td>climatological surface temperature</td>
</tr>
<tr>
<td>17) Moisture initialization</td>
<td>climatological monthly zonal mean at all levels</td>
</tr>
</tbody>
</table>
Table 1 (con't)

<table>
<thead>
<tr>
<th>No.</th>
<th>Description</th>
<th>Details</th>
</tr>
</thead>
<tbody>
<tr>
<td>18)</td>
<td>Temperature initialization</td>
<td>At s-level: monthly mean ocean surface temperature over ocean. Over land, extrapolation downward from first layer assuming 5°C km⁻¹ lapse rate. At upper levels: computed hydrostatically from analyzed pressure field; temperature of ground surface initialized at temperature of s-level</td>
</tr>
<tr>
<td>19)</td>
<td>Initialization</td>
<td>modified divergence (see Washington and Baumhefner, 1975)</td>
</tr>
</tbody>
</table>
Figure 1.
Figure 3.
Figure 4.
Figure 6.
Figure 7.
Figure 9.
Figure 11.
Figure 15.
Figure 16c.
Figure 17c.
Figure 18c.
Figure 19 (top).  Figure 20 (bottom).
Figure 21 (top).  Figure 22 (bottom).
Figure 23 (top). Figure 24 (bottom).
Figure 29.
Figure 32.
Figure 32c.
Figure 34.
Figure 34c.
Figure 35.
Figure 35c.
Figure 36c.
Figure 37c.
Figure 38.
Figure 38c.
Figure 39.
Figure 39c.
Figure 40.
Figure 44.
Figure 51b.
Figure 52.
Figure 53a.
Figure 53b.
Figure 54.
Figure 55a.
Figure 55b.
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Figure 57.
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Figure 60.
Figure 61a.
Figure 61b.
Figure 62.
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