REVIEW

Regional Models of the Atmosphere in Middle Latitudes

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ABSTRACT

This review describes recent development in operational and research limited-area numerical weather prediction models in middle latitudes. The current skill of limited-area models is summarized through the use of conventional measures of verification such as $S_o$ scores, root-mean-square errors and correlations between forecast and observed changes. Additional measures of verification, which measure the skill or realism of regional models in reproducing atmospheric structure on those scales, are discussed. Use of a uniform set of verification measures such as those discussed here would facilitate model comparisons and assessment of the impact of changes in model components on short-range (0–48 h) forecasts.

Three major components of regional models are discussed. These include numerical aspects (e.g., the grid structure, boundary conditions and the approximations to the analytic differential equations), physical aspects (modeling surface and boundary layer processes, condensation and evaporation, and radiation) and the analysis and initialization procedure. The paper emphasizes the impact of these components on the forecast rather than the details of each component.

The main conclusion of the paper is that further increases in the overall skill of operational regional forecasts are likely to occur through improvements in all the components of limited-area models. Improvements in various components developed and tested in research models are currently being incorporated in several operational models, and some modest but significant improvements in regional forecast skill are likely over the next five years.

1. Introduction

Operational numerical weather prediction of synoptic-scale (characteristic wavelengths greater than 2500 km) atmospheric flows began in the United States in 1954 with a barotropic model (Fawcett, 1977; Shuman, 1978). Multilevel, baroclinic models were developed gradually in the 1960s and 1970s. With the operational implementation of these models came a slow, but steady improvement in the skill of large-scale forecasts of sea-level pressure and 500 mb heights and winds. This improvement could be related mainly to higher resolution (vertical and horizontal), increases in domain from hemispheric to global, and improvement in analysis and initialization techniques and physical parameterizations.

Despite the improvements in the skill of predicting the large-scale wind and pressure patterns, improvements in predicting precipitation, which exhibits greater variability on scales much smaller than the synoptic scale, have been very slow (Committee on Atmospheric Sciences, 1980; Ramage, 1982).

The long-realized fact that much significant weather, in addition to precipitation, occurs on the mesoscale (Orlanski, 1975) stimulated considerable research on limited-area, or regional numerical modeling when computers became large and fast enough to make such high-resolution feasible. However, the extension of synoptic-scale models, with typical horizontal grid sizes of 400 km, to regional models, with grid sizes from 50–250 km, has not been straightforward. There are four principal difficulties: 1) numerical difficulties associated with lateral boundary conditions, 2) limitations in the density of initial data, particularly upper-air data, on the regional scale, 3) the increase in importance of and difficulty in modeling diabatic, topographic, and surface effects, and 4) the reduction in the inherent predictability of smaller-scale atmospheric circulations. The first two are solvable with existing technology, requiring only implementation of existing computer and advanced observational systems. Considerable progress in developing improved models of physical effects of terrain, surface and boundary layer fluxes of momentum, heat, and water vapor, latent heating associated with condensation and evaporation, and radiative effects has also been made in research models. Thus, we may hypothesize that the only fundamental dif-

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difficulty blocking significant improvements to regional-scale, or meso-\(\alpha\) scale, predictions lies in the increasingly stochastic nature of smaller-scale phenomena, as discussed by Tennekes (1978). While the ultimate limits to predictability of regional-scale weather are unknown, I believe there is evidence that significant improvements in predictions of meso-\(\alpha\) scale phenomena could be made immediately. These improvements could be obtained through a combination of better use of existing data, incorporation of recent research results in operational models, and implementation of an improved data network.

This paper reviews recent operational and research regional-scale\(^2\) numerical models (summarized in Table 1) and discusses their current limitations, and the prospects for improvement of predictions on this scale for time periods of 12–48 h. To limit the scope of this paper, only models that have been tested on real data in extratropical regions are reviewed, despite the fact that models of similar scale are showing promise in the tropics as well (Krishnamurti et al., 1979; Harrison and Fiorino, 1982; Fiorino et al., 1982).

2. Quantitative measures of forecast skill and realism of simulations

Before reviewing the components of regional models, it is worthwhile to review methods of judging the accuracy and skill of models, not only to be able to compare the relative performance of different models but also to evaluate the impact of future changes in the numerical, physical or data aspects of the models. Two nonexclusive types of verifications can be identified, those that measure the skill of forecasts and those that measure the degree to which the model forecast or simulation realistically simulates atmospheric behavior. Examples of the first, more conventional type of verification are \(S_1\) scores, rms errors, and threat scores, which will be reviewed in this section. In addition to these traditional methods of verification, evaluation of mesoscale and smaller-scale predictions and simulations using primitive equation models requires a second type of verification. This is because the amplitude of features predicted by the mesoscale models often becomes larger while the scale becomes smaller. Thus, minor displacement errors in time or space can produce enormous errors at individual points. Despite these large errors, the prediction of features of the correct amplitude and structure in approximately the right region may provide useful information. Examples include the prediction of heavy convective rainfall somewhere in a watershed, the prediction of a major hurricane making landfall within a zone along a coast, or the prediction of severe downslope winds along the front range of the Rocky Mountains sometime during a given 24 h period. An example of the second type of verification is the model’s kinetic energy spectrum. If, over a large number of cases, a model produced a spectrum similar to that of the atmosphere, it would be considered a realistic model in this respect. In analogy with general circulation models, the statistical characteristics of the model atmosphere over an ensemble of forecasts or simulations such as the kinetic energy spectrum may be viewed as the model’s “climate.” A model could have a good mesoscale climate but poor average skill scores.

This section discusses some objective verification techniques that can be used to judge the skill of mesoscale forecasts and the realism of mesoscale simulations. Because of the many degrees of freedom in mesoscale models and the computational expense required to run a large number of forecasts, it would be desirable to develop and utilize these measures early, so that investigators can compute a uniform set of comparable measures of skill in the evaluation of experimental forecasts. The use of the same criteria will enable more effective model intercomparisons and more meaningful evaluation of the impact on the forecast of different data, analysis, and initialization techniques and physical parameterizations.

A summary of useful quantitative measures of forecast skill is presented in Table 2, together with estimates of the current capability of regional models, where available.

\(a.\) Measures of forecast skill

1) \(S_1\) scores

Synoptic-scale models have been verified over the years by calculation of objective indices, or scores that reflect the skill in predicting the mass (pressure or height) fields and the precipitation occurrence and amount. The most common measure of skill in forecasting the pressure or height is the \(S_1\) score (Tewes and Wobus, 1954), which measures the skill in predicting the horizontal gradient of a scalar field. Because of the strong geostrophic (or gradient) relationship between the pressure gradient and large-scale flow in extratropical regions, the \(S_1\) score for pressure is also a good measure of skill in predicting the synoptic scale wind field. The \(S_1\) score is defined as

\[
S_1 = 100 \frac{\sum |e_o|}{\sum |G_o|},
\]

where \(e_o\) is the error of the forecast pressure difference, and \(G_o\) is the maximum of either the observed or forecast difference between two points. The summation is over all points in the verification region.

\(^2\) In this paper, regional-scale models refer to limited-area models with horizontal resolution of 50–250 km. These models have sufficiently fine meshes to resolve many meso-\(\alpha\) scale phenomena, which have characteristic horizontal scales of 250–2500 km (Orlanski, 1975).
<table>
<thead>
<tr>
<th>Model reference</th>
<th>Analysis and initialization</th>
<th>Numerics</th>
</tr>
</thead>
<tbody>
<tr>
<td>LFM (Gerrity, 1977; Newell and Deaven, 1981)</td>
<td>NMC global analysis 1st guess and regional analysis (successive scans)</td>
<td>( \nabla \cdot V = 0 )</td>
</tr>
<tr>
<td>MFM (Hovermale, 1982, personal communication)</td>
<td>Any NMC regional or global analysis</td>
<td>Normal mode</td>
</tr>
<tr>
<td>ANMRC (McGregor et al., 1978; Gauntlett et al., 1978; Leslie et al., 1981)</td>
<td>Successive corrections ( V, T )</td>
<td>None</td>
</tr>
<tr>
<td>British Meteor. Office (Golding, 1982, personal communication)</td>
<td>( p, T, q, V ) of Global O/I</td>
<td>Remove ( \dot{D} )</td>
</tr>
<tr>
<td>Japan Met Agency Tatsumi (1982)</td>
<td>( p, T, q, V ) analysis</td>
<td>Bal Eq. + ( \omega ) Eq. ( \rightarrow V )</td>
</tr>
<tr>
<td>Navy-NORAPS Hodur (1982)</td>
<td>( p, T, q, V ) From 6 h forecast of global model modified by observations of ( T, V )</td>
<td>Several options involving balance equation</td>
</tr>
<tr>
<td>MASS (Kaplan et al., 1982)</td>
<td>LFM first guess, enhanced by surface &amp; significant data</td>
<td>Remove ( \dot{D} )</td>
</tr>
</tbody>
</table>

### A. Operational models

### B. Research models
<table>
<thead>
<tr>
<th>Source</th>
<th>Methodology</th>
<th>Resolution</th>
<th>Data Type</th>
<th>Variables</th>
<th>Model</th>
<th>Filter Type</th>
<th>Porous Sponge</th>
<th>Observation Dependence</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boudra (1981)</td>
<td>NMC analysis first guess, mandatory plus significant level data analyzed on ( \theta )</td>
<td>40, 50, 50 x 40</td>
<td>Polar stereo, unstaggered</td>
<td>15, ( z )</td>
<td>4th order</td>
<td>Leapfrog, time filter</td>
<td>Porous</td>
<td>No</td>
</tr>
<tr>
<td>Pennsylvania State University/NCAR (Anthes et al., 1982a)</td>
<td>LFM first guess, enhanced by surface and significant level data</td>
<td>90, 191, 41 x 41</td>
<td>Lambert conformal, staggered ( &quot;B&quot; ) grid</td>
<td>10, ( \sigma - p )</td>
<td>2nd order</td>
<td>Brown-Campana, time filter</td>
<td>Time-dependent from observation</td>
<td>Yes</td>
</tr>
<tr>
<td>Drexel/NCAR LAMPS (Chang et al., 1981, 1982; Maddox et al., 1981)</td>
<td>NMC analysis first guess, enhanced by mandatory plus significant level data</td>
<td>1.25° lat. ( \times ) 1.60° long.</td>
<td>Lat/long unstaggered grid</td>
<td>14, ( \sigma - z )</td>
<td>4th order</td>
<td>Brown-Campana, time filter</td>
<td>Porous</td>
<td>Yes</td>
</tr>
<tr>
<td>GFDL (Ross and Orlanski, 1982)</td>
<td>NMC analysis ( \rightarrow ) GFDL N40 grid ( \rightarrow ) fine-mesh model ( (D \text{ and } f \text{ interpolated}) )</td>
<td>61.5, 180, 38 x 30</td>
<td>Polar stereo, staggered ( &quot;C&quot; ) grid</td>
<td>N.A., ( z )</td>
<td>2nd order</td>
<td>Leapfrog, time filter</td>
<td>Time-dependent from large-scale model</td>
<td>No</td>
</tr>
<tr>
<td>GFDL N80 (Miyakoda, 1973; Miyakoda and Rosati, 1977)</td>
<td>Mandatory levels of ( \phi, T, q )</td>
<td>135, 150, 160 ( \times ) 160</td>
<td>Polar stereo, unstaggered</td>
<td>9, ( \sigma - p )</td>
<td>2nd order</td>
<td>Leapfrog, time filter</td>
<td>Sponge</td>
<td>Yes</td>
</tr>
<tr>
<td>British Meteor. Office (Carpenter, 1979; Golding, 1982 personal communication)</td>
<td>( p ) obtained from large-scale model</td>
<td>10, 60, 62 ( \times ) 61</td>
<td>Cartesian, staggered ( &quot;C&quot; ) grid</td>
<td>10, ( z - z_c )</td>
<td>2nd order</td>
<td>Semi-implicit, time filter</td>
<td>Time-dependent from larger scale model</td>
<td>Yes</td>
</tr>
<tr>
<td>NGM (Phillips, 1978, 1979, personal communication)</td>
<td>NMC regional O/I analysis</td>
<td>Normal mode</td>
<td>( \Delta s/C_g ) for each grid ( \Delta s )</td>
<td>3 nested grids, 2-way interaction</td>
<td>( \sigma - p )</td>
<td>Eliassen-Lax-Wendroff</td>
<td>Outer grid: symmetry at equator</td>
<td>Yes</td>
</tr>
</tbody>
</table>

**List of Symbols**

- \( \Delta s \): horizontal grid size
- \( C_g \): fastest gravity wave speed
- \( V \): horizontal vector velocity
- \( T \): temperature
- \( p_0 \): surface pressure
- \( q \): mixing ratio
- \( \zeta \): vorticity
- \( l \): mixing length scale
- \( \omega \): \( dp/dt \)
- \( e \): normalized vertical coordinate
- \( C_D \): drag coefficient
- \( K \): diffusion coefficient
- \( D' \): vertical integral of divergence
- \( R_i \): Richardson number
- Kuo: variation of Kuo (1974) cumulus parameterization
- \( \Delta z \): vertical grid spacing
- \( \theta \): potential temperature
- \( \phi \): geopotential height
- \( z_e \): surface elevation

Table 1 continued on page 1310
<table>
<thead>
<tr>
<th>Model reference</th>
<th>Surface heat flux</th>
<th>Surface evaporation</th>
<th>PBL (including surface layer)</th>
<th>Radiation (free atmosphere)</th>
<th>Clouds and precipitation</th>
<th>Internal mixing</th>
<th>Computer time for 24 h forecast (Not including pre- and post-processing)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LFM ( Gerrity, 1977; Newell and Deaven, 1981)</td>
<td>Indirect PBL heating</td>
<td>Sea only</td>
<td>Bulk, $1.5 \times 10^{-3} &lt; C_D &lt; 9.5 \times 10^{-3}$</td>
<td>Yes, 1.44 K/day above clouds</td>
<td>Convective adjustment</td>
<td>Saturation criterion 90-96%</td>
<td>Filter</td>
</tr>
<tr>
<td>MFM (Hovermale, 1982, personal communication)</td>
<td>Sea only</td>
<td>Sea only</td>
<td>Same as LFM</td>
<td>No</td>
<td>No</td>
<td>Kuo</td>
<td>N.A.</td>
</tr>
<tr>
<td>ANMRC (McGregor et al., 1978; Gauntlett et al., 1978; Leslie et al., 1981)</td>
<td>No</td>
<td>No</td>
<td>Bulk, $C_D = 4 \times 10^{-4}$ over land $C_D = 1 \times 10^{-4}$ over water</td>
<td>No</td>
<td>No</td>
<td>Kuo</td>
<td>Saturation criterion 95%</td>
</tr>
<tr>
<td>British Meteor. Office (Golding, 1982, personal communication)</td>
<td>Yes—energy balance</td>
<td>Yes—depends on surface type</td>
<td>Explicit, $K(z_0, R_i)$</td>
<td>Yes—climatological clouds</td>
<td>Yes—climatological clouds</td>
<td>Entrainment cloud model</td>
<td>Saturation criterion 100%</td>
</tr>
<tr>
<td>Japan Met Agency Tatsumi (1982)</td>
<td>Sea only</td>
<td>Sea only</td>
<td>Bulk, $C_D$ varies with stability and wind speed</td>
<td>No</td>
<td>No</td>
<td>Convective adjustment</td>
<td>Saturation criterion 100%</td>
</tr>
<tr>
<td>Navy-NORAPS Hodur (1982)</td>
<td>Sea only</td>
<td>Sea only</td>
<td>Bulk, $C_D = 1 \times 10^{-3}$ over water, $2 \times 10^{-3}$ over land</td>
<td>No</td>
<td>No</td>
<td>Kuo</td>
<td>Saturation criterion 100% evaporation allowed</td>
</tr>
</tbody>
</table>

**A. Operational models**

**B. Research models**
<table>
<thead>
<tr>
<th>Model</th>
<th>Boundary Conditions</th>
<th>Flux Formulation</th>
<th>Saturation Criterion</th>
<th>Physical Processes Considered</th>
<th>Computational Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boudra (1981)</td>
<td>Over water only</td>
<td>Explicit, O'Brien (1970) $K$ profile, prognostic Eq. for $h$</td>
<td>No</td>
<td>Yes</td>
<td>Saturation criterion 100%, no evaporation</td>
</tr>
<tr>
<td>Pennsylvania State University/NCAR (Anthes et al., 1982a)</td>
<td>Over water only</td>
<td>Bulk, $1.5 \times 10^{-3}$ $&lt; C_D &lt; 2.0 \times 10^{-3}$</td>
<td>No</td>
<td>No</td>
<td>Kuo-type (Anthes, 1977)</td>
</tr>
<tr>
<td>Drexel/NCAR LAMPS (Chang et al., 1981, 1982; Maddox et al., 1981)</td>
<td>Surface energy budget</td>
<td>Explicit, O'Brien $K$ profile</td>
<td>No</td>
<td>Yes</td>
<td>Sequential plume model (Kreitzberg and Perkey, 1976)</td>
</tr>
<tr>
<td>GFDL (Ross and Orlanski, 1982)</td>
<td>Yes, assumes constant lapse rate at surface</td>
<td>Yes, assumes fixed surface relative humidity</td>
<td>Explicit, $K(Ri)$</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>GFDL N80 (Miyakoda, 1973; Miyakoda and Rosati, 1977)</td>
<td>Yes, surface energy budget</td>
<td>Yes, constant moisture availability</td>
<td>Explicit, geometric profile of $K$, $C_D \sim 2 \times 10^{-3}$</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>British Meteor. Office (Carpenter, 1979; Golding, 1982, personal communication)</td>
<td>Yes, surface energy budget</td>
<td>Yes, surface energy budget</td>
<td>Explicit, geometric $K$ profile</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>NGM (Phillips, 1978, 1979, personal communication)</td>
<td>Sea only</td>
<td>Sea only</td>
<td>Bulk, $C_D$ varies with geography and wind speed</td>
<td>No</td>
<td>No</td>
</tr>
</tbody>
</table>

**List of Symbols**

- $\Delta z$: vertical grid spacing
- $\theta$: potential temperature
- $\phi$: geopotential height
- $z_s$: surface elevation
- $\Delta z$: horizontal grid size
- $C_D$: drag coefficient
- $K$: diffusion coefficient
- $D^z$: vertical integral of divergence
- Ri: Richardson number
- Kuo: variation of Kuo (1974) cumulus parameterization
- $q$: mixing ratio
- $\xi$: vorticity
- $l$: mixing length scale
- $\omega$: $dp/dt$
- $\sigma$: normalized vertical coordinate
- $p_s$: surface pressure
Table 2: Quantitative measures of evaluating regional model forecasts. Typical values refer to 24 h forecasts.

### A. Skill of forecasts

<table>
<thead>
<tr>
<th>Measure</th>
<th>Typical values (1980–82)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sea level pressure</td>
<td>45</td>
</tr>
<tr>
<td>850 mb height</td>
<td>40</td>
</tr>
<tr>
<td>700 mb height</td>
<td>30</td>
</tr>
<tr>
<td>500 mb height</td>
<td>25</td>
</tr>
<tr>
<td>300 mb height</td>
<td>20</td>
</tr>
</tbody>
</table>

### Rms errors

<table>
<thead>
<tr>
<th>Vector wind (m s⁻¹)</th>
<th>Height (m)</th>
<th>Temperature (°C)</th>
<th>Specific humidity (g kg⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface</td>
<td>2.5</td>
<td>3</td>
<td>4</td>
</tr>
<tr>
<td>850 mb</td>
<td>5</td>
<td>30</td>
<td>4</td>
</tr>
<tr>
<td>700 mb</td>
<td>7</td>
<td>35</td>
<td>3</td>
</tr>
<tr>
<td>500 mb</td>
<td>10</td>
<td>40</td>
<td>3</td>
</tr>
<tr>
<td>300 mb</td>
<td>15</td>
<td>45</td>
<td>4</td>
</tr>
</tbody>
</table>

### Threat scores—

- **Precipitation (0.25 mm or more)**: 0.35
- **(2.5 cm or more)**: 0.20

### Areas of indices or parameters related to thunderstorm occurrence

- Convective instability
- Lifted, K indices
- Wind shear

### Characteristics of features

- Minimum pressure of cyclones: ±4 mb
- Maximum speed of jet streaks
- Error in position of features (e.g., cyclone center)

### Band-passed difference fields of conventional variables (temperature, pressure, moisture, winds)

- Space-filtered
- Time-filtered

* Reliable estimates not yet available.

The $S_1$ score is sensitive to the distance between verification points, as well as the scales of motion present in the forecast and analysis. A valid comparison of $S_1$ scores, therefore, requires similar spacing of the verification points and filtering of the model forecasts and analysis in a consistent way. Twenty years of experience with the $S_1$ score of sea level pressure at the National Meteorological Center (NMC) have shown the practical range between essentially perfect and worthless forecasts to be 30–80 (Fawcett, 1977). The $S_1$ scores for 30 h forecasts of SLP at NMC have decreased steadily from ~65 in 1955 to 52 in the early 1970s (Fawcett, 1977). Recent operational model $S_1$ scores for 24 h forecasts of SLP (Table 3, Fig. 1) indicate representative values of 40–50.

Most reports of research model simulations have not included $S_1$ scores. Those that have are not directly comparable with the operational scores because they have been done for a small number of cases under research rather than operational conditions and probably have a somewhat different spacing of verification data and scales of motion in the forecast. With these limitations, there is some evidence that further improvements can be made. For example, Anthes and Keyser (1979) found an average 24-h $S_1$ score for sea level pressure in 32 cases of 39.1 which compared favorably to an average of 45.9 for the Navy's operational model and 73.4 for persistence. Koch (personal communication, 1982) indicates a significant improvement over the LFM in a number of 24 h predictions using a high-resolution research model developed by Kaplan et al. (1982).

### B. Realism of Forecasts

#### Correlation matrix

The spatial correlation between observed and forecast gridded scalar fields are calculated for various lags, or offsets, or the observed grid relative to the forecast grid. The higher the maximum correlation and the less the lag associated with the maximum, the better the forecast.

#### Structure function

The structure function is a measure of the fraction of variance associated with scales of motion smaller than a given value. It ranges in value from zero at zero distance to twice the temporal variance of a variable at large distances.

#### Spectra

- Kinetic energy
- Temperature

#### Terms in budget equations

- Kinetic energy
- Vorticity
- Water vapor
- Temperature

2) **CATEGORICAL FORECAST SCORES**

A categorical forecast is a yes or no forecast of an event, such as occurrence of a precipitation amount (usually 0.25 mm or more), at a given point during
TABLE 3. $S_1$ score for 24 h forecasts of sea-level pressure.

<table>
<thead>
<tr>
<th>Model</th>
<th>$S_1$ Score</th>
</tr>
</thead>
<tbody>
<tr>
<td>Australian Numerical Meteorology Research Center, 1978, 1979 average (Leslie et al., 1981)</td>
<td>46.5</td>
</tr>
<tr>
<td>NMC Limited-area Fine-mesh Model (LFM) (Feb-Sep 1976)</td>
<td></td>
</tr>
<tr>
<td>East of Denver</td>
<td>44.8</td>
</tr>
<tr>
<td>West of Denver</td>
<td>50.7</td>
</tr>
<tr>
<td>(LFM (computed from 49 point) 1976</td>
<td>45</td>
</tr>
<tr>
<td>latitude-longitude grid 1977</td>
<td>45</td>
</tr>
<tr>
<td>centered over U.S.</td>
<td>1978</td>
</tr>
<tr>
<td>(Newell and Deaven, 1981)</td>
<td>1979</td>
</tr>
</tbody>
</table>

* Denotes $S_1$ scores for hemispheric model.

a specified time period. A measure of success of categorical forecasts is the percentage of correct forecasts. Records show annual averages of correct forecasts of precipitation occurrence over the contiguous United States of 81–86%, with little evidence of improvement over the period 1966–1977 (Ramage, 1982).

Since 1965, the National Weather Service has issued probability forecasts of 0.25 mm or more of precipitation. The reliability of these forecasts over the entire United States has been remarkably good (Fig. 2), suggesting the value of these forecasts to those activities affected by precipitation.

3) THREAT SCORES

In contrast to the prediction of occurrence versus nonoccurrence of measurable precipitation, the threat score measures the skill in predicting the area of precipitation amounts over any given threshold (Fig. 3). The threat score $TS$ is defined by

$$TS = \frac{CFA}{(FA + OA - CFA)},$$  \hspace{1cm} (2a)

where $CFA$ is the correctly forecast area bounded by a given precipitation amount, $FA$ is the forecast area, and $OA$ is the observed area. A second form of the threat score, which is easily calculated from numerical models, is

$$TS = \frac{C}{F + R - C},$$  \hspace{1cm} (2b)

where $C$ is the number of stations (or grid points) correctly forecast to receive a threshold amount of precipitation, $F$ is the number of stations forecast, and $R$ is the number of stations observing the amount. This form of the threat score is the same as the Critical Success Index (CSI) discussed by Donaldson et al. (1975).

Subjectively prepared threat scores at NMC of precipitation in excess of 2.5 cm in the period 0–24 h have shown an annual average of $\sim$0.20 over the
United States since 1960 with no apparent trend (Committee on Atmospheric Sciences, 1980). Threat scores produced by NMC's operational regional model (the Limited Area Fine-mesh Model, or LFM) for 0.25 mm of precipitation in the 12-24 h forecast period are considerably higher (averaging \( \approx 0.40 \)) and have shown a slight increase since 1976 (Fig. 4). Fig. 4 also demonstrates an annual variation of skill, with significantly lower skill in the summer.

In addition to precipitation forecasts, threat scores can be usefully applied to other parameters of significance to small-scale weather phenomena. Examples are measures of stability that are statistically related to thunderstorms, such as areas of convective instability or threshold values of various indices (e.g., lifted index, K-index). Reap and Foster (1979) discuss results of a statistical technique based on numerical model output parameters to forecast the probability of thunderstorms.

4) \textbf{BIAS SCORE}

The \textit{bias score} measures the tendency of a model to forecast too small or too large an area of a given amount of precipitation. In terms of an area of precipitation, it is defined as

\begin{equation}
B = \frac{FA}{OA},
\end{equation}

while in terms of points (stations) it is

\begin{equation}
B = \frac{F}{R}.
\end{equation}

As shown in Fig. 4, there was little bias in the LFM model until about 1981 when a rather strong positive bias \((B \approx 1.4)\) developed. According to Hovermale (personal communication, 1982), this bias occurred when an unrealistically dry boundary-layer moisture analysis was replaced by a more realistic (and more moist) analysis. Because the convective precipitation parameterization, which had been tuned to yield little bias with the dry analysis, was not altered, an excessive number of points received the threshold amount of precipitation and a positive bias developed.

5) \textbf{PROBABILITY ELLIPSES}

If the stochastic component of mesoscale predictions is known for various phenomena, probabilistic forecasts can be developed which give users a quantitative measure of uncertainty associated with deterministic forecasts. An example is the set of probability ellipses associated with hurricane track prediction (Neumann and Hope, 1972); in addition to forecasting the most probable location of a hurricane 24, 48 and 72 h in advance, the probability that the center will fall within ellipses of different sizes is also given. Similar probability forecasts could be generated for many mesoscale phenomena over land. At present, probability forecasts for the occurrence of precipitation, for frozen precipitation and for thunderstorms are issued by the National Weather Service. The severe weather “boxes” issued by the National Severe Storms Forecast Center outlining probable areas of severe thunderstorms represent crude probability forecasts.

6) \textbf{RMS ERRORS}

In addition to the \(S_1\), threat, and bias scores, a common measure of accuracy is the \textit{rms} error. Table 4 gives two examples of \textit{rms} wind, height and temperature errors in a 24 h forecast on a 99 km mesh (Phillips, 1978). It shows typical \textit{rms} vector wind errors of 5-10 m s\(^{-1}\), height errors of 25-75 m, and temperature errors of 1.5-4.5°C.

7) \textbf{CORRELATION COEFFICIENTS}

Correlation between forecast and observed changes are useful measures of prediction skill, but they have

\begin{figure}
\centering
\includegraphics[width=\textwidth]{figure4.png}
\caption{Mean monthly threat score (top) and bias score (bottom) for 12-24 h LFM forecasts of 0.25 mm or more of precipitation (Newell and Deaven, 1981; Deaven, personal communication, 1982).}
\end{figure}
Table 4. RMS errors for two experimental 24-h forecasts on a 99-km grid of nested grid model (Phillips, 1978).

<table>
<thead>
<tr>
<th>p (mb)</th>
<th>1200 GMT 9 Jan 1975</th>
<th>0000 GMT 19 Nov 1975</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>rms Vector wind errors (m s⁻¹)</td>
<td>rms Vector wind errors (m s⁻¹)</td>
</tr>
<tr>
<td>850</td>
<td>7.6</td>
<td>5.4</td>
</tr>
<tr>
<td>500</td>
<td>6.8</td>
<td>6.8</td>
</tr>
<tr>
<td>300</td>
<td>15.5</td>
<td>10.0</td>
</tr>
<tr>
<td>200</td>
<td>13.0</td>
<td>9.5</td>
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<tr>
<td></td>
<td>rms height errors (m)</td>
<td>rms height errors (m)</td>
</tr>
<tr>
<td>850</td>
<td>41.7</td>
<td>17.8</td>
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<tr>
<td>500</td>
<td>44.3</td>
<td>24.7</td>
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<tr>
<td>300</td>
<td>66.4</td>
<td>34.8</td>
</tr>
<tr>
<td>200</td>
<td>78.6</td>
<td>43.3</td>
</tr>
<tr>
<td></td>
<td>rms temperature errors (°C)</td>
<td>rms temperature errors (°C)</td>
</tr>
<tr>
<td>850</td>
<td>4.2</td>
<td>4.5</td>
</tr>
<tr>
<td>500</td>
<td>1.8</td>
<td>1.5</td>
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<tr>
<td>300</td>
<td>3.1</td>
<td>2.6</td>
</tr>
<tr>
<td>200</td>
<td>4.6</td>
<td>1.7</td>
</tr>
</tbody>
</table>

not been reported extensively for either operational or research models. An exception is the regional model of the Japan Meteorological Agency. Nitta et al. (1979) report monthly averages of correlation coefficients for 24-h changes of 500 mb heights and surface pressure ranging from ~0.65 in summer to more than 0.8 in winter (Fig. 5).

8) Characteristics of Phenomena

Characteristics of significant phenomena, such as the minimum pressure of cyclones, maximum wind speed, or temperature and moisture gradients, can be tabulated over a number of forecasts and plotted against the corresponding observed values. Errors in the position of features, such as cyclones, are also of interest. This statistic is routinely reported for tropical cyclone forecasts. Hollingsworth et al. (1980) report a threat score measuring the skill in forecasting cyclone positions.

9) Prediction Matrix of Occurrence versus Nonoccurrence of Events

It is often important to predict the occurrence or nonoccurrence of an event such as thunderstorms, cyclogenesis, development of a mesoscale convective complex, or occurrence of an area of precipitation greater than a specified amount. Within preset limits, the exact location and amplitude are not necessarily considered. Scoring can be done over a number of cases according to a prediction matrix of number of forecasts of the event's occurrence or nonoccurrence versus the observed events (Table 2a). From this table, a threat score, accuracy, false alarm rate, bias, probability of detection, and misses can be calculated.

10) Scale Separation of Errors

Most methods of evaluating the skill in numerical models calculate errors associated with the total predicted field. As discussed by Bettge and Baumhefner (1980), separation of the total error into the errors associated with different scales in the forecast is often useful in identifying sources of model error. Bettge and Baumhefner (1980) present a method to separate various scales in a limited-area domain. The method consists of applying a digital band-pass filter to the fields. The errors associated with different scales can be computed by comparing the band-passed fields to the analyses, which are filtered in the same way. In addition to considering band-passed difference fields of conventional variables, the error variance of each band can be computed and normalized by the climatological observed variance associated with those scales.

11) Summary of Operational Skill Scores

Objective measures of the skill of operational numerical models indicate a slow but steady increase in the short-range (0–24 h) prediction of sea-level pressure, although substantial room for improvement still exists. The same is true of 500 mb heights (Shuman, 1978). An increase in the skill of predicting precipitation has been more difficult to show. The percentage of correct categorical (yes–no) forecasts of measurable precipitation have remained around 85% in the past decade, showing only very slight improvement. However, the threat score of 12–24 h precipitation forecasts by the LFM have shown slow but steady improvement since 1976. Forecasts of the probability of precipitation are highly reliable, e.g., when forecasts of 50% probability of measurable precipitation are made, precipitation occurs 50% of the time. All measures of forecast skill show greater skill in winter than in summer, which reflects the smaller-scale nature of significant weather systems in the summer.

Fig. 5. Annual variation of the monthly mean correlation coefficient between the observed and calculated 24-h changes. The solid line denotes the coefficient of the surface pressure and the dotted line that of the geopotential height at the 500 mb level (Nitta et al., 1979).
b. Statistical measures of realism of simulations

Because of the increasingly random component of mesoscale predictions as the spatial scale decreases, skill scores may indicate a poor forecast, and yet the forecast model may be quite realistic for understanding the evolution of a phenomenon and may even have practical utility. An extreme, hypothetical example illustrates this paradox. Suppose a small-scale model were developed that perfectly predicted the structure, intensity, and track of typhoons, except for small position errors resulting from errors in the speed of the storm. Fig. 6 shows hypothetical forecast and observed sea-level pressure patterns, with the predicted storm lagging behind the observed storm (assumed to be moving toward the northeast) by 100 km. The pressure pattern was computed from the empirical formula (Holland, 1980)

$$\frac{p_s - p_0}{p_e - p_0} = \exp(-A r^B) \quad 0 \leq r \leq 10 R_0, \quad (4)$$

where $p_s$ is the sea-level pressure, $p_0$ is the pressure at the center of the storm, $p_e$ is the pressure in the environment beyond $10 R_0$, $R_0$ is the radius of maximum wind speed, and $A$ and $B$ are constants. In this example, $p_0$ is 900 mb, $p_s$ is 1015 mb, $R_0$ is 40 km, $A$ is 253 km$^{1.5}$ and $B$ is 1.5.

For a two day forecast in which the observed storm is assumed to have moved at a constant speed of 30 km h$^{-1}$, the error shown in Fig. 6 represents an error in the model storm speed of 2.1 km h$^{-1}$ or a time lag of 3.6 h. Although most people would agree that the above forecast would have great utility and represents significant skill, many conventional measures of skill would indicate a worthless forecast. Fig. 7 shows the rapid increase of error associated with the $S_1$ score, rms error of pressure and vector wind speed, and threat score for increasing position errors. In the example above, the $S_1$ score of SLP is 93 and the rms pressure and wind errors are 9 mb and 12.5 m s$^{-1}$, respectively.

1) Correlation matrix

As the previous example demonstrates, a regional model might forecast the correct intensity and shape of a field but displace the field by some small distance. A correlation matrix scoring method (Tarbell et al., 1981) is a measure of skill in predicting the pattern of a scalar field such as rainfall. An observed analysis is computed on the model grid and spatial correlations between observed and predicted variables are computed for various north–south and east–west lags, or offsets of the observed grid with respect to the forecast grid. Grid points in data-void regions are not included. The result is a matrix that contains information about the skill of the model in predicting patterns. A matrix containing a few large positive correlation coefficients and a large number of small or negative correlation indicates considerable variance in the predicted and observed fields and that the model is predicting the observed structure, though not necessarily in the correct location. In the above example of the tropical storm forecast, the proper shift would yield a maximum correlation of 1.0, indicating a perfect prediction of pattern. In contrast, a smooth forecast with little structure will show a smaller maximum positive correlation and less variation for various lags. Fig. 8 compares two experimental rainfall forecasts, verifying at 1200 GMT 25 January 1978. The forecast with more structure (middle panel) has a maximum correlation coefficient of 0.87 when the analysis grid is displaced five grid points to the east and four grid points to the north (Table 5).

2) Structure function

The structure of a regional model forecast or simulation can be compared quantitatively with that of the atmosphere by the structure function (Gandin, 1963), defined as

$$b(r_1, r_2) = m(r_1, r_1) + m(r_2, r_2) - 2m(r_1, r_2). \quad (5)$$

Here the $m$ values are correlation functions for the deviations of the meteorological variables from their time mean values and the $r$ are position vectors of...
Fig. 7. Measures of forecast skill as a function of position error of hypothetical tropical cyclone model described in text. (a) $S_1$, score of sea level pressure, (b) rms error of sea level pressure, (c) rms error of vector wind, (d) threat score of rainfall of given amount as a function of $\alpha$, the ratio of position error to the diameter of the rainfall area.

the observation pairs 1 and 2. Thus, $m(r_1, r_2)$ is an autocorrelation function (the variance at station 1) while $m(r_1, r_2)$ is the covariance. We note from (5) that, as the station separation approaches zero, $b$ approaches zero (for perfect observations). If the covariance vanishes at infinite station separation, $b$ approaches twice the station variance. Thus, the ratio of the structure function to twice the mean station variance is a measure of the fraction of variance associated with scales smaller than the station spacing. When applied to observational data sets, as done by Barnes and Lilly (1975), all pairs of observations are considered and the correlations and structure functions are often grouped and displayed as a function of distance separation. Fig. 9 shows examples of the correlation and structure function for thunderstorm conditions in Oklahoma (Barnes and Lilly, 1975). A comparison of the results for temperature and mixing ratio indicates that there is much more variance associated with small scales for moisture. Thus, only 22% of the total variance of temperature occurs at scales less than 200 km, while for moisture, 68% of the variance is associated with these scales. Similar calculations on model data would indicate whether
the model simulates scales of motion similar to those in the atmosphere.

3) Spectra

Another measure of the fidelity with which a model reproduces or simulates the atmosphere is the spectrum of various quantities such as kinetic energy, temperature, water vapor, or vertical velocities.

c. Role of special data sets in verification

Some of the above measures of forecast skill and realism of a model's simulation require mesoscale observations that are not yet available operationally. For example, special observing problems occur with respect to verifying boundary-layer structure, circulations over mountainous regions and over oceans, and precipitation everywhere. Regional-scale models often produce more spatial and temporal variability than can be resolved by operational analyses. Verification of this mesoscale structure using the above methods is one important objective of field programs designed to provide special mesoscale data sets. UCAR (1983) reviews current observational capability for obtaining the needed mesoscale observations.

3. Components of regional models

Three major components of regional numerical weather prediction models can be identified. Nu-
Numerical aspects include the accuracy of the numerical approximations to the analytic differential equations. Physical aspects include the modeling or parameterization of important energy sources or sinks, such as radiation, condensation and evaporation, and frictional dissipation. Finally, a model perfect in the first two aspects would be useless for operational prediction without initial data and a method of analysis and initialization. Table 1 summarizes these aspects of some operational and research regional models. This section reviews these components and tries to identify those which are currently the most important in determining the accuracy of operational models. Rather than provide detailed reviews of the components themselves, which is unnecessary because individual reviews exist elsewhere, I concentrate on the impact of these components on regional forecasts and simulations. Because systematic tests on a large number of cases in which all components are varied have not been made, some of the conclusions are subjective and based on only a few case studies.

a. Numerical aspects

Numerical components of regional-scale models include the choice of vertical and horizontal grids, the approximations to the temporal and spatial derivatives in the partial differential equations governing the flow, and the formulation of the lateral boundary conditions. There are many variations of these components which have been studied extensively. Excellent summaries and reviews appear in the literature (Table 6), and there is no need to provide further detailed reviews here. Instead, we discuss the most significant issues involved in the choice of a particular grid structure, numerical method, and lateral boundary formulation.

1) Grid structure

Horizontal grids may have each variable (three velocity components—pressure, temperature, and moisture) defined at every grid point (nonstaggered grid), or may have individual variables defined on separate grids that are offset from each other (staggered grids). As indicated in Table 1, both nonstaggered and staggered grids have been used successfully in regional models. There is evidence, however, that staggered grids provide more accurate solutions for a given number of grid points because many of the spatial derivatives can be evaluated over horizontal distances only half as great as those associated with
Table 6. Reviews and discussions of numerical aspects of limited-area meteorological models.

<table>
<thead>
<tr>
<th>General</th>
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<tr>
<td>Eliasen (1980)</td>
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<td>Pielke (1981)</td>
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<td>WMO (1979)</td>
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<tr>
<th>Horizontal and vertical grid structures</th>
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<tr>
<td>Arakawa and Lamb (1977)</td>
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<td>Grotjahn (1977)</td>
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<table>
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<tr>
<th>Approximations to nonlinear partial differential equations</th>
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<tr>
<td>Finite-difference schemes</td>
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<tr>
<td>Haltiner and Williams (1980)</td>
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<tr>
<td>Richtmyer and Morton (1967)</td>
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<td>Kreiss and Oliger (1973)</td>
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<td>Mesinger and Arakawa (1976)</td>
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<th>Spectral/pseudo-spectral schemes</th>
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<td>Orszag (1971)</td>
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<td>Bourke et al. (1977)</td>
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<th>Finite-element methods</th>
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<tr>
<td>Strang and Fix (1973)</td>
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<td>Cullen (1976)</td>
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<tr>
<th>Lateral boundary conditions</th>
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<td>Moretti (1969)</td>
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<tr>
<td>Shapiro (1970)</td>
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<tr>
<td>Oliger and Sundström (1978)</td>
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<tr>
<td>Sundström and Elvius (1979)</td>
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Nonstaggered grids. Arakawa and Lamb (1977) discuss theoretical reasons for the increased accuracy associated with staggered grids. The superiority of a staggered grid over an unstaggered grid was shown in the three-dimensional simulation of tropical cyclones by Anthes et al. (1971) and Anthes (1972). Increased accuracy can also be obtained by staggering variables in the vertical, and most limited-area models use staggered vertical grids. The only disadvantage associated with staggered grids is their requirement for shorter time steps (a consequence of the higher horizontal resolution).

2) Numerical methods

A large number of numerical approximations to the partial differential equations exist. As discussed by Pielke (1981), the methods may be classified as finite-difference schemes, spectral/pseudo spectral methods, finite-element schemes and interpolation schemes. However, with the exception of the spectral methods, all methods produce a set of finite-difference equations, the main differences being the accuracy of the approximations, the conservation properties, and the computational cost and complexity.

Because of their simplicity and well-tested behavior under a variety of meteorological conditions, finite differences are used in the great majority of regional models. Most of these methods involve simple centered differences and are accurate to the second order (the errors in the finite differences are proportional to the second power of the grid length or time step). However, because truncation errors associated with spatial differences are known to be a significant source of model error (Miyakoda et al., 1971), several research models utilize higher-order finite-difference approximations to some of the derivatives in the complete set of equations. In a global model, Williamson (1978) found an improvement in the phase speed of shorter waves when a second-order scheme was replaced by a fourth-order scheme. Campana (1979) showed that a fourth-order version of a coarse-mesh (381 km) model gave nearly as accurate forecasts in the 0–48 h range as did a second-order, finer-mesh (190.5 km) version of the model. These results were confirmed by Leslie et al. (1981), who compared six 24 h forecasts differing in order of the advection terms and in grid size. They found an improvement in mean SLP $S_1$ score from 47.7 to 45.0 when the second-order advection terms were replaced by fourth-order approximations. This improvement, which required only a slight increase in computational time, was almost as great as that associated with halving the grid length, which required an eight-fold increase in computational time.

Considerable effort has gone into the design of finite-difference schemes that conserve (in an adiabatic, inviscid model) integral quantities such as mass, total water, total energy, and potential entrophy (Arakawa and Lamb, 1977, 1981). Except for the conservation of first-order quantities (e.g., mass and water), the general consensus is that, for regional short-range prediction models, local accuracy is more important than global conservation properties (Benwell et al., 1971). However, detailed comparisons designed to assess the value of conservative schemes have not been made with regional forecast models. It is possible that schemes which conserve potential entrophy would provide improved forecasts in regional models over high terrain, although Nakamura (1978) found truncation errors over steep mountains to be considerably smaller than expected.

The temporal integration schemes may be classified as explicit, in which the temporal rate of change of all variables depends only on present or past values of the variable; or implicit, in which the tendencies depend on future variables as well. Solution of the implicit schemes is possible through relaxation or matrix inversion techniques. While the explicit schemes are simpler to implement, the implicit schemes are more efficient because they may be stably integrated with substantially longer (2–8 times longer) time steps. Robert et al. (1972), Gauntlett et al. (1976), and McGregor et al. (1978) have successfully used implicit schemes in baroclinic models.

Compared to the forecast errors introduced by spatial truncation errors, the error caused by temporal truncation errors is probably insignificant. This is
because meteorologically significant features have important variations on scales as small as two to four times the horizontal grid length where truncation errors are large. However, these features have periods greater than an hour or more, which is many times that of the time steps used in regional models (Table 1). While temporal truncation errors adversely affect high-frequency gravity waves (Janjić and Wiin-Nielsen, 1977; Collins, 1980), the overall effect on the forecast is probably minimal, as evidenced by the success of implicit models which use time steps several times longer than similar explicit models (Seaman and Anthes, 1981).

Most regional models require a method to prevent energy from accumulating in the shortest resolvable wavelengths and high temporal frequencies. While some finite-difference schemes minimize this problem by their inherent damping properties (such as the Euler-backward time integration scheme which heavily damps high-frequency waves), most models utilize explicit methods to control these unwanted features. For controlling high temporal frequencies, a widely used method is a time filter (Asselin, 1972). To control the short wavelengths, horizontal diffusion terms are used most often. These terms may be proportional to $\nabla \cdot K \nabla$ or $\nabla^2 K \nabla^2$ of the quantity where $K$ is an eddy diffusivity which may be constant or depend on some property of the flow, such as deformation. The latter has the desirable property of being considerably more scale-selective, and provides more damping of the shortest waves and less damping of the intermediate waves (Williamson, 1978).

3) Lateral boundary conditions

Lateral boundary conditions are undoubtedly a major source of error in regional models. Conditions are required which allow internally generated waves of all frequencies to pass out of the domain, while allowing meteorologically significant information to propagate into the domain. While no completely acceptable method has been demonstrated (because of the ill-posed mathematical nature of the problem), a number of reasonably successful schemes have been tested and used in regional models. One popular method, often called the sponge method, involves utilizing an increased horizontal eddy viscosity in a band around the lateral boundaries (Perkey and Kreitzberg, 1976). This zone of high viscosity damps waves propagating out of the interior before they can reflect off the boundary. A diffusion term may be added to the prognostic equations in this zone to allow large-scale information obtained from a synoptic-scale model to propagate into the domain. The movable fine-mesh model of NMC (Hovermale and Livzezy, 1977) follows this approach. In two experiments with a coarse-mesh limited-area model, Baumhefner and Perkey (1982) found that the errors associated with the Perkey-Kreitzberg boundary conditions were considerably less ($\sim 20\%$) than the total forecast error in the 0–48 h time period.

A second method involves the use of approximate radiative boundary conditions which minimize the reflection of waves (Orlanski, 1976) even without a sponge zone. Recent results of Orlanski with a limited-area model indicate that a version of radiative boundary conditions, in which vorticity and divergence are extrapolated outward at the boundary rather than the velocity components themselves, is successful in minimizing distortion at the boundaries. Miyakoda and Rosati (1977) have found the radiation boundary conditions to be superior to the sponge technique.

A third, and perhaps the most satisfactory, solution to the lateral boundary problem in limited-area models is to nest the regional model in a larger-scale model. Considerable success with two-way interacting nested grids has been demonstrated by Jones (1977) and Phillips (1978). Discussion and reviews of nested grid techniques are provided by Phillips and Shukla (1973) and Elsberry (1978). An example of the effect of lateral boundary conditions on 24 h forecasts of a regional model and the validity of a one-way interacting nested-grid concept is provided by Leslie et al. (1981). Fig. 10 shows the average 24-h $S_1$ scores from six experimental forecasts. In the nonnested forecasts, the variables at inflow points are assumed to be in steady state. In the nested grid forecasts, time-dependent values are obtained from a hemispheric model. The use of time-dependent boundary conditions results in a significantly reduced $S_1$ score at all levels.

In summary, most regional-scale numerical models utilize finite-difference schemes with second-order

![Fig. 10. Mean 24 h $S_1$ scores as a function of height for nested grid and nonnested forecasts (Leslie et al., 1981).](image-url)
accuracy. It is probable that the largest errors associated with the numerical aspects of the model arise from the lateral boundary conditions and spatial truncation errors. These can be minimized at a computationally reasonable cost by the adoption of staggered horizontal grids, higher-order difference schemes and nested grids. Implicit temporal integration schemes can reduce the computation cost by a factor of two or more with little reduction in accuracy, because temporal truncation errors do not appear to be an important part of the total forecast error.

4) DATA ANALYSIS AND INITIALIZATION

Two important components of regional models are the analysis and initialization steps. In the analysis phase, grid point values of the model’s variables are estimated from diverse types of surface and upper-air data at irregularly spaced locations. A widely used analysis scheme in regional models is the method of successive corrections, in which a first-guess field is modified by a series of scans which use observations to modify the first-guess errors. Cressman (1959) introduced this method; others have suggested refinements and variations (Barnes, 1964).

Optimum interpolation (O/I) is an analysis method, introduced by Gandin (1963), designed to select a set of weighting coefficients which minimize the rms error of the analysis over a large ensemble of synoptic situations. The weights depend on the spatial covariances among the difference (first-guess minus observation) fields. Various O/I schemes have been tested by Bleck (1975), Schlatter (1975), and Bergman (1979).

Otto–Bliesner et al. (1977) compared four analysis schemes (a Cressman analysis scheme, a global multivariate O/I scheme, an isentropic O/I scheme, and a subjective analysis). Although differences were relatively minor, there was some evidence that the O/I schemes did not resolve the amplitude of some of the features. McPherson et al. (1979) found that an O/I method resulted in a better resolution of the wind field in one case study than did a global spectral analysis scheme. In addition, the O/I scheme produced a considerably superior analysis of humidity. Reimer (1980) used a univariate O/I method on isentropic coordinates to test the impact of grid resolution on the analysis. Over a data-rich region (central Europe), he found a significant improvement by utilizing a fine grid for the analysis (Fig. 11).

The above comparisons of analysis methods consider only a limited number of cases and did not test the impact of the different analyses on a subsequent forecast. The performance of the ANMRC model, however, indicates that an improved analysis scheme can make a significant positive impact on the forecast. A decrease in SLP S1 score at 24 h of about two points was noted when a variational analysis scheme (Seaman et al., 1977) was introduced.

Because of the limited number of comparisons, it is difficult to reach firm conclusions regarding the superiority of one type of analysis scheme compared to another. However, there is some evidence that local analysis methods are better than spectral methods. There is less evidence of the forecast benefits to be gained by utilizing the computationally more expensive O/I methods over the simpler Cressman-type schemes. There is some evidence that analysis on isentropic surfaces is preferable to analysis on other surfaces, because the horizontal scale of variation of the wind is larger on θ surfaces, especially in frontal zones.

Most objective analyses result in mass and wind fields that are unbalanced, i.e., the temporal tendencies are much greater (an order of magnitude or more) than the correct values. The result is the generation of inertia–gravity waves. Although in most cases these waves do not interact greatly with the meteorologically significant waves, they are a nuisance in that they contaminate output fields of interest, such as sea-level pressure and vertical velocity. It is therefore desirable to adjust the variables to obtain a balanced state at the start of the forecast, a process known as initialization (Kasahara, 1982). Leith (1980) has discussed the balanced state in terms of the concept of a slow manifold. When on the slow manifold, the divergence and vorticity are consistent with a slowly evolving, meteorologically significant forecast. The initialization phase of a model forecast is designed to modify the objective analysis to obtain initial values that correspond to a solution on the slow manifold.

Two methods have shown great promise for obtaining balanced initial conditions. When periodic domains are considered (global or hemispheric), the initial analysis can be projected on the model’s eigenfunctions or normal modes, and the unwanted, fast modes eliminated by setting their amplitude to zero (Baer and Tribbia, 1977). Phillips (1981) shows how variational analyses combined with a nonlinear normal mode initialization can improve the initial analysis in a more significant way than the elimination of noise. Reviews of this normal mode method are provided by Daley (1981) and Kasahara (1982). In spite of the fact that the scheme is not as well-suited to limited-area models which involve nonperiodic domains, progress is being made (Briere, 1982). In addition, the method may be applied in regional models by initializing on a global domain and interpolating the global fields to the fine-mesh, limited-area model.

A related scheme, which is more suitable for direct initialization of regional models, is the bounded-derivative method (Browning et al., 1980; Kasahara, 1982). In this method, the analyzed data are adjusted
to reduce the temporal derivatives to meteorologically realistic values. Although the bounded-derivative scheme has potential for future models, it has not yet been tested in regional models.

Other methods of initialization have involved using various combinations of the divergence equation and "omega" equation to obtain winds from the mass field or vice versa. For example, in several models (Table 1), the divergence equation with the terms involving the divergence set to zero (balance equation) is used to obtain the nondivergent wind component $V_{ND}$ from the mass field, while the $\omega$-equation is used to estimate the divergent wind component. In order to minimize the amplitude of the external gravity wave (Benwell and Bretherton, 1968), the vertically integrated mass divergence is often eliminated (Washington and Baumhefner, 1975). Although these techniques have been partially successful in reducing noise in models (Fig. 12), their impact on the slower modes has been more difficult to show, because of

Fig. 11. Vertical cross sections of relative humidity generated by objective analysis for 1200 GMT 2 March 1977, where B, C, and D denote 381, 190.5 and 95.75 km grid resolutions, respectively. Analysis A was produced subjectively. The horizontal extent of the cross-sections are $\sim 1900$ km (Reimer, 1980).
the weak interaction between the slow and fast modes.

A difficulty with the initialization schemes mentioned so far is that the strong, small-scale vertical motions associated with latent heating in regions of heavy rainfall are not resolved. A method that utilizes observed rainfall rates to obtain vertical motions (and the associated divergent wind component) has been proposed by Tarbell et al. (1981). An improvement in the very short-range (0–6 h) forecast of precipitation was shown with this scheme, but the impact after 12 h was small.

Analysis and initialization schemes, of course, depend on adequate observations, and there is evidence that limitations in data coverage, particularly in the Southern Hemisphere, contribute significantly to errors in regional model forecasts. Leslie et al. (1981) found a decrease in the annual average SLP $S_1$ score from 48 in 1978 to 45 in 1979, when an enhanced data base associated with FGGE was available. The extra data came from ocean buoys and the TIROS-N satellite. It is not clear to what extent higher-resolution data would improve regional model forecasts over the United States. However, it is likely that some improvement could be obtained with more dense coverage, especially in precipitation forecasts, since considerable structure in the meteorological variables, especially moisture, exists in scales of motion unresolved by the current radiosonde network. Prospects for improved mesoscale initial data from remote-sensing (both surface-based and from satellites) are discussed in detail by UCAR (1983).

b. Physical aspects

The physical components of a numerical model include the modeling of energy sources and sinks associated with fluxes of heat, moisture, and momentum at the earth’s surface, in the planetary boundary layer (PBL) and occasionally the free atmosphere. They also include radiation and changes of phase of water in clouds of all types and scales. Many of these physical processes are associated with scales of motion smaller than those resolved by the model. These processes are referred to as “subgrid-scale” processes and relating their cumulative effect on the flow to the scales of motion resolvable by the model is known as parameterization. The concept of parameterization (and subgrid-scale effects) for global models is discussed in GARP Publication Series No. 8 (WMO, 1972). According to this report, successful parameterization requires a number of steps.

i) Identification of the process.

ii) Determination of its importance for the resolvable scales of motion.

iii) Intensive studies of individual cases in order to establish the fact that the relevant physics and dynamics are adequately understood.

iv) Formulation of quantitative rules for expressing the location, frequency of occurrence, and intensity of the subgrid-scale processes in terms of the resolvable scale.

v) Formulation of quantitative rules for determining the grid-scale averages of the transports of mass, momentum, heat, and moisture and verification of these rules by direct observations.

The representation of physical processes in regional models and the impact of these processes on short-range, regional forecasts and simulations are reviewed in this section.

1) SURFACE PROCESSES

The properties of the earth’s surface, including roughness, temperature, albedo, and moisture availability (percent of saturation) directly affect the vertical fluxes of momentum, heat, and water vapor into (out of) the lowest layer of the model and hence represent sources and sinks of energy. Although these processes have been studied extensively through theoretical and observational programs by the boundary-layer research community, they are only recently beginning to be studied in the context of their relationship to tropospheric flows in regional models.

Treatment of fluxes of heat and moisture in regional models varies from their neglect to rather sophisticated parameterizations based on a surface energy budget. Because the sea surface temperature is relatively constant over the forecast time periods, calculation of fluxes over water is relatively straightforward, and most models compute these fluxes over water utilizing bulk aerodynamic formulas.

Several regional models utilize a surface energy budget to obtain the ground temperature and moist-
ture availability (degree of saturation of the ground). With these parameters and an estimate of air temperature and water vapor, heat and moisture fluxes may be computed using similarity theory (Blackadar, 1979; Pielke, 1981). Blackadar developed an economical, accurate method for predicting the ground temperature and surface heat fluxes (Deardorff, 1978). This model, when coupled to a boundary-layer model, depends on the net radiation (a strong function of cloud cover), albedo, moisture availability, and, to a lesser extent, the thermal capacity of the soil. The model is simple enough for incorporation in operational regional models.

2) SURFACE-LAYER AND PLANETARY BOUNDARY-LAYER PROCESSES

There are essentially two methods of parameterizing the surface layer (0–100 m) and the planetary boundary layer (PBL) in regional models. Most models (Table 1) use the well-known bulk aerodynamic method which treats the surface layer and PBL as a single layer and models the surface fluxes of heat, moisture, and momentum by exchange coefficients. The depth of the PBL may be fixed or vary in time (e.g., Deardorff, 1972), while the exchange coefficients may be constant or vary with roughness or static stability. The bulk-aerodynamic method is simple, computationally efficient, and has been reasonably successful in regional models.

In recent years, a number of high-resolution boundary-layer models have been developed and tested within a one-dimensional framework (Blackadar, 1979; Pielke, 1981). These models, though requiring more computer time (an additional five layers or so) provide for more generality than the bulk PBL models, for example, during the transition from well-mixed conditions to stratified nocturnal conditions in which strong vertical gradients of temperature, wind and moisture often exist. Blackadar presents additional arguments for the need for high-resolution PBL models. Considerable testing of various high-resolution PBL models in a one-dimensional framework is discussed in the literature (e.g., Burk, 1977; Chang, 1979; Yamada and Mellor, 1979).

Extensive studies of the role of surface fluxes and PBL processes in regional models have not been carried out, but a few research modeling studies indicate the importance of the PBL on regional-scale flows over periods as short as 0–24 h. For example, sea level pressure forecasts have depended rather strongly on surface friction. Graystone (1962), Bushby (1968), Danard (1969a), and Anthes and Keyser (1979) found differences of 5–20 mb in the minimum pressures of cyclones in 24 h forecasts with and without friction.

A simplified argument, following Danard (1969a), yields some insight into the effect of surface friction on the surface pressure tendency. The equation of motion, neglecting horizontal diffusion, can be written

\[ \mathbf{V} = \mathbf{V}_s + \frac{1}{f} \mathbf{k} \times \left( \frac{d\mathbf{V}}{dt} - \frac{1}{\rho} \frac{\partial}{\partial z} \sigma \right) \]

where \( f \) is the Coriolis parameter, \( \mathbf{V} \) is velocity, \( \mathbf{V}_s \) is geostrophic velocity, \( \rho \) is density, and \( \sigma \) is stress. Substitution of (6) into the pressure tendency equation

\[ \frac{\partial p_0}{\partial t} = -g \int_{0}^{\infty} \nabla_h \cdot \rho V dz \]

yields

\[ \frac{\partial p_0}{\partial t} = -g \int_{0}^{\infty} \left( \nabla_h \cdot \rho V_s + \nabla_h \cdot \rho f^{-1} \mathbf{k} \times \frac{d\mathbf{V}}{dt} \right) dz \]

\[ + g f^{-1} \mathbf{k} \cdot \nabla_h \times \tau_0. \]  

The first two terms in (8) represent dynamical and thermodynamical processes (density advection and changes of vorticity) throughout the depth of the atmosphere. The third term represents the instantaneous effect of surface friction. Surface friction plays a more complicated role than might be indicated at first by (8), because vertical motions induced by friction modify the temperature and pressure aloft and therefore change the first two terms. However, it is instructive to consider the order of magnitude of the friction term on the instantaneous pressure tendency. We express the surface stress \( \tau_0 \) by the usual quadratic law

\[ \tau_0 = \rho_0 C_D V^2, \]

where \( V \) is the wind speed at some level and \( C_D \) is the drag coefficient appropriate for the wind at that level. If we linearize (9) by taking \( V \) as a representative constant value and substitute into the third term of (8), we have

\[ \frac{\partial p_0}{\partial t} \approx g \rho_0 f^{-1} C_D V^2. \]  

For \( g = 9.8 \text{ m s}^{-2}, \rho_0 = 1.02 \text{ kg m}^{-3}, \) and \( f = 10^{-4} \text{ s}^{-1}, \) (10) becomes

\[ \frac{\partial p_0}{\partial t} \approx 100 C_D V \xi(kPa \text{ s}^{-1}), \]

where \( V \) and \( \xi \) are expressed in m s\(^{-1}\) and s\(^{-1}\), respectively. As shown by (10), the instantaneous effect of surface friction in the pressure tendency varies linearly with \( C_D \) and with the square of the wind speed, since vorticity is approximately proportional to wind speed. Table 7 gives pressure tendencies for various values of \( \xi \) and \( V \) for \( C_D = 2 \times 10^{-3} \). From this table, the effect of friction in a 24 h forecast varies from negligible for light winds to extremely important (24 h pressure changes greater than 190 mb). Of course, the tendencies given in Table 7 could not be realized
TABLE 7. Instantaneous pressure change due to surface friction alone (mb h⁻¹) for $C_p = 2 \times 10^{-3}$.

<table>
<thead>
<tr>
<th>$\zeta \times 10^{-4}$</th>
<th>1</th>
<th>5</th>
<th>10</th>
<th>20</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.01</td>
<td>0.007</td>
<td>0.036</td>
<td>0.072</td>
<td>0.144</td>
</tr>
<tr>
<td>0.05</td>
<td>0.036</td>
<td>0.180</td>
<td>0.360</td>
<td>0.720</td>
</tr>
<tr>
<td>0.10</td>
<td>0.072</td>
<td>0.360</td>
<td>0.720</td>
<td>1.44</td>
</tr>
<tr>
<td>0.50</td>
<td>0.360</td>
<td>1.800</td>
<td>3.60</td>
<td>7.20</td>
</tr>
<tr>
<td>1.00</td>
<td>0.720</td>
<td>3.60</td>
<td>7.20</td>
<td>14.40</td>
</tr>
</tbody>
</table>

as 24 h pressure changes because of the feedback into the dynamical processes aloft.

The above discussion indicates the importance of boundary-layer friction on cyclone intensity. Other studies, using more sophisticated PBL models, have indicated the importance of surface fluxes on the temperature and moisture structure of the PBL in addition to the pressure. Yamagishi (1980) utilized a medium-resolution PBL model based on similarity theory and the level 2 (Mellor and Yamada, 1974) turbulent closure model in a regional forecast of a cold air outbreak over the Sea of Japan. His model simulated the observed heat and moisture fluxes and the height of the mixed layer reasonably well when accurate sea surface temperatures were specified.

Although no systematic comparisons of the effect of different PBL parameterizations on regional model forecasts have been made, Miyakoda and Sirutis (1977) studied the response of a general circulation model to three different parameterizations. The schemes tested over a 30 day integration were 1) Mellor and Yamada's (1974) level 2.5 closure model, 2) a dry convective adjustment model (Manabe et al., 1965), and 3) a mixed-layer model developed by Randall and Arakawa. Both the level 2.5 closure model and the mixed-layer model produced more realistic simulations than did the forecast with convective adjustment, which showed excessive cooling in the lowest 400 m.

The effects of soil moisture on a model with 2° latitude-longitude resolution over West Africa was investigated by Walker and Rowntree (1977) in idealized experiments. When a dry surface typical of the Sahara was replaced by moist land, the rainfall associated with transient disturbances increased significantly. Not only was precipitation affected in the first two days of the simulation, a strong positive feedback between increased precipitation and increased soil moisture was noted out to 20 days, which suggests a mechanism for the persistence of precipitation anomalies.

Leslie (1980) used a scheme based on a surface energy balance to study the generation of the summer heat lows over Australia in more than a month of 24 h forecasts. In a comparison to forecasts without surface heating, the mean error in minimum pressure associated with the heat low was reduced from 2.5 to 0.9 mb, and the forecasts were judged superior by independent observers in 35 of 37 cases.

Using a model with 100 km horizontal resolution, a prognostic surface energy budget, and a medium-resolution boundary-layer model, Benjamin (personal communication, 1982) found that surface heat, moisture, and momentum fluxes had a large impact on forecasts as short as 12 h over the central United States under severe weather conditions. Using data from the SESAME-I (10–11 April 1979) and SESAME-IV (9–10 May 1979) experiments, he found significant differences in boundary-layer temperature, specific humidity, and surface pressure between 12 h forecasts with and without surface fluxes (Fig. 13). These differences were caused by differential heating associated with topographic variations, cloud cover, and variable surface characteristics (notably moisture availability), differential evaporation, and changes in the horizontal transport of mass, heat, and water vapor associated with the changing pressure field.

In summary, surface fluxes of heat, moisture, and momentum have been shown to be important in 0–48 h forecasts on the regional scale in a number of case studies. There is also evidence that horizontal variations in surface parameters and cloud cover can produce significant horizontal variations in PBL structure on these time and space scales in numerical models. While it has not yet been shown that the use of a surface energy budget over land coupled with a

FIG. 13. Difference fields in 12 h forecasts with and without surface fluxes of heat and water vapor. Forecasts were initialized at 1200 GMT 10 April 1979. Thick solid lines are surface pressure differences in mb. Dashed lines are temperature differences at lowest level in model in K. Thin solid lines are differences in specific humidity at lowest level in model in g kg⁻¹ (S. G. Benjamin, personal communication, 1982).
medium-resolution boundary-layer model will significantly improve regional-scale forecasts, there is enough evidence to warrant testing of a model with these properties on a large number of cases.

3) CONDENSATION AND EVAPORATION PROCESSES

The release of latent heat of condensation represents an important source of energy for synoptic-scale cyclones (Aubert, 1957; Danard, 1964; Tracton, 1973) and is also important in modifying the largerscale environment. Observational studies (Ninomiya, 1971; Maddox et al., 1981; Fritsch and Maddox, 1981) have shown the development of anticyclonic perturbation flows in the upper troposphere over mesoscale regions of precipitation. The increasing baroclinicity often induces an upper-level jet streak north and west of the convective system.

Numerical models have successfully simulated the development of mesoscale convective systems and the effect of latent heat on the environmental flow. Chang et al. (1982) isolated the effect of latent heating on a 24 h forecast by subtracting a forecast without latent heating from a control forecast which contained heating. Fig. 14 presents the 24 h forecast height, temperature, and vector wind differences (wet-dry) on the 700, 500 and 300 mb pressure surfaces. As expected, the maximum differences occur over the regions of intense precipitation. The major patterns are a cold low at 700 mb, a warm low at 500 mb, and a warm high at 300 mb. Throughout the troposphere there is a net increase in the area-mean (calculated over the region depicted in the figures) 24 h predicted temperature due to the latent heat release. This temperature increase is largest at 300 mb (0.85°C), decreasing in magnitude downward to 700 mb (0.03°C). The height difference fields indicate an average increase in the 700–500 mb thickness of 5 m and an increase in 500–300 mb thickness of 13 m. More significant to the dynamics, however, are the large horizontal gradients in temperatures and heights, which produce changes in the winds and vertical motions. Comparing the vertical velocity fields (not shown) between the wet and dry simulations, the wet case produces much larger upward motion in the area of positive temperature difference at 500 mb. This indicates that the latent heating exceeds the adiabatic cooling and, thus, the air column is forced to expand. Consequently, relatively high and low pressure areas are generated in the upper and lower troposphere, respectively.

At 300 mb, the generation of high pressure to the north fills the northern portion of the large-scale trough, and the generation of low pressure to the south deepens the southern portion of the trough. As a result, a cut-off low forms in the wet run. The corresponding change in wind circulation enhanced the north-south temperature advection. These processes, in conjunction with latent heating, appear to be responsible for the distribution of positive temperature difference over a rather wide region. At 500 mb, the temperature difference is confined to the areas of large precipitation, indicating that latent heating and adiabatic cooling are the main factors in determining its pattern, i.e., the effects of the latent heat are not rapidly advected out of the region. At both 500 and 300 mb, the increase in temperature due to the inclusion of condensation processes occurs largely in the colder air. Thus the upper-level baroclinicity is reduced. At 700 mb, the change in temperature is related more to cold advection than to diabatic processes. In response to the enhancement of the pressure gradient, the winds increase their cyclonic circulation at 700 and 500 mb and their anticyclonic circulation at 300 mb. As evidenced by the increased vertical motion, the wet case’s low-level winds have a stronger cross-isobaric component than do those of the dry case. The maximum wind difference is 21 m s\(^{-1}\) at 700 mb, 14 m s\(^{-1}\) at 500 mb, and 35 m s\(^{-1}\) at 300 mb. At all three levels, this maximum occurs over the northwest quadrant relative to the precipitation field. This preference is similar to that shown by Maddox et al. (1981) for mesoscale convective complexes (MCCs).

The strong effects of latent heat shown in Fig. 14 have been found in other models with rather different parameterizations of convective heating. Anthes et al. (1982a) showed that latent heating induced a divergent anticyclonic wind perturbation near 300 mb with a jet streak of more than 15 m s\(^{-1}\). In the boundary layer the latent heat generated a pressure decrease of more than 7 mb and a cyclonic circulation with perturbation wind speeds greater than 10 m s\(^{-1}\). Similar results were obtained by Maddox et al. (1981) and Ninomiya and Tatsumi (1981).

It is well-known from theoretical and numerical studies that the vertical distribution of convective heating is extremely important in determining the evolution and structure of tropical cyclones (Anthes, 1982). As regional models are increasingly applied to study middle-latitude phenomena, it is becoming clear that the vertical distribution of latent heating is also a crucial factor in the development of real and model extratropical cyclones (Tracton, 1973; Anthes and Keyser, 1979).

Primitive equation models have demonstrated the sensitivity of meso-\(\alpha\) scale extratropical circulations to rather small changes in the specified vertical distribution of heating whenever substantial (greater than \(\sim 2\) cm of rain per 12 h) precipitation is predicted. Anthes and Keyser (1979) showed an example of a 12 h forecast in which lowering the maximum in the specified vertical distribution of convective heating from 480 to 600 mb produced a cyclone with minimum pressure 11 mb lower. The greater proportion of heat released in the lower troposphere de-
stabilizes the atmosphere and permits a much more rapid intensification. This interpretation is consistent with Sutcliffe's (1947) development theory, in which the greatest brake on a developing cyclone is the adiabatic cooling associated with upward motion (Petterssen, 1956, p. 329). It is also consistent with
Staley and Gall's (1977) study which showed that the wavelength of maximum growth in a baroclinically unstable environment shifts toward smaller wavelengths (∼2000 km) as the lower troposphere becomes less stable.

Observational studies have demonstrated the importance of evaporation on the meso-γ and meso-β structure of precipitating systems. Zipser (1969) showed that evaporation was an important (indeed, essential) process in the maintenance of a tropical precipitating system. Moncrieff and Miller (1976), Ogura and Liou (1980), and Houze and Betts (1981) discuss the importance of evaporation in producing the downdraft structure of thunderstorms and squall lines. Brown's (1979) numerical modeling study showed that evaporation was the primary mechanism in driving a meso-β scale downdraft. While considerable effort has been devoted toward understanding the role of evaporation on small mesoscale features, relatively little attention has been paid to the effects of evaporation on meso-α scale features.

Although it has been demonstrated conclusively that latent heat release and its vertical distribution can influence significantly the evolution of regional forecasts, the problem of accurately parameterizing the effects of cumulus convection is still unsolved. As discussed by Anthes et al. (1982b), cumulus convection affects the environment through diabatic heating and cooling associated with condensation, evaporation, freezing and melting, through vertical fluxes of sensible heat, moisture and momentum, and through horizontal pressure perturbations. The processes are highly nonlinear and depend in complex ways upon the size spectrum of clouds and on the environmental flow.

Some success with cumulus parameterization schemes of varying complexity has been reported by Kuo (1965, 1974), Arakawa and Schubert (1974), Kreitzberg and Perkey (1976), Anthes (1977), Hayes (1977), Sundqvist (1978), and Fritsch and Maddox (1981b). The Arakawa and Schubert, and the Kuo type parameterizations have been verified by diagnostic studies in the "semi-prognostic" sense, as discussed by Lord (1980) and Krishnamurti et al. (1980). Fig. 15 shows the observed and computed rainfall rates from a cumulus parameterization based on Kuo's (1974) scheme. Although not a complete test in the sense that feedbacks between the convective heating and dynamical fields are not allowed, such studies provide encouragement that realistic cumulus parameterization in regional-scale models is possible.

Perhaps the simplest parameterization of cumulus convection is the convective adjustment scheme (e.g., Manabe et al., 1965), which consists of the adjustment of model lapse rates that exceed the wet adiabatic value under conditions of convective instability. Convective adjustment schemes tend to under-forecast convective rainfall rates while simultaneously overpredicting the area of convective rainfall (Hayes, 1977). After introducing a scheme in which the parameterization of deep cumulus convection was controlled by large-scale convergence of water vapor and conditional instability (a Kuo type parameterization), the operational UK regional model showed increases in convective rainfall over England in summer by more than 40%, and "significant improvements" in the "forecast patterns and amounts of rainfall" (Hayes, 1977).

Ninomiya and Tatsumi (1980), in a simulation of Baiu frontal rainfall, found that a moist convective adjustment scheme which worked well in a 381 km mesh forecast gave unrealistic results with a 77 km mesh.

Parameterizing cumulus convective effects as a function of the resolvable scale becomes questionable as the grid size becomes smaller than ∼100 km. For finer meshes, the separation between the resolvable and convective scales becomes less and the model may begin to simulate the same clouds it is also trying to parameterize. For high-resolution models, therefore, it may be preferable to abandon the concept of parameterization in favor of explicit treatments of condensation and evaporation through the introduction of prediction equations for cloud and precipitation water. This strategy has shown some success in studies by Rosenthal (1978) and Ross and Orlanski (1982).

4) LAYERED CLOUDS AND RADIATIVE EFFECTS

Clouds greatly modify the shortwave and longwave radiation budget and thereby exert an important control on the evolution of the planetary boundary layer. The energy budget at the earth's surface is responsible for generating important mesoscale circulations such
as sea breezes, mountain–valley breezes, and in producing conditions of low-level stability that are favorable for moist convective systems. Thus even nonprecipitating clouds are probably important in perturbing regional-scale flows.

Clouds can have an important effect on the atmosphere over short time periods through differential heating between clear and cloudy regions and through destabilization of the atmosphere by strong cloud-top cooling (Danard, 1969b; Cox, 1969; Gray and Jacobson, 1977). In spite of their importance, most regional models either neglect nonprecipitating clouds entirely or parameterize their radiative effects in a crude manner based on the mean layer relative humidity. Layered, nonprecipitating clouds are particularly difficult to parameterize in numerical models because they are often much thinner than the vertical resolution of the models. An example of the difficulty in relating middle tropospheric cloud amounts to mean layer relative humidity is shown in Fig. 16.

Slingo (1980) used statistics from GATE data to develop a parameterization for the percentage of low, medium, high and convective clouds. Because observations showed that layer clouds were concentrated over disturbances and decreased rapidly away from the disturbance, quadratic relationships between relative humidity \( R \) and cloud amount were developed:

\[
C_H = (R - 80)^2/400, \quad R \geq 80
\]

\[
C_M = (R - 65)^2/1225, \quad R \geq 65
\]

\[
C_L = (R - 80)^2/400, \quad R \geq 80
\]

where \( C_H \), \( C_M \) and \( C_L \) are the percentage of high, medium, and low clouds, respectively. This scheme was tested in a large-scale model (2° latitude–longitude mesh) and was found to predict the cloud distribution reasonably well, including the difficult-to-predict stratocumulus clouds. Similar, although not necessarily identical, parameterizations may be useful in models of extratropical systems.

An alternative approach to the statistical methods of estimating cloud cover is to predict the clouds explicitly. Sundqvist (1978) proposes a scheme involving a prognostic equation for cloud liquid water content. The scheme, tested in a one-dimensional model, gave realistic vertical distribution of cloud water, precipitation, and evaporation.

Probably the most important radiative effect on short-range regional model forecasts is the absorption of solar radiation at the surface, which is the main driving mechanism for surface sensible and latent heat fluxes. In the free atmosphere, changes of temperature due to the absorption of shortwave radiation during the day and emission of longwave radiation at all times are approximately 1–2°C per day when averaged over layers of the thickness of most numerical models. Because these diabatic heating rates are small compared to other diabatic and adiabatic rates of temperature changes, most regional models neglect radiative effects above the surface.

While radiation parameterizations are uncommon in present regional models, considerable effort has been devoted to modeling radiation in general circulation models. These methods, which range in complexity from direct numerical integration of the radiation transfer equations to the use of climatological means, are reviewed in WMO (1972).

Although layered-mean diabatic heating rates associated with radiation are small, rates over thin layers in the vicinity of cloud tops can be large and affect the development of clouds. Using a five-band model to calculate the infrared heating rate above layered clouds, Roach and Slingo (1979) found cooling rates of more than 8°C h\(^{-1}\) in a thin (1 mb) layer above the cloud.

Interaction of radiation and clouds may be significant in modulating cloud growth and precipitation, and may be responsible for the observed diurnal variations in tropical precipitation (Gray and Jacobson, 1977). Possible mechanisms are the destabilization of cloud tops during the night and enhancement of

![Fig. 16. Scatter diagram of observed middle cloud amount against mean relative humidity for the layer 700–500 mb (Slingo, 1980).](image-url)
mesoscale circulations through differential radiative heating between cloudy and clear regions (Anthes, 1982).

Compared to the efforts devoted to parameterizing surface fluxes, planetary boundary layer processes, and cumulus convection, and to evaluating the effects of parameterizations in regional models, there have been only limited studies on radiative effects. In one of the few studies, Kubota (1981) considered the radiative effects on an eight-day Northern Hemisphere forecast. The major effect of clouds was to reduce the average cooling in the lower troposphere and increase the cooling in the upper troposphere, which represents an increased greenhouse effect. Offsetting the greenhouse effect slightly were an increase in reflected solar radiation (albedo effect) and a reduction of surface heating (ground effect) in the experiment with clouds.

Hollingsworth et al. (1980) compared two 10-day forecasts with different radiation parameterizations (as well as different parameterizations of other physical processes). In the simpler scheme, the radiative absorbers (H₂O, CO₂ and O₃) and the clouds were specified as temporally constant functions of latitude and height. In the second scheme, the moisture and clouds varied, allowing for a cloud-radiative feedback. CO₂ was constant in space and time, and O₃ was a specified function of all three spatial coordinates. Although differences in mean heating rates over the 10-day period were several degrees Celsius per day, there were only minor differences in the forecasts.

4. Summary and conclusions

This paper has reviewed the current status of operational and research numerical weather prediction over limited-area, regional domains. Over the past decade there has been intensive research activity in several operational centers and research institutions around the world. While increases in the accuracy or skill of operational short-range forecasts have occurred over this period of time, there remains much room for improvement, especially in the prediction of significant weather such as quantitative precipitation. These improvements, if they occur, will probably not result from a major breakthrough in any one aspect of the regional models, because there are so many numerical and physical factors that are important in producing an accurate prediction or a realistic simulation. These include the accuracy and density of the initial data, the analysis and initialization methods, the numerical treatment of the partial differential equations, and the modeling or parameterization of physical effects, including surface fluxes, boundary layer processes, release of latent heat, moist convective transports, clouds, and radiation.

A number of research models investigating the above effects have demonstrated the importance of each on a limited number of cases. What is needed now is a systematic testing of regional models which include the potential improvements proposed in the research models. These tests, and future research model experiments, would be most effective if a comprehensive set of verification statistics were computed. In addition to the conventional measures of skill such as S₁ scores, other less conventional statistics such as the structure function of various variables or the separation of errors by scale would be useful in assessing a model's realism, the rate of improvement over time and in model inter-comparisons. Examples of these measures of skill are provided in Section 2.

In constructing improved models, there are always tradeoffs between accuracy and computational requirements. In addition, a balance must be maintained between accuracy and computational expense among the various components of the model. A highly sophisticated, computationally expensive physical parameterization, for example, is unjustified unless the errors associated with other physical and numerical approximations are of the same order as the errors introduced by that particular process.

To close, I speculate, based on the results reviewed in this paper, how a state-of-the-art regional model might be constructed in order to make further improvements in forecast skill. This hypothetical model would contain about 20 layers and utilize fourth-order spatial finite differencing on a staggered horizontal grid with an implicit (or semi-implicit) temporal integration scheme. The high resolution (~50 km) portion of the grid would be nested within a larger scale, global model. It would be analyzed and initialized by a combination of an optimal interpolation analysis scheme and a nonlinear normal mode initialization procedure.

The physical parameterizations would include a medium-resolution (about five layers) PBL formulation coupled with a surface energy budget and a diurnal cycle. The parameterization of the effects of cumulus convection would depend on the dynamic and thermodynamic structure and evolution of the resolvable scales (such as moisture convergence or rate of destabilization and the vertical temperature and moisture structure). The effect of nonprecipitating layered clouds would be considered in the surface energy budget by relating the amount of these clouds to the mean layer relative humidity.

While no one model containing all of the above features exists, each feature may be found in at least one model and many may be found in several. Such a model is computationally feasible with existing computers, and evidence may be found in the literature for the importance and potential for improvement in each factor. While a significant advancement in the total model's performance resulting from the net effect of many small improvements cannot be
predicted with certainty, if an overall improvement is to occur, there appear to be few alternative strategies.

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