Development and Use of the NCAR GCM

A Report of the GCM Steering Committee
When development of a General Circulation Model (GCM) at NCAR began in 1965, it was intended to make the model widely available to the atmospheric science community as a facility for numerical experimentation, as well as to fulfill a need at NCAR for studying the large-scale dynamics of the atmosphere. Although the principles of formulating the GCM are well known, building an advanced GCM requires extensive research into mathematical formulations, numerical analysis, parameterizations of physical processes and efficient programming. Above all, it consumes a large amount of computing time for testing and experimentation.

It is imperative, therefore, that the NCAR GCM be flexible enough that modifications can be readily incorporated with a minimum amount of recoding. In other words, scientists should be able to perform experiments knowing only those details of the model which apply directly to their particular work. To achieve this objective, we have designed the model program in modular fashion so that each physics component can be added or withdrawn without affecting the main dynamical framework.

The demand to use the existing GCMs for experimentation by scientists inside and outside NCAR has increased significantly in the past few years. This, along with a general increase in computer use
at NCAR, has decreased the computer time available for the development of more flexible and advanced GCMs and, consequently, there is a need for setting priorities.

To review past and present GCM activities at NCAR and to plan for the future, the GCM Steering Committee (see members listed on page vii) has been set up by F. Bretherton. One principal task of the committee was to prepare and discuss the following sections of this report:

1. **Classification of NCAR GCMs** (describing the characteristics of different model versions).
2. **Climatological Statistics** (compiled from computer simulations with various NCAR GCMs and comparison with observed statistics).
3. **Model Sensitivity for Climate Simulation.**
4. **One- to Seven-Day Forecasts of Large-Scale Atmospheric Flow Generated from the GCM** (application of the GCM to medium-range weather prediction experiments including objective data analysis techniques).
5. **Conclusions.**

Sections 1 and 3 were prepared by W. Washington, Section 4 by D. Baumhefner and Sections 2 and 5 by A. Kasahara, who also served as editor for the entire document. Those assisting in the writing and reviewing were R. Chervin, T. Schlatter, J. Williams, D. Williamson and the committee members. Since the principal objective in writing
this document is to review specifically GCM activities at NCAR, the presentation of GCM results and the related discussion are confined only to the NCAR GCM. Active discussions during the meetings focused on the strengths and weaknesses of the NCAR GCM and the areas needing improvement; these are summarized in Section 5. As a result of the extensive review, the committee is now in a better position to set priorities with the GCM and to establish effective coordination between GCM users and the Computing Facility. The committee hopes that this document will help GCM users to better judge the performance of the NCAR GCMs, design needed numerical experiments and determine the areas needing improvement.

In consultation with the GCM Advisory Panel members (currently A. Arakawa, D. Houghton, R. Lindzen and P. Merilees), the GCM Steering Committee will act on research proposals relating to the development and use of the GCM, allocate necessary computing and manpower resources and follow through on their accomplishments.

May 1975
GCM STEERING COMMITTEE MEMBERS:

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References
1.0 Classification of NCAR GCMs

1.1 First-Generation Model

Development of the first-generation NCAR GCM began in 1965 with acquisition of a Control Data Corporation 3600 computer allowing solution of a low-resolution version of the GCM. The early version had two 6-km thick layers in the vertical and a horizontal resolution of 5° in latitude and longitude with longitudinal grid skipping poleward of 60°. The physical processes were quite simple compared to the current versions. For example, the mean climatological distribution of January surface temperature was specified over the entire globe; we also assumed a saturated earth's surface. Orography was not included, nor was an explicit calculation of moisture. The latent heat release was assumed proportional to the upward vertical velocity. The horizontal eddy viscosity was constant. An infrared radiation computer program was developed by Sasamori (1968a,b) based on radiation-chart approximations of fluxes resulting in considerable savings of computer time. Cloudiness was specified from zonal climatology. During these early years, a great deal of attention was given to numerical schemes, particularly Kasahara (1965), Houghton et al. (1966) and Williamson (1966). In fact, we have continually examined numerical schemes for improving the model (Williamson, 1968, 1969, 1970, 1971; Williamson and Browning, 1973). Details
of the first-generation model, as well as the basic framework for later generations, are described in Kasahara and Washington (1967).

In 1968-1969, an explicit prediction of moisture was added because the earlier assumption of saturation led to continuous atmospheric warming. With inclusion of a full hydrologic cycle, this no longer occurs because a balance between evaporation and precipitation is maintained. A non-constant horizontal eddy viscosity was added as a function of flow deformation and the change greatly reduced the overdamping of the baroclinic disturbances in the model. Washington and Kasahara (1970) compared the results of the two-layer model with observed data and, in addition, computed momentum, energy and moisture budgets in terms of eddy and mean circulations.

The first-generation model is no longer available, but Table 1.1 summarizes its characteristics.

1.2 Second-Generation Model
1.2.1 Tropospheric version

Development of the second-generation model began in 1968 when the first-generation model proved inflexible in programming additional features. The major changes incorporated into the second-generation model were more efficient programming, diagnostically determined cloudiness, surface temperature calculation over non-ocean regions based upon a local surface energy balance and inclusion of orography. Since the NCAR GCM uses geometric
height as the vertical coordinate and the earth's surface is not a coordinate surface, a special procedure is needed to incorporate the earth's orography into the model. Orography was first included in a two-layer version of the model (Kasahara and Washington, 1969) and later an improved treatment was developed for a six-layer version (Kasahara and Washington, 1971).

Details of the basic second-generation model appear in Oliger et al. (1970). This model has six layers with a thickness of 3 km and a horizontal resolution of 5° in latitude and longitude. Various eddy viscosity formulations, convective parameterizations, surface hydrology, seasonal changes and boundary-layer parameterizations have been tested. Kasahara and Washington (1971) discussed the results of a January simulation with the six-layer version including momentum, energy and moisture budgets. The general characteristics are summarized in Table 1.2; in Table 1.3 are the specific characteristics of the version used in most climate sensitivity experiments (see Section 3).

1.2.2 Tropospheric-stratospheric model

To investigate the lower stratosphere, we developed a model having the same vertical resolution as the tropospheric version, but with six additional layers at the top, thus extending the model height from 18 to 36 km. The method of calculating solar and infrared radiation is described in Sasamori et al. (1972).
Ozone heating was included in the radiation calculation in the lower stratosphere. Since the ozone processes are not known well enough to predict, we specified a latitude-height climatological distribution for the radiation calculation. The description of this model is contained in Kasahara et al. (1973) and Kasahara and Sasamori (1974), with computed results for a January simulation with and without large-scale mountain effects.

1.2.3 High-resolution global tropospheric model

The horizontal resolution of the 5° tropospheric model is too coarse for prediction of many features of regional interest. To alleviate this problem, we reduced the horizontal resolution from 5° to 2.5°. This higher resolution model includes elaborate parameterization schemes. Two new features are the Krishnamurti and Moxim (1971) convective parameterization and predictive equations for snow cover and soil moisture (Washington, 1974). Table 1.2 summarizes the physical processes included in this model, and Table 1.5 lists the computer time for various model versions.

1.2.4 Limited-area model (LAM)

For numerical weather prediction studies on a very fine scale, we have developed a limited-area version of the second-generation model. The lateral boundary conditions for this model are obtained by first running a forecast with a low-resolution global model and generating inflow boundary conditions for the limited-area model. The outflow boundary conditions are provided by
Lagrangian extrapolation (see Williamson and Browning, 1974, for details). As shown in Table 1.5, the limited-area model has been tested on several different meshes of varying resolution.

1.3 Third-Generation Model

Development of an improved and more flexible third-generation NCAR GCM was begun in 1972. The numerical procedure of reducing gridpoints at high latitudes in the second-generation model results in low accuracy. Williamson and Browning (1973) showed that substantial improvement could be made by restoring skipped gridpoints and eliminating computationally unstable modes by Fourier filtering. The new model also uses a fourth-order finite-difference scheme to reduce truncation errors. To run efficiently on the next-generation computers, the new model adopts transformed height (Kasahara, 1974) as the vertical coordinate. The program is designed so that with relatively minor modifications the user can choose a variable vertical increment and even select a different variable as the vertical coordinate. Thus far, the options in the program have been tested only for geometric height. Various physical processes in the third-generation model are essentially the same as in the second-generation model. However, the solar and infrared radiation programs are improved and cloudiness can be included at all levels rather than the two levels in the second-generation model. A new
boundary-layer treatment is being developed. Table 1.4 summarizes the physical processes included in the model.
Table 1.1 First-Generation NCAR GCM Characteristics

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<table>
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<tbody>
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<td>1)</td>
<td>Vertical coordinate</td>
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<td>2)</td>
<td>Horizontal domain</td>
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<td>3)</td>
<td>Horizontal resolution (latitude and longitude)</td>
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<tr>
<td>4)</td>
<td>Horizontal approximation of derivatives</td>
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<td>5)</td>
<td>Vertical resolution and number of layers</td>
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<tr>
<td>6)</td>
<td>Height of the top</td>
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<td>7)</td>
<td>Horizontal diffusion</td>
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<td>8)</td>
<td>Moisture processes</td>
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<td>Surface hydrology</td>
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<td>Surface temperature calculation</td>
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<td>Cloudiness</td>
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<td>Convective parameterization</td>
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<td>Seasonal change</td>
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<td>Ozone heating</td>
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<td>Boundary layer</td>
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<td>16)</td>
<td>Orography</td>
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<td>17)</td>
<td>Ocean assumption</td>
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<tr>
<td>18)</td>
<td>Computer time</td>
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Table 1.2 Second-Generation NCAR GCM Characteristics

<table>
<thead>
<tr>
<th></th>
<th>Vertical coordinate</th>
<th>1) Vertical coordinate geometric height</th>
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<tbody>
<tr>
<td>2)</td>
<td>Horizontal domain</td>
<td>global, hemispheric and limited-area versions</td>
</tr>
<tr>
<td>3)</td>
<td>Horizontal resolution (latitude and longitude)</td>
<td>5°, 2.5°, global; 1.25°, 5/8° limited area</td>
</tr>
<tr>
<td>4)</td>
<td>Horizontal approximation of derivatives</td>
<td>second-order</td>
</tr>
<tr>
<td>5)</td>
<td>Vertical resolution and number of layers</td>
<td>3 km, 6 or 12 layers in global; 6, 12, 24 layers in limited area</td>
</tr>
<tr>
<td>6)</td>
<td>Height of the top</td>
<td>18 or 36 km in global, variable in limited area</td>
</tr>
<tr>
<td>7)</td>
<td>Horizontal diffusion</td>
<td>variable eddy viscosity</td>
</tr>
<tr>
<td>8)</td>
<td>Moisture processes</td>
<td>explicit hydrologic cycle, Kuo-Krishnamurti convective parameterization</td>
</tr>
<tr>
<td>9)</td>
<td>Surface hydrology</td>
<td>ground saturated or explicit prediction of snow cover and soil moisture (in some versions)</td>
</tr>
<tr>
<td>10)</td>
<td>Surface temperature calculation</td>
<td>computed from surface energy balance</td>
</tr>
<tr>
<td>11)</td>
<td>Cloudiness</td>
<td>diagnostically determined from relative humidity and vertical motion at 3 and 9 km</td>
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<tr>
<td>12)</td>
<td>Convective parameterization</td>
<td>convective adjustment (see 8)</td>
</tr>
<tr>
<td>13)</td>
<td>Seasonal change</td>
<td>perpetual months or seasonal change</td>
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<tr>
<td>14)</td>
<td>Ozone heating</td>
<td>in stratospheric version</td>
</tr>
<tr>
<td>15)</td>
<td>Boundary layer</td>
<td>bulk transfer coefficient method and Deardorff formulation</td>
</tr>
<tr>
<td>16)</td>
<td>Orography</td>
<td>realistic consistent with resolution</td>
</tr>
<tr>
<td>17)</td>
<td>Ocean assumption</td>
<td>climatological surface temperature; or computed from surface energy</td>
</tr>
<tr>
<td>18)</td>
<td>Computer time</td>
<td>see Table 1.5</td>
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1There are many model variants in terms of physical processes and resolution.
<table>
<thead>
<tr>
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<th>Second-Generation NCAR GCM Used for Most Climate Sensitivity Experiments&lt;sup&gt;1&lt;/sup&gt;</th>
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<td>1)</td>
<td>Vertical coordinate</td>
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<td>2)</td>
<td>Horizontal domain</td>
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<td>3)</td>
<td>Horizontal resolution (latitude and longitude)</td>
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<tr>
<td>4)</td>
<td>Horizontal approximation of derivatives</td>
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<tr>
<td>5)</td>
<td>Vertical resolution and number of layers</td>
</tr>
<tr>
<td>6)</td>
<td>Height of the top</td>
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<td>Horizontal diffusion</td>
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<tr>
<td>8)</td>
<td>Moisture processes</td>
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<td>Boundary layer</td>
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<td>17)</td>
<td>Ocean assumption</td>
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<td>18)</td>
<td>Computer time</td>
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</tbody>
</table>

<sup>1</sup>There are many model variants in terms of physical processes and resolution.
Table 1.4 Third-Generation NCAR GCM Characteristics

<table>
<thead>
<tr>
<th></th>
<th>Description</th>
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<tr>
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<td>Horizontal domain</td>
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<td>3</td>
<td>Horizontal resolution (latitude and longitude)</td>
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<tr>
<td>5</td>
<td>Vertical resolution and number of layers</td>
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<tr>
<td>6</td>
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<td>17</td>
<td>Ocean assumption</td>
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<td>18</td>
<td>Computer time</td>
</tr>
</tbody>
</table>

1This model is still in early development stage and available only for limited use.
Table 1.5 Versions of NCAR GCM

<table>
<thead>
<tr>
<th></th>
<th>CPU Hours (7600)</th>
<th>CRUs</th>
<th>Wall Clock Hours</th>
</tr>
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<tbody>
<tr>
<td><strong>A. SECOND-GENERATION</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Global</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6 layer, 5°</td>
<td>0.2</td>
<td>600</td>
<td>0.24</td>
</tr>
<tr>
<td>12 layer, 5°</td>
<td>0.4</td>
<td>1,200</td>
<td>0.48</td>
</tr>
<tr>
<td>6 layer, 2.5°</td>
<td>1.6</td>
<td>4,800</td>
<td>1.92</td>
</tr>
<tr>
<td><strong>Limited Area</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6 layer, 1.25°</td>
<td>1.0-2.0</td>
<td>3,000-6,000</td>
<td>1.2-2.4</td>
</tr>
<tr>
<td>12 layer, 1.25°</td>
<td>2.0-4.0</td>
<td>6,000-12,000</td>
<td>2.4-4.8</td>
</tr>
<tr>
<td>24 layer, 1.25°</td>
<td>4.0-8.0</td>
<td>12,000-24,000</td>
<td>4.8-9.6</td>
</tr>
<tr>
<td>24 layer, 0.625°</td>
<td>32.0-64.0</td>
<td>96,000-192,000</td>
<td>38.4-76.8</td>
</tr>
<tr>
<td><strong>B. THIRD-GENERATION</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6 layer, 5.625°</td>
<td>0.33</td>
<td>990</td>
<td>0.4</td>
</tr>
<tr>
<td>6 layer, 2.8125°</td>
<td>2.8(^{6})</td>
<td>8,400(^{6})</td>
<td>3.4(^{6})</td>
</tr>
<tr>
<td>24 layer, 5.625°</td>
<td>1.50(^{3,6})</td>
<td>4,500(^{3,6})</td>
<td>1.8(^{3,6})</td>
</tr>
</tbody>
</table>

1 The first number indicates layers in vertical; the second is size of horizontal grid in degrees of longitude and latitude.

2 Assume that 1 hour CPU (Control Data Corporation 7600) = 2,700 CRUs. The estimated cost of 1 CRU = $0.30.

3 Estimated.

4 Computer time needed for running LAM depends on the size of forecasting domain.

5 These times will be somewhat less with the version making more use of parallel processing of instructions where the program is not input-output limited. This requires recoding time-consuming portions of the model.
2.0 Climatological Statistics

We present in this section the climatological statistics generated by various NCAR GCMs. Short-range predictions are largely influenced by initial conditions (including the data initialization procedure). When the period of prediction is extended beyond one week or so, the model gradually approaches its own statistical equilibrium state—called model climate. It is important in modeling research to examine the make-up of model climate, how closely it compares with the real climate and how fast the model approaches its climate. In investigating these questions, we consider (1) global mean energies, (2) zonal mean meteorological variables and (3) meridional transports by mean circulation and eddy motions. If we call the zonal mean state the first-moment quantity, meridional transports may be referred to as second-moment quantities. Since we do not yet have reliable estimates of the third-moment quantities from the real atmosphere, we will not cover here energy budgets in detail. In addition to these three aspects, we briefly discuss the spectra and the geographical distributions of atmospheric variables.

We confine our presentation primarily to results from the second-generation models unless otherwise stated. The full processor for computing transports and energy statistics for the third-generation models has not been completed.
2.1 Global Mean Energies

The global mean values of zonal kinetic energy $\overline{K}$, eddy kinetic energy $K'$, internal energy $I$ and potential energy $P$ are computed in the models. Here,

$$\overline{K} = \frac{1}{2} \rho (u^2 + v^2)$$
$$K' = \frac{1}{2} \rho (u'^2 + v'^2)$$
$$I = C_v \rho T$$
$$P = g \rho z$$

where $u$ and $v$ denote the longitudinal and meridional components of horizontal velocity, $\rho$ density, $T$ temperature, $C_v$ specific heat of air at constant volume, $z$ geometric height, $g$ acceleration due to gravity and for any variable $A$

$$\overline{A} = \frac{1}{2\pi} \int_0^{2\pi} A d\lambda \quad (\lambda: \text{longitude}) \quad (1)$$

and

$$\hat{A} = \frac{A}{\rho} , \quad A = \hat{A} + A', \quad \rho A' = 0 .$$

Quantity $\overline{I}$ is a measure of the density-weighted zonal mean temperature, where $\overline{P}$ denotes a measure of zonal mean density at a given height.

Figure 2.1 shows the time variations of global mean $\overline{K}$ and $K'$ for two cases--with and without the earth's orography for January simulated by a $5^\circ$ 6-layer model (Kasahara and Washington, 1971).
The no-mountain case (NM) has a constant elevation of 10 m everywhere over land. For each case, $\bar{K}$ tends to increase during the period, but $K'$ reaches equilibrium after Day 20 or so. A question is then raised concerning the long-term trend of $\bar{K}$, namely, whether $\bar{K}$ will ever reach equilibrium. To look into this matter, we plotted on Fig. 2.2 the time variations of $\bar{K}$ for other runs simulating January with a 5° 6-layer model. Case 138 is identical to Case M (mountain case) shown in Fig. 2.1. Case 156 is an updated version of Case 138 in which many minor improvements (such as changing albedo values over the oceans and boundary-layer formulations) were incorporated in the model. Case 331 includes the surface pressure adjustment (see Section 3.4). Case 138 has not been run beyond Day 70 because Case 156 was more realistic. Note that $\bar{K}$ started to level off beyond Days 40-50 for both 138 and 331. On the same figure, we plotted $\bar{K}$ and $K'$ from a high-resolution (2.5°) version of the 6-layer model. For this case, $\bar{K}$ reached equilibrium after Day 40 or so, while $K'$ varied around a constant mean value throughout the period. At the right, we indicate the observed ranges of $\bar{K}$ and $K'$. It is significant that $\bar{K}$ is overestimated for both the 2.5° and 5° models, whereas $K'$ is underestimated, particularly in the 5° model. The 2.5° model nearly doubled the level of $K'$ which is much closer to the observed than the 5° model.
The reason why $\overline{K}$ is overestimated in the models may be related to the fact that the mean meridional temperature gradient in the models is overestimated in the middle latitudes (for example, see Fig. 2.8). As a consequence, the intensity of zonal wind becomes exaggerated (compare Fig. 2.9(a) and (b)). One way to reduce the mean meridional temperature in the models is to lower the upper tropospheric temperature in the tropics, which at present is about several degrees higher than the observed. The reduction of upper tropospheric temperature in the tropics is observed in the GCM with a different convective parameterization (discussed in Section 3.5.4). It is entirely possible that the reduction of the mean meridional temperature gradient may affect the weakening of baroclinic activity in the middle latitudes and lower the magnitude of $K'$ somewhat.

Figure 2.3 illustrates the time variations of the global mean $\overline{K}$ and $K'$ for July (Williams, 1974) with the $5^\circ$ 6-layer model; both quantities reach equilibrium. The control case is the standard run without modification of the boundary conditions for ice-age experiments.

Figure 2.4 shows the time variations of the global mean $\overline{K}$, $K'$ and $\overline{I}$ for January simulated by the $5^\circ$ 12-layer tropospheric-stratospheric model (Kasahara et al., 1973). While $\overline{K}$ and $\overline{I}$ still increase after Day 120, $K'$ reaches equilibrium after Day 20. We found that the increase of $\overline{I}$ is due to a gradual increase
of mean density $\bar{\rho}$ rather than the increase of mean temperature $\bar{T}$. The increase in mean density is caused by the accumulated truncation error referred to in Section 3.4.

We have also run 18 months with the $5^\circ$ 12-layer model by allowing day-by-day progression of the sun's declination and updating the ozone distribution and the ocean surface temperature interpolated on a daily basis from monthly mean fields. Detailed analyses of this case have not yet been completed and will be reported elsewhere, but we show in Fig. 2.5 the time variations of $K'$, $\bar{K}$ and $\bar{T}$. This run incorporates surface pressure adjustment (Section 3.4). $K'$ appears to be in statistical equilibrium during the 18 months, whereas $\bar{K}$ shows a clear seasonal variation indicating maximum around December-January and minimum in July-August. On the other hand, the global mean internal energy $\bar{I}$ reaches maximum around July-August and minimum about December-January.

Figures 2.6 and 2.7 depict the latitude-height distributions of $\bar{K}$ and $K'$, respectively, for three versions of the 6-layer model with different horizontal grid resolutions for a January simulation (Wellick et al., 1971). The three different cases are identified by the horizontal grid size indicated in the upper right corner of each figure. The patterns of $\bar{K}$ do not vary appreciably for the three resolution cases, whereas those of $K'$ demonstrate the significant influence of horizontal grid resolution.
As seen from Fig. 2.2, the eddy kinetic energy level of the 2.5° model shows considerable improvement over that of the 5° model. We will come back to this point again in Section 2.4.

2.2 Zonal Mean States

We present zonal mean values of computed atmospheric-state variables, $\bar{u}$ (longitudinal velocity component), $\bar{v}$ (meridional velocity component), $\bar{w}$ (vertical velocity component), $\bar{T}$ (temperature), $\bar{q}$ (specific humidity) and cloudiness. All are averaged in time for 30 days.

Figures 2.8 and 2.9 show the latitude-height distributions of $\bar{T}$, $\bar{q}$, $\bar{u}$, $\bar{v}$ and $\bar{w}$ for January simulated by the 5° 6-layer model with and without the earth's orography and compared with observed data (Kasahara and Washington, 1971). The heavy solid lines are contours for the no-mountain case and the thin lines are for the mountain case. Note that the zonal mean temperature distribution near the surface and 1.5 km in the middle latitudes for the mountain case should be ignored since the surface temperature is calculated at a different elevation. Although the two cases simulate generally well the lower tropospheric temperature field, both fail to produce the colder upper troposphere in the tropics. This difficulty seems to be due, in part, to the upper boundary condition, i.e., vertical motion vanishes at the 18-km top. The model fails to simulate the warm lower stratosphere; this shortcoming may be attributed to the lack of solar heating by ozone.
in the lower stratosphere in this particular version. On the other hand, the computed specific humidity distributions agree well with the observed.

As evident from Fig. 2.9, the computed westerly jet maxima appear too high in altitude. The Northern Hemisphere (N.H.) jet is correctly located in the observed latitude, but the Southern Hemisphere (S.H.) jet appears somewhat closer to the equator for cases with and without mountains compared to the observed. A major deficiency in the simulation of the zonal wind field is the lack of easterlies in the tropics. It is speculated that an excessive amount of horizontal diffusion of momentum and/or a deficiency in the vertical momentum diffusion may be responsible for this failure. Recently, it was shown by simulation with a general circulation model that the increase of vertical momentum diffusion contributed to the formation of deeper easterlies in the tropics (Stone et al., 1974). The subgrid-scale mixing in the tropics may be related to the parameterization of cumulus convection. Research is being conducted to clarify this question.

The zonally averaged meridional wind component \( \overline{v} \) and vertical wind component \( \overline{w} \) shown in Fig. 2.9 illustrate the Hadley circulation. In comparison with the observed patterns, the simulation somewhat overestimates the intensity of the Hadley circulation. The intensity of the Hadley circulation is very much dependent
on the amount of released latent heat of condensation, which is overestimated in this model. Improvement in the parameterization of cumulus convection may alleviate the problem. We should not overlook, however, the fact that $K'$ in the $5^\circ$ model is underevaluated and then compensated for by overevaluation of $\overline{K}$ (see discussion of $\overline{K}$ on page 15). Thus, it may not be as simple as just speculated, since horizontal grid resolution is an equally important factor.

Williams et al. (1974) and Williams (1974) analyzed the results of the July simulation with the $5^\circ$ 6-layer model. Since details are described in these papers, we do not discuss them here except to point out that several model deficiencies in the January simulation also appear in the July simulation. After taking into account these model deficiencies, the simulation of ice-age climate reveals broad agreement with paleoclimatological reconstructions.

For stratospheric simulations, Figs. 2.10 and 2.11 present the latitude-height distributions of $\overline{T}$, $\overline{u}$, $\overline{v}$ and $\overline{w}$ for Case NM (without mountains), Case M (with mountains) computed for January by the $5^\circ$ 12-layer model (Kasahara et al., 1973) and compared with the long-term observed means.

The temperature field in the troposphere is generally well simulated, but a few discrepancies are noted in the stratosphere. The minimum temperature at the tropopause in the tropics is about $10^\circ$C colder and the height of the computed tropopause is a few
kilometers too high. The vertical temperature stratification in the stratosphere is too stable. We will return later to the question of diagnosis.

The computed zonal wind distributions, in general, agree well with the observed. The tropospheric westerly jets are clearly separated from the polar night jet in the N.H. and the stratospheric easterly jet in the S.H. With the mountains, the intensity of the polar night jet is reduced considerably. As shown in Section 3.5.3, the presence of mountains is crucial to the formation of the Aleutian High in the stratosphere. The deviation of the jet stream from an otherwise almost circular symmetric pattern apparently resulted in a weaker zonally averaged westerly flow.

The zonally averaged meridional wind component $\overline{v}$ and vertical wind component $\overline{w}$ shown in Fig. 2.11 constitute the Hadley circulation. In the troposphere, the three-cell structure is reproduced. In the stratosphere, there are essentially two ascending and two descending motion areas spanning both hemispheres. In the N.H., ascending motions are found over the polar region and descending motions in the middle latitudes. This branch of descending motion connects directly with the downward branch of the N.H. tropospheric Hadley cell through a tropopause break.
The seasonal run with the 12-layer model, as indicated earlier, demonstrates a clear reversal of the tropospheric Hadley circulation from January to July and July to January. The Hadley circulation becomes considerably weaker during the transition seasons of April-May and October-November.

Wellick et al. (1971) investigated the dependence of the latitude-height distributions of $\overline{T}$, $\overline{q}$, $\overline{u}$ and $\overline{v}$ on horizontal grid resolutions comparing the results of January simulation using 10°, 5° and 2.5° meshes with the 6-layer model. The grid resolution does not significantly affect the distributions of temperature and specific humidity, but the 2.5° model seems to simulate the Hadley circulation closer to the observed. The 2.5° model, however, did not simulate the deep tropical easterlies; these results were based on an early version of the 2.5° model. Many improvements in the model have been made since then and a climatological examination of the latest 2.5° model is now being conducted (see Section 3.5.4).

The rate of heating/cooling in the model is expressed by

$$ Q = Q_{dv} + Q_{dH} + Q_{al} + Q_{as} + Q_c $$

where the five parts of the heating rate are due to vertical diffusion of sensible heat $Q_{dv}$, horizontal diffusion of sensible heat $Q_{dH}$, longwave cooling of the atmosphere $Q_{al}$, absorption of
solar radiation by water vapor $Q_a$ and the release of latent heat of condensation of water vapor $Q_c$.

The five heating rates are calculated in the model. Figure 2.12 shows the latitudinal distribution of vertically integrated total heating $\int \rho Q \, dz$ from the earth's surface to the top of the model atmosphere for January calculated from the $5^\circ$ 6-layer model (Kasahara and Washington, 1971). Considering the difficulty in evaluating the so-called observed data, agreement between the computed and observed distributions is reasonably good.

Just as $Q$ is the source term in the thermodynamic equation, $M + \rho E$ is the source term for the continuity equation of water vapor, where $M$ is the rate of condensation of water vapor and $\rho E$ the rate of change of water vapor content due to subgrid-scale diffusion. In effect, the vertically integrated $\int (\bar{M} + \bar{\rho E}) \, dz$ from the earth's surface to the model top is equivalent to the difference between precipitation and evaporation per unit area in a zonally averaged meridional cross-section. Figure 2.13 shows the difference between zonally averaged evaporation and precipitation calculated from the $5^\circ$ 6-layer model for January. Agreement between the computed and observed is again reasonably good.

Figure 2.14 presents the computed zonal mean total cloudiness for January and July simulated with the $5^\circ$ 6-layer model and
the results are compared with observed data (Williams, 1974). In July, the computed cloudiness agrees well with the observed data in the N.H. of Telegadas and London (1954). In January, the computed values are larger than observed in the N.H. higher latitudes and lower in the N.H. tropics. For the S.H., the agreement between the computed and van Loon's (1972) data is satisfactory in January, but in July the computed values are too large between 35-55°S and too low for 0-30°S.

2.3 Meridional Transports by Mean Circulation and Eddy Motion

We show here the meridional transports:

<table>
<thead>
<tr>
<th>Mean Circulation</th>
<th>Eddy Motion</th>
</tr>
</thead>
<tbody>
<tr>
<td>Horizontal momentum</td>
<td>$\rho u \hat{v}$</td>
</tr>
<tr>
<td>Sensible heat</td>
<td>$\rho T \hat{v}$</td>
</tr>
<tr>
<td>Water vapor</td>
<td>$\rho q \hat{v}$</td>
</tr>
</tbody>
</table>

Since the density changes appreciably with height, the latitude-height distributions are given without the density factor.

Figure 2.15 presents the latitude-height distributions of eddy momentum transports: (a) observed total (transient plus standing) eddy momentum transport for December-February; (b) observed transient eddy momentum transport for December-February and (c) 30-day mean total eddy momentum transport.
computed by the 5° 6-layer model without mountains for January (Kasahara and Washington, 1971). Although the magnitude of computed transport in the middle latitudes is underestimated, the overall distribution agrees with the observed pattern.

Figure 2.16 shows the vertical integrals of $\bar{\rho}u' v' \cos \varphi$ and $\rho u' v' \cos \varphi$ as functions of latitude, where $a$ is the earth's radius and $\varphi$ is latitude. Compared with the observed distributions by Oort and Rasmusson (1971) for January over the N.H., the computed transport by mean circulation is overestimated, but the computed eddy transport agrees fairly well with the observed. Similar comparisons appear in Fig. 2.17 for the vertical integrals of water vapor transports and in Figs. 2.18 and 2.19 for the vertical integrals of sensible transports by mean circulation and eddy motion.

For a July simulation with the 5° 6-layer model, Williams (1974) presents in Figs. 2.20 and 2.21 the vertical integrals of $\bar{\rho}u' v' \cos \varphi$ and $\rho u' v' \cos \varphi$ as functions of latitude. Note that in January the momentum transport by mean circulation in the tropics is northward, whereas in July it is southward. On the other hand, the momentum transport by eddy motion does not change much from January to July.

Similarly, Fig. 2.22 shows the vertical integrals of water vapor transports by mean circulation and eddy motion. Although
the water vapor transport by mean circulation is overestimated, the shift of direction from January to July is evident. This is an important feature of the earth's atmosphere, namely, the transports of momentum, water vapor and sensible heat by the Hadley circulation change sign from January to July, but the eddy transports primarily by baroclinic eddies in the middle latitudes do not vary appreciably from one season to another.

For a January simulation with the 5° 12-layer model, the left side of Fig. 2.23 presents the zonally averaged mean cross sections for the northward flux of westerly momentum of the no-mountain case, mountain case and observed long-term mean. The right shows the same as left but for the northward flux of sensible heat for January (Kasahara et al., 1973). The observed data are taken from Newell et al. (1969). The magnitude of eddy transports in the stratosphere is unreliable due to the sparsity of observed data. Hence, the observed data in the stratosphere should be used only as a guide. The computed eddy momentum transport in the troposphere is underevaluated, as expected from the 5° resolution model.

The tropospheric portion of the present results is directly comparable to that of the 5° 6-layer model, since both tropospheres have the same 6 layers. The addition of the stratosphere, however, did little to improve the simulation of the troposphere. On the
other hand, we found that the stratosphere is influenced significantly by the tropospheric circulation. The primary source of eddy kinetic energy in the stratosphere is the vertical transport of wave energy through the tropopause (Kasahara and Sasamori, 1974).

In connection with the temperature field simulated by the 5° 12-layer model (Fig. 2.10), we pointed out that the vertical temperature stratification in the stratosphere is stabler than the observed. The question is then raised that the radiative balance may not be well maintained. To understand the zonally averaged temperature distribution, we need detailed information on dynamics as well as radiative balance.

By applying the density-weighted zonal averaging (1) to the thermodynamic equation, we find

\[ C_v \rho \frac{\partial T}{\partial t} + C_v \frac{\partial}{a} \frac{\partial T}{\partial \phi} + C_v \rho \theta \frac{\partial T}{\partial z} + \frac{C_v}{a \cos \phi} \left( \rho T' \theta' \cos \phi \right) + C_v \frac{\partial}{\partial \phi} \left( \rho T' \omega' \right) \]

\[ = -C(I,K') - C(I,K + P) + \rho \Omega \]

where

\[ C(I,K') = \rho \left( \nabla \cdot \omega' + \frac{\partial \omega'}{\partial z} \right) \]

\[ \nabla \cdot \omega' = \frac{1}{a \cos \phi} \left[ \frac{\partial u'}{\partial \lambda} + \frac{\partial}{\partial \phi} (v' \cos \phi) \right] \]

\[ C(I,K + P) = \rho \left[ \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (\theta \cos \phi) + \frac{\partial \theta}{\partial z} \right] \]
The term \( C(\overline{I}, K') \) indicates energy conversion from \( \overline{I} \) to \( K' \) if the quantity is positive and vice versa if negative. \( C(\overline{I}, \overline{K + P}) \) shows energy conversion between \( \overline{I} \) and \( \overline{K + P} \).

The time change of zonally averaged temperature thus depends on horizontal and vertical advections of zonal mean temperature, horizontal and vertical divergences of eddy heat fluxes, energy conversions between \( \overline{I} \leftrightarrow K' \) and \( I \leftrightarrow \overline{K + P} \) and heating \( \rho Q \).

Figure 2.24 presents the latitudinal distribution of each term in the above equation integrated vertically in the stratosphere between 18 and 36 km. We took a time average of 30 days so that the first term in the equation is assumed negligible. The net heating term \( \sum_s \rho Q \Delta z \) is positive between \( 20^\circ \text{N}-30^\circ \text{S} \) on the order of \( 2 \times 10^3 \text{ gm sec}^{-3} \). This value is considerably smaller than the maximum net vertically integrated heating in the troposphere which is approximately \( 10^5 \text{ gm sec}^{-3} \). Although the net heating contributes substantially to the zonal mean temperature budget, the energy conversion between \( \overline{I} \leftrightarrow \overline{K + P} \) and the divergence of eddy heat flux both contribute significantly.

The convergence of the horizontal transport of sensible heat by eddies \( C_v \overline{\nu' T'} \) tends to compensate the conversion term \(-C(\overline{I}, \overline{K + P})\). The dominant term in \( C(\overline{I}, \overline{K + P}) \) is \( \overline{\rho \omega} / \Delta z \). The vertically integrated value of the latter is approximately equal to \(-\sum_s \rho g \Delta z \) which corresponds approximately to adiabatic
compressional heating by the mean vertical motion. In the stratosphere, a major descending branch exists between 20-55°N and an ascending branch poleward from 55°N (Fig. 2.11, right side). The distribution of \( C(\overline{I, K+P}) \) in Fig. 2.24 reflects the mean vertical motion pattern. For more detailed analysis of the energetics of the stratosphere for this case, see Kasahara and Sasamori (1974).

Before concluding this subsection, we should point out the dependence on horizontal grid resolutions of various meridional transports by mean circulation and eddy motions. Wellick et al. (1971) discussed in detail the latitude-height distributions of meridional transports of momentum, heat and moisture by eddy motion and mean circulation for the three different horizontal resolution versions (10°, 5° and 2.5°) simulating January and compared the results with observed data. General conclusions from this resolution experiment are: (1) the 2.5° 6-layer model reduces the intensity of the Hadley circulation which is too strong in the 5° models, and (2) the 2.5° 6-layer model increases the strength of eddy transport, particularly eddy momentum transport in the middle latitudes which is too weak in the 5° models.

2.4 Spectra of Meteorological Variables

2.4.1 Zonal spectra of kinetic energy

To further demonstrate the effects of increased horizontal grid resolution on eddy motions, we present in Fig. 2.25 the
spectral distribution of zonal eddy kinetic energy at 40N from the three different grid versions.

Consistent with the conclusions in Section 2.3, we find that the horizontal grid resolution has a significant influence on the intensity of eddy activity. The 10° mesh generates weak eddy activity in the high wavenumber components; this deficiency is improved in the 5° mesh. A major difference between the 5° and 2.5° mesh cases is found in the increase of kinetic energy in the low wavenumber components 1 to 4. Since large-scale atmospheric motions are quasi-two-dimensional and nondivergent, we may be able to interpret this phenomenon in light of two-dimensional turbulence theory. The eddy kinetic energy in the middle latitudes owes its major energy source to the disturbances of wavenumbers, say, 6 to 8 generated by baroclinic instability. The input energy is then cascading toward longer waves and vorticity is cascading toward shorter waves (Leith, 1968). The mesh with higher horizontal grid resolution is capable of resolving baroclinic waves more accurately and produces less energy dissipation (due to nonlinear viscosity) at the wavenumbers of maximum baroclinic instability (Merilees, 1975). Thus, more energy may be transported to the longer waves accounting for a significant increase in kinetic energy in the longwave regime of wavenumbers 1 to 4 for the 2.5° mesh case.
Figure 2.26 shows the distribution of specific kinetic energy as a function of zonal wavenumber based on data from a January simulation with the 5° 12-layer model (Kasahara et al., 1973). In the N.H. middle stratosphere (31.5 km), the eddy kinetic energy decreases sharply as the wavenumber increases, except in the tropics where the kinetic energy does not depend much on wavenumber. In the S.H., the eddy kinetic energy is generally very small compared with that in the N.H. because the eddy activity in the S.H. is grossly underestimated in the simulation. In the N.H. lower stratosphere and mid-troposphere, the computed magnitudes of eddy kinetic energy are smaller than the observed January and February mean values in the low wavenumbers. This deficiency in computed eddy kinetic energy may be caused by truncation errors of the finite-difference scheme in the 5° model as discussed earlier.

2.4.2 Wavenumber-frequency spectra

To identify the nature of planetary-scale disturbances in simulated data, Tsay (1974) analyzed wavenumber-frequency spectra using output from the 5° 12-layer model. The wavenumber-frequency spectra for the zonal wind component at the equator and 25.5 km show two strong peaks at wavenumber 1, eastward-moving with a period of about 24 days and wavenumbers 2-3, eastward-moving with a period of about 12 days. These waves strongly resemble atmospheric Kelvin waves detected in the tropical stratosphere (Wallace
The spectra of the meridional wind component at the equator and 22.5 km indicate a distinct peak for a westward-moving wave with wavenumber 3 and a period of about 4.61 days. These waves strongly resemble the mixed Rossby-gravity waves described by Rosenthal (1965), Matsuno (1966) and Lindzen (1967). The spectra of pressure at the equator at 3 and 24 km have a distinct peak with wavenumber 1 and a period of about 5 days. A similar peak appears at all model levels and in the middle latitudes. These waves are identified as free oscillations of the second kind, similar to Rossby-Haurwitz waves (Longuet-Higgins, 1968), and are observed in the atmosphere (Madden and Julian, 1972).

2.5 Geographical Distributions of Meteorological Variables

Distributions of meteorological variables in latitude-height meridional cross-sections are convenient ways to represent the vertical structure of the atmospheric general circulation, but give no information on the longitudinal variations of variables. We show here the geographical distributions (averaged for 30 days) of cloudiness, precipitation, surface temperature and sea-level pressure for January simulated by the 5° 6-layer model (Kasahara and Washington, 1971).

Figure 2.27 presents the calculated cloudiness distribution at 3 km—(a) no mountain case and (b) mountain case. These
patterns are compared with the observed total ocean cloudiness for December-February after McDonald (1938) in Fig. 2.28. Also from TIROS satellite nephanalyses, Clapp (1964) has drawn a total cloudiness distribution for December-February (Fig. 2.29). The large-scale features in the computed cloudiness pattern agree in general with the observed distribution (see also Section 2.2). The high cloudiness poleward of 40° in both hemispheres has been modeled correctly. We notice some tendency for less cloudiness over land than ocean in the winter hemisphere and more cloudiness in the summer hemisphere. Over the Sahara Desert, the cloudiness is small, as expected. Some failures in the simulation of cloudiness are noted off the West Coast of South America and the Western African Coast in the Atlantic. These coastal areas experience low stratus clouds not predicted by the present cloudiness parameterization.

Figure 2.30 shows the computed mean precipitation distribution [mm (90 days)^{-1}] which may be compared with the observed pattern in the same units after Möller (1951) for December-February shown in Fig. 2.31. The major dry and wet areas are relatively well simulated. For example, the major wet areas on the computed distribution coincide with the intertropical convergence zones over the Pacific, Atlantic and Indian Oceans. The major dry areas are found over the Eastern Pacific along the West
Coast of South America, the Southern Atlantic west of the African Coast, the Southern Indian Ocean west of Australia and over the Sahara Desert. The computed distribution, however, gives a higher precipitation rate than observed, particularly in the tropics. In the model calculation, we did not allow for any evaporation from falling raindrops. Also, the model gives a higher rate of evaporation from the earth's surface and more condensation heating than observed, although the evaporation rate minus the precipitation rate in the model agrees with observations, as discussed in Section 2.2.

Figure 2.32 shows (a) the computed mean surface temperature over each hemisphere, and (b) the observed distributions of surface mean ambient temperature for January from Crutcher and Meserve (1970) for the N.H. and Taljaard et al. (1969) for the S.H. Agreement is good over the North American continent, but the cold air pool in eastern Siberia is not well simulated. The computed pattern shows clearly the presence of cold air over the Himalayan Plateau, but the observed does not clearly show such an effect.

In Fig. 2.33 (a) is the computed mean sea-level pressure over each hemisphere and (b) is an observed distribution for January. One serious problem in evaluating sea-level pressure is that its evaluation is subject to an assumption of temperature distribution from ground height to sea level. In the model, we assume a constant
temperature in the mountains which is the same as the mean temperature in the layer directly above the mountains. This procedure tends to make the sea-level pressure higher than that obtained by the usual method of assuming a temperature profile inside the mountains. The Icelandic Low is too strong, whereas the intensity of the Aleutian Low is greatly underestimated. For the S.H., the locations of the subtropical highs are fairly well simulated, but the model fails to properly simulate the low-pressure belt around 55°S. It is suspected that the numerical handling of Antarctica is not satisfactory and we need further investigation of the method of incorporating the earth's orography in the model (see Section 3.5.3).
Fig. 2.1 Time variation of global mean $\overline{K}$ and $K'$ for mountain case (M) and no-mountain case (NM) (January simulation).
Fig. 2.2 Time variation of global mean $\bar{K}$ and $K'$ for Cases 138, 156 and 331 of the 5° model and a 2.5° model case (January simulation).
Fig. 2.3 Time variation of global mean $\bar{K}$ and $K'$ for control, ice-age and snow-cover cases (July simulation).
Fig. 2.4 Time variation of global mean $\bar{K}$, $\bar{I}$ and $K'$ for January simulation with 5° 12-layer tropospheric-stratospheric model with the earth's orography.
Fig. 2.5 Time variation of global mean $K'$, $\bar{K}$ and $\bar{I}$ for an 18-month period of a seasonal run with $5^\circ$ 12-layer tropospheric-stratospheric model with the earth's orography.
Fig. 2.6 Latitude-height distributions of $\bar{K}$ for three mesh cases (January simulation). Lower right shows observed $\bar{K}$ for January based on geostrophic wind (N.H. from Jenne et al., 1971, and S.H. calculated from Crutcher and Meserve, 1970).
Fig. 2.7 Latitude-height distributions of $K'$ for three mesh cases. Lower right shows observed $K'$ for January after Oort and Rasmusson (1971).
Fig. 2.8 Latitude-height distributions of zonally averaged fields: (a) zonal mean temperature for Case NM (heavy lines) and Case M (thin lines); (b) observed temperature for January; (c) zonal mean specific humidity for Case NM (heavy lines) and Case M (thin lines); and (d) observed specific humidity for January (Peixoto and Crisi, 1965).
Fig. 2.9 Latitude-height distributions of zonally averaged fields: (a) zonal wind component for Case NM (heavy lines) and Case M (thin lines); (b) observed zonal wind component for January; (c) meridional wind component for Case NM (heavy lines) and Case M (thin lines); (d) observed meridional wind component for December-February (Vincent, 1969); (e) vertical wind component for Case NM (heavy lines) and Case M (thin lines); (f) observed vertical wind component for December-February (Vincent, 1969).
Fig. 2.9 Continued
Fig. 2.10 Left, from top down: zonally averaged mean temperature of Case NM, Case M and observed long-term mean. Right: zonally averaged mean west-to-east velocity component of Case NM, M and observed long-term mean.
Fig. 2.11 Left, from top down: zonally averaged meridional velocity of Case NM, M and observed long-term mean. Right: same as left but for vertical velocity.
Fig. 2.12 Latitudinal distributions of vertical integrals of total heating rate $\rho Q$. Solid line is observed for December-February (Newell et al., 1969); dashed line and dots are computed for Cases NM and M; open squares are observed (Katayama, 1962).
Latitudinal distributions of the difference between evaporation and precipitation computed from the model for Case NM (dashed line) and Case M (dots). Also, the same quantity computed indirectly from moisture balance is shown for Case NM (squares) and Case M (crosses). Solid line denotes observed for December-February.
Fig. 2.14 Latitudinal distributions of zonal averages of total cloudiness. Observed data from Telegadas and London (1954) for the N.H. and van Loon (1972) for the S.H.
Fig. 2.15 Latitude-height distributions of eddy momentum transports: (a) observed total (transient plus standing) eddy momentum transport for December-February; (b) observed transient eddy momentum transport for December-February; and (c) total eddy momentum transport for Case NM (January simulation).
Fig. 2.16 Latitudinal distributions of vertical integrals of $\rho \sigma \hat{u} \hat{v} \alpha \cos \phi \Delta z$ and $\sum \rho \sigma u'v' \alpha \cos \phi \Delta z$ for Cases NM and M and observed from Oort and Rasmusson (1971) (January simulation).
Fig. 2.17 Vertical integrals of $\overline{\rho \sigma v'q'}$ as function of latitude (January simulation).
Fig. 2.18 Latitudinal distributions of vertical integral of $C_{vpT}$ for Case NM (solid line) and Case M (dots). Dashed line shows observed for January from Oort and Rasmusson (1971).
Fig. 2.19 Similar to Fig. 2.18, but for vertical integrals of $\frac{C_p \sigma v' T'}{\Delta z}$ as a function of latitude (January simulation).
Fig. 2.20 Vertical integrals of $\overline{\rho \sigma \hat{u} \hat{v}} \cos \phi \Delta z$ as a function of latitude for July cases.
Fig. 2.21 Vertical integrals of $\rho \underline{u} \underline{v} \cos \phi \Delta z$ as a function of latitude for July cases.
Vertical integrals of $\bar{\sigma} \bar{v} \bar{q}$ (upper) and $\bar{\rho} \bar{\sigma} \bar{v} \bar{q}'$ (lower) as a function of latitude (July simulation).

**Fig. 2.22**
Fig. 2.23 Left, from top down: zonally averaged mean cross-sections of $u'v'$ for Case NM, M and observed. Right: same as left, but for $T'v'$ (January simulation).
Fig. 2.24 Latitudinal distributions of component terms in zonal internal energy budget equation integrated from 18 to 36 km. January simulation with the 5° 12-layer model.
Fig. 2.25 Spectra of kinetic energy KE vs. wavenumber k for three mesh cases compared with observed (lower right). Straight line shows a -3 power distribution.
Fig. 2.26 Distribution of specific kinetic energy vs. wavenumber at 31.5, 19.5 and 4.5 km and at 60°N, 30°N, equator, 30°S and 60°S. Solid lines denote Case M and dashed lines Case NM; observed (Teweles, 1963) by dots for January 1958 and crosses for February 1958.
Fig. 2.27 Computed January mean cloudiness at 3 km. (a) No-mountain case and (b) mountain case. Contour interval in tenths.
Fig. 2.28 Observed total cloudiness for December-February (McDonald, 1938). Contour interval in tenths.
Observed total cloudiness from satellite observations (Clapp, 1964). Stippled shading represents cloudiness > 75%; hatched area 50-75%; no shading < 50%. Solid lines are isopleths of 25% within white area, while dashed lines are either 65 or 35%.

Fig. 2.29
Fig. 2.30 Computed January mean distribution of precipitation. (a) no-mountain case, (b) mountain case. Contour interval is mm (90 days)⁻¹.
Fig. 2.31 Observed mean distribution of precipitation (mm per season) for December-February (Möller, 1951). (A season is assumed to be 90 days.)
Fig. 2.32 (a) Computed geographical distributions of surface temperature (contour interval 5°C) for January.
Fig. 2.32 (b) Observed mean distributions of surface ambient temperature for January.
Fig. 2.33 (a) Computed geographical distributions of sea-level pressure (contour interval 4 mb) for January.
Fig. 2.33 (b) Observed mean distributions of sea-level pressure for January.
3.0 Model Sensitivity for Climate Simulation

We summarize here the model aspects relating to climate simulation and discuss what has been learned from the climate experiments carried out thus far.

3.1 Finite-Difference Grid Resolution

3.1.1 Effects of horizontal resolution

Wellick et al. (1971) investigated the effect of varying the horizontal grid resolution in the GCM on simulation of climatological statistics. With the 6-layer model, three different versions with horizontal grid increments of 10°, 5° and 2.5° in longitude and latitude (centered difference operators are calculated over twice the grid length) were run and the results compared with observed climatological statistics. For the simulation of the wind field, the 10° mesh is definitely too coarse; the baroclinic eddies are almost absent. Although the 5° version gives reasonable zonal mean states, we need the 2.5° version to reproduce more accurately higher-moment quantities such as eddy kinetic energy and eddy transports of momentum and heat (Section 2.1-2.4). We probably need a grid resolution of the order of 1° for satisfactory simulation of the baroclinic waves with a second-order, finite-difference scheme.

3.1.2 Effects of vertical resolution

Systematic tests are being conducted with a short-range
prediction version of the GCM to determine the minimum number of vertical levels necessary for realistic atmospheric forecasting (Section 4). However, for climate simulation we have compared the 2-layer model ($\Delta z = 6$ km) (Washington and Kasahara, 1970) with the 6-layer model ($\Delta z = 3$ km) as reported in Kasahara and Washington (1971). Essentially all aspects of the simulation improved with increase in vertical resolution.

We are trying to find the best method of choosing the vertical distribution of grid points. It appears that we need high vertical resolution near the earth's surface and the jet stream level (near the tropopause) because strong vertical gradients exist in these regions. The answer may also depend partly on the choice of vertical coordinate.

3.2 Height of Model Top and Upper Boundary Conditions

It has been suggested that the simple upper boundary condition that vertical motion $w = 0$ at the model top causes distortion of wave modes and, therefore, this upper boundary condition may contribute to errors in weather prediction and climate simulation. One effect of this boundary condition may be seen from the performance of the 5° 12-layer model in which the 6 lower layers correspond exactly to the 5° 6-layer model of the troposphere. In the 12-layer model, the top boundary condition $w = 0$ is at 36 km and in the 6-layer model 18 km. From analyses of wave
characteristics in the tropics, we found that the 12-layer model simulates atmospheric Kelvin waves and mixed Rossby-gravity waves, both of which carry energy upward in accordance with observations. For more details, see Kasahara et al. (1973) and Tsay (1974). With regard to the vertical propagation of planetary waves, the 12-layer model is free from the artificial constraint of $w = 0$ at 18 km. In comparing the tropospheric circulation, we did not find improvement with the 12-layer model compared with the 6-layer model, except that the upper tropospheric jet streams have closed-off contours.

For forced oscillations in a simple analytical model, the upper boundary condition may be selected by applying the so-called radiation condition. For free oscillations, however, the radiation condition depends on the frequency of wave motions. Since the radiation condition applied to inviscid and adiabatic motions is designed to simulate the absorption of upward-propagating waves in the upper atmosphere due to molecular viscosity and conductivity, one can introduce a small dissipation (Bretherton, 1969) in the upper boundary condition. Upward-propagating waves may also be absorbed at a critical level where the mean horizontal velocity is equal to the horizontal phase velocity and momentum is transferred to the mean flow (Booker and Bretherton, 1967). Thus, in a realistic numerical model where the mean horizontal velocity depends on height and effects of viscosity and conductivity are included at a sufficiently high level
of the atmosphere, the selection of the upper boundary condition may not be as crucial as in a simple inviscid atmospheric model, as far as the simulation of the troposphere.

3.3 Computer Precision

The sensitivity of the GCM to arithmetic precision has been tested for short- and long-term simulations. The results of the short-range forecast error growth are discussed in Section 4. For long-term simulations with results averaged over the last 30 days, Williamson and Washington (1973) found few differences between a 48-bit mantissa calculation and 24- and 21-bit runs. An error growth experiment with a small random initial difference was performed to determine the significance of the results. The differences in all experiments were quite small in the latitude-height distribution of wind components.

3.4 Sea-Level Pressure Adjustment and High Wavenumber Patterns in the Pressure Field

The second-generation model uses first-order difference approximations near domains blocked by orography and linear interpolation in the vicinity of the poles to fill values at grid points which are skipped to avoid computational instability. The model therefore has only first-order accuracy in these areas and neither procedure conserves mass. Figures 3.1 and 3.2 show zonally averaged pressure as a function of time at sea level (extrapolated) and \( P_3 \) (9 km).
for the $5^\circ$ 6-layer experiment. The arrow at the right of the figures indicates a 10% change; the changes are almost uniform with latitude and height. We have modified the second-generation version to maintain the globally averaged value of surface pressure every time step. The procedure eliminates the mass change problem while affecting very little climatological statistics such as zonal wind, temperature, moisture etc.

Another peculiar aspect of surface pressure in the second-generation model is a $2\Delta\lambda$ oscillation of about 5 mb in the unsmoothed output. This oscillation is most evident in the tropics over the oceans. It is caused by the separation of grid lattice due to a centered difference scheme and it does not interact with the wind and temperature fields. It can be suppressed by Fourier chopping of the pressure field.

3.5 Parameterization of Physical Processes

3.5.1 Radiation parameterization

Several radiation sensitivity experiments have been carried out (Washington, 1971b) showing that a small random error in the radiation calculation has virtually no effect on climate simulation. This may not be the case with systematic changes in the radiation calculation. The Fels and Kaplan (1975) test of a detailed infrared radiation calculation in the $5^\circ$ model led to much colder upper tropospheric temperatures in the tropics than does Sasamori's (1968a,b)
simplified method. Neither case includes the absorption of solar radiation by ozone. The difference between the two cases is caused by a small but systematic difference in the cooling rates of the two methods which interacts eventually with other physics as well as the dynamics. In Figs. 3.3 and 3.4, we show the zonally averaged cooling rates at various latitudes between the test (solid line) using the refined method and the control (dashed) using Sasamori's method. The first figure is near the initial time (within a half day) and the second is 39½ days later. Note that the systematic difference in cooling rates is largest in the upper tropical troposphere and the differences decrease with time. The temperature structure in Fig. 3.5 has changed dramatically by becoming cooler (~ 25° in the tropics) with the Fels-Kaplan method. This also affects other atmospheric statistics such as $\bar{K}$, $K'$, $\overline{u}$ etc. (see Fels and Kaplan, 1975, for details). Note that the Sasamori method takes two orders of magnitude less computer time than the Fels-Kaplan method. The Fels-Kaplan study shows that systematic differences in infrared radiation calculations have produced seemingly dramatic temperature changes, but the effects of changes in comparison with the natural variability of the atmosphere have not been fully explored. Also, we do not yet know the degree of feedback between dynamics and atmospheric radiation in the presence of clouds. Attempts to refine radiation calculations in GCMs should be accompanied by sensitivity studies to determine the degree of their impact on the atmosphere and by investigations of the influence of cloudiness on radiation calculations.
3.5.2 Cloudiness--radiation--ocean-surface sensitivity studies

Since the NCAR GCM adopts fixed ocean-surface temperatures (OST), the response of the GCM surface temperature to changes in the radiation balance is restricted to grid points in non-ocean areas. This necessarily reduces the sensitivity of the global (or zonal) mean surface temperature to changes in radiation. However, even with this restriction several aspects of cloudiness--radiation--ocean-surface feedback have been studied.

(1) Response of globally averaged surface temperature to deliberately imposed increases or decreases in cloudiness: We first computed a January control case (29-day average) having a global low cloud amount (LCA) of 0.501 and global surface temperature ($T_g$) of 11.21°C. The control is compared to two cloud perturbation cases—one where LCA is increased to 0.610 and $T_g$ becomes 10.65°C and the other where LCA decreases to 0.376 and $T_g$ becomes 11.56°C.

(2) Response of cloud amount to deliberate changes in OST: Two experiments were performed with global $\pm 2^\circ$C changes in OST and then compared to a January control case and a random experiment—a control case plus small random perturbation in initial conditions. Some results and differences between the control and other experiments are summarized below.
GCM EXPERIMENTS \( \Delta T_G(\degree C) \) \( \Delta T_1(\degree C) \) \( \Delta q(g/kg) \) \( \Delta RH(\%) \) \( \Delta LCA(\%) \) \( \Delta OST(\degree C) \)

<table>
<thead>
<tr>
<th></th>
<th>( \Delta T_G(\degree C) )</th>
<th>( \Delta T_1(\degree C) )</th>
<th>( \Delta q(g/kg) )</th>
<th>( \Delta RH(%) )</th>
<th>( \Delta LCA(%) )</th>
<th>( \Delta OST(\degree C) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Higher OST--control</td>
<td>+2.29</td>
<td>+2.40</td>
<td>+0.94</td>
<td>-3.3</td>
<td>-2.74</td>
<td>+2.0</td>
</tr>
<tr>
<td>Lower OST--control</td>
<td>-2.52</td>
<td>-2.77</td>
<td>-0.87</td>
<td>+1.7</td>
<td>+1.53</td>
<td>-2.0</td>
</tr>
<tr>
<td>Random--control</td>
<td>-0.27</td>
<td>-0.36</td>
<td>-</td>
<td>-</td>
<td>-0.66</td>
<td>-</td>
</tr>
</tbody>
</table>

where \( T_1 \), \( q \) and \( RH \) denote the temperature, specific humidity and relative humidity at 1.5 km.

The above indicates that the increased OST experiment leads to increased stability and specific humidity but decreased relative humidity and cloudiness, whereas the decreased OST experiment produces changes in these variables of opposite sign and comparable magnitude. Comparisons with a random run suggest that these changes are significant for the 29-day average model experiment. These experiments show a positive feedback between \( T_G \) and model-generated LCA.

Experiments in which OST is increased by 2\( \degree C \) for latitude belts 0\( \degree \)-10\( \degree \)S and 10\( \degree \)-20\( \degree \)N suggest that regional changes in OST produce a response in cloudiness different from that produced by global changes in OST (Fig. 3.6). In addition, further experiments reveal that the sensitivity of low cloudiness to changes in \( T_G \) is relatively unaffected by changes in the drag (dimensionless) coefficient for evaporation, although absolute values of humidity, precipitation etc. are significantly different. Results are being analyzed by Schneider and Washington and the work will be completed by October 1975.
3.5.3 Effects of earth's orography

There is no question about the importance of the earth's orography in numerical models for weather forecasting and climate simulation. It is, nevertheless, difficult to define precisely the role of orography, which acts as a barrier to the atmospheric circulation and affects the geographical distribution of sensible and latent heating over the globe as well. Another difficulty comes from the many technical problems in the incorporation of realistic mountain effects in models. Kasahara (1966, 1967), Kasahara and Washington (1969, 1971) and Kasahara et al. (1973) have investigated the various aspects of the influence of the earth's orography on atmospheric motions.

In determining the global-scale features of tropospheric climate, the thermal effect of continentality dominates the dynamic effect of orography. However, the detailed climatic features are significantly different with and without orography. For stratospheric simulations, the semipermanent anticyclone over the Aleutian region in the lower stratosphere during winter was simulated only in the model with realistic orography (see Fig. 3.7). The improvement with orography is due to increased amplitude of wavenumber 1 in the stratosphere as a result of the increased vertical transport of planetary wave energy. For more detail, see Kasahara et al. (1973).
In the troposphere, the dynamic effects of the Himalayas and Rockies are evident, although their significance may be underestimated in the $5^\circ$ mesh models. We need a high-resolution model to evaluate fully the impact of orography on regional climate. For example, studies are underway by Washington to estimate the effects of the Himalayas and East African highlands on the Northern Hemisphere summer monsoon.

We mentioned that the method of incorporating mountains in the second-generation model used first-order finite differences near the mountains which do not conserve mass or pressure. With a transformed vertical coordinate in the third-generation model, there is no inherent problem of conservation of mass or pressure; however, the truncation errors associated with the pressure gradient terms in the horizontal momentum equation may have a large effect on the evolution of flow near the mountains. The proper numerical treatment of mountains remains a largely unsolved problem in atmospheric modeling.

3.5.4 Convective parameterization

Two methods of convective parameterization have been tested in the second-generation model—the simple convective adjustment described by Washington and Kasahara (1970) and a scheme devised by Krishnamurti and Moxim (1971), based on Kuo (1965). The first method makes use of two assumptions: (1) the adjustment follows
the moist adiabat for upward motion and the dry adiabat for downward motion, and (2) the internal energy is conserved. The temperature change caused by latent heat released in the lower atmosphere is shifted to the upper troposphere by the adjustment procedure. In the Krishnamurti-Moxim method, the moisture convergence in the lower troposphere is assumed to release latent heat in proportion to the difference between the moist adiabat and temperature distribution. Thus, more heat is released in the upper troposphere. The two methods lead to moderate differences in statistics, but it is difficult to conclude which is better overall (see Washington and Baumhefner, 1974; Washington and Daggupaty, 1975). For example, the Krishnamurti-Moxim scheme leads to tropical lapse rates closer to observed than the convective adjustment, but the rainfall rates are smaller than observed. Also, the overall temperatures in the tropics are colder than normal. To correct these deficiencies in their scheme, several proposals were made as described in the above references.

3.5.5 Drag coefficient

Prior to examining the effect of mountain wave momentum flux in the NCAR GCM, Chervin and Lilly performed a series of sensitivity tests with the 5° model to determine the effect of the surface drag coefficient on momentum transport and upper-level winds. When the drag coefficient was increased by a factor of two at all grid points, the lower-level winds decreased as expected, but an increase
in the upper-level winds was also observed. This resulted from an increase in the upper-level poleward temperature gradient caused by increased upward latent heat flux in the tropics and the resulting convective adjustment in the upper levels. The opposite effects were observed when the drag coefficient was reduced by a factor of two. However, these results were very much dependent upon the minimum surface velocity of 5 m/s prescribed by the GCM in the lower boundary condition for the flux terms. Since this minimum was, in fact, imposed at a majority of the grid points, the stress components obeyed a linear drag law as opposed to the expected quadratic dependence on velocity. By the same reasoning, the fluxes of sensible heat and water vapor were relatively insensitive to changes in the surface velocity. See Section 3.6.4 for the plan to improve the formulation of the planetary boundary layer. It is expected that Chervin and Lilly will complete this work by September 1975.

3.5.6 Clear-air turbulence

It is known that $\bar{K}$ in the GCMs is too large, partly because the present GCMs do not properly take into account internal dissipation such as by clear-air turbulence. In a joint project with Pennsylvania State University, we investigated the effect of increased vertical eddy viscosity as a function of the Richardson number (see Schenk, 1974, for
experiment details). The experiments exhibited some puzzling features; for example, we found an unexplained increase in $K$ after an initial expected decrease. Because of time restraints on the publication of a thesis, this work awaits further study by the Pennsylvania State group.

3.5.7 Wave momentum flux parameterization

Chervin and Lilly followed up on earlier work of Lilly (1972) in which a constant drag force was inserted into the GCM over a region coinciding approximately with the Rocky Mountains. In this latest effort, drag forces were added at high and low levels of the troposphere in separate experiments proportional to the square of a low-level velocity. This crude attempt at a parameterization demonstrated that the reduction in zonal wind was nearly the same in the high- and low-level drag experiments.

3.5.8 Horizontal diffusion

Two forms of horizontal diffusion were tested in the second-generation model. One, developed by Smagorinsky (1965), treats the horizontal eddy viscosity as a function of the deformation of flow; the other, suggested by Leith (1969), has horizontal eddy viscosity proportional to the gradient of vorticity. Leith's formula has a higher scale dependence than Smagorinsky's, but for the particular finite-difference approximations used in the second-generation model, Leith's
did not work as well. The model needed more total diffusion to keep the calculation from blowing up by nonlinear instability. The finite-difference equations in the second-generation models are such that the deformation becomes very large but the gradient of vorticity remains small when $2\Delta x$ waves are generated. Also, compared with the deformation, the vorticity may be a poor measure of gravity waves generated in the tropics by latent heat release. We concluded that the details of the finite-difference scheme are as important as the form of horizontal diffusion.

Experiments have been carried out with the third-generation model (fourth-order accuracy in finite differencing) to determine if Fourier chopping can control small-scale disturbances without significantly damping the large-scale baroclinic waves (Williamson, 1974a). The procedure does not work satisfactorily for the low-resolution (~ 5.6°) grid. The buildup of excessive energy in the shortest waves not filtered out resulted in overly large poleward eddy transports with an accumulation of moisture and momentum in the polar regions. The polar atmosphere becomes saturated resulting in unrealistic condensation and associated warming there. It should be noted that with the tests carried out so far there is also too much kinetic energy in the high wavenumbers compared with observed. Tests are underway with other nonlinear diffusion
operators similar to that proposed by Kreiss and Oliger (1973) and the deformation formulation used in the second-generation model.

Leith (1971) has suggested another approach to handle the nonlinear diffusion in a prediction model. The idea is to introduce an eddy-dissipation function that exactly preserves the sharply truncated -3 power spectrum by compensating for missing nonlinear interactions as computed by the eddy-damped Markovian approximation. This method has not yet been tried.

3.6 Lower Boundary Conditions

3.6.1 Ocean-surface temperature (OST)--Pacific, Atlantic and Indian Oceans

We are conducting several experiments on OST anomalies. The first series with OST changes, carried out by Houghton et al., (1974), tested the effect of a 1-2°C temperature anomaly in the western North Atlantic with the 5° model. The differences between the control and anomaly experiment were small and difficult to evaluate quantitatively. The results with a 2.5° model are presently being studied jointly with the University of Wisconsin. In this new set of experiments, we have added a random experiment so that an estimate of noise level can be included in the study.

In a similar, but distinct set of experiments, we are jointly studying with Scripps Institution of Oceanography the effects of OST anomalies in the North Pacific. These experiments indicate that an anomaly pattern of -4°C in the Western Pacific and +4°C in
the Eastern Pacific has relatively little effect on the hemispheric temperature patterns except near the area of the anomaly itself. In fact, most changes were over the warm anomaly and confined to the lowest model layers.

To measure significance, we first completed a calculation of standard deviation of temperature from 5 computed Januaries and compared the differences among a random-control, anomaly-control and superanomaly-control. In the superanomaly experiment, we multiplied the normal anomaly by a factor of 3. From the analysis by Chervin et al. (1975), we have an idea of how much time-averaging is required to separate the signal from the natural variability. Figures 3.8-3.11 show the differences in 30-day means of temperature at 1.5 km between the three experiments (random, normal anomaly and superanomaly) and the control case. Figure 3.8 is averaged from Days 31-60, Fig. 3.9 from 61-90 and Fig. 3.10 from 91-120. If we average over 90 days, as in Fig. 3.11, the major effect of the superanomaly shows up in the Eastern North Pacific; we also find that the normal anomaly experiment has nearly the same response as the random experiment. Although we have not completed analysis of these experiments, we find little evidence of definite downstream effects from the anomalies.
3.6.2 Surface albedo

Two sets of GCM experiments have been conducted indicating a large sensitivity to surface albedo, especially in the Northern Hemisphere summer. In a seasonal run with the stratospheric version of the second-generation model, an 18-month calculation was performed wherein the sun's declination and ocean temperature were changed but the ice-snow line was held fixed (inadvertently!) to a mean January position. The significance of the ice-snow line is that this fixes an upper limit of 0°C for the surface temperature. In this run, the higher latitudes in the Northern Hemisphere summer remained too cold compared to the observed. To examine this aspect more closely, we changed the ice-snow line in the mean July position, and the simulated July atmospheric patterns were closer to the observed.

In another set of experiments on the response of the simulated atmospheric circulation to glacial boundary conditions and extensive snow cover, Williams (1975) showed that the effect of snow cover on the Northern Hemisphere is more important than the dynamic effect of glacier height. Bearing in mind that the changes in albedo are quite drastic in both experiments, it is difficult to judge if smaller, more subtle changes of albedo will have noticeable effect on the atmospheric simulation. See also Section 3.6.5.
3.6.3 Surface hydrology

In the early versions of the second-generation model, we assumed saturated ground over all non-ocean regions and a Bowen ratio of unity. This method generated unrealistic precipitation over dry parts of the world such as the Sahara Desert. In later experiments with this version, the Bowen ratio was reduced to 1/10, thus, in turn, reducing substantially the precipitation over desert areas.

The parameterization of soil moisture and snow cover (Washington, 1974) distinguishes between liquid and solid precipitation at ground level and computes the melting of snow and evaporation as a function of soil moisture. It contains the major processes leading to the formation of deserts and tropical rain forests. The new surface hydrology parameterization showed more realistic precipitation patterns over the desert areas. Used with the convective adjustment, the new surface hydrology parameterization led to below-normal rainfall rates over the tropical land areas and above-normal amounts over oceans. With the Krishnamurti-Moxim convective parameterization scheme, however, it led to more realistic rainfall over land areas. See also the precipitation patterns in Washington and Daggupaty (1975).
3.6.4 Planetary boundary layer

The current NCAR GCMs adopt a bulk transfer coefficient method to calculate the vertical transfer of sensible and latent heat between the atmosphere and the earth's surface and the momentum dissipation at the earth's surface.

To improve parameterizations of the planetary boundary layer (PBL) in the GCM, J. Deardorff and R. Benoit developed a method involving prediction of the thickness of the PBL in terms of large-scale variables such as vertical velocity and potential temperature. The thickness of the PBL, a fundamental quantity in the PBL parameterization, changes with time from day to night and varies horizontally as well and depends not only on the amount of surface heat and moisture fluxes, but also on cloud population.

Prediction equations are formulated for wind velocity, temperature and moisture at the "anemometer" level near the earth's surface. According to Deardorff (1972) and others, the variables at the anemometer level can be estimated better if calculations are based on the height of the PBL and the surface fluxes of heat, moisture and momentum. This requires substantial modification of the GCM code, now underway.

3.6.5 Ice-age boundary conditions

Using the second-generation model, Williams et al. (1974) investigated the influence of glacial-period boundary conditions
and extensive July snow cover. In the ice-age cases, the results show temperatures in July more like January of today: Icelandic low shifted southward, more meridional pressure gradient, intensified Hadley cell, greater transient cyclones over the Atlantic, greatly reduced global mean eddy kinetic energy in July reflecting absence of monsoon, and reduced precipitation. In the July snow cover case, which had albedos and ocean surface temperatures like the ice-age case but with present-day orography, we find a weakened Northern Hemisphere summer monsoon, intensified troughs over Eastern North America and Western Europe, increased pole-to-equator temperature gradient, jet stream (Northern Hemisphere) stronger and farther south and increased cloudiness over snow-covered areas.

3.6.6 Surface energy balance ocean model

As a simple ocean model to predict the ocean-surface temperature (OST), we constructed a 3-m thick "swamp" model with an insulated bottom and no oceanic currents; the swamp temperature is calculated based on the local surface energy balance. This model is designed to investigate the positive feedback between cloudiness amount and surface temperature suggested by sensitivity experiments. Several cases have been run and demonstrate such a positive feedback. Furthermore, in cases where the model clouds are opaque to solar radiation, but not allowed to exceed 80% at a grid point, this positive feedback leads to unrealistic ocean
temperatures (i.e., ocean temperature reaches below 0°C), but if
the clouds are permitted to diffusely transmit to the surface 50%
of the solar radiation reaching them, unrealistically cold ocean
temperatures do not appear in a 90-day simulation experiment.

For the perpetual January experiment, it is not possible, of
course, to obtain the calculated OST as it is observed. The object
of this study is not to reproduce the correct OST, but to determine
the magnitude of imbalance of various energy fluxes at the earth's
surface through the calculation of OST. The time change of OST is
dependent on the residual of various energy fluxes at the surface,
some having large absolute magnitudes. Thus, a study of this kind
is useful to examine the local energy imbalance for some climate
studies. Analysis of these experiments will be completed by October
1975 by Schneider and Washington.

3.6.7 Thermal pollution studies

Washington (1971a, 1972) conducted several preliminary
experiments at the request of Oak Ridge National Laboratories to
determine the possible effects of thermal pollution on the climate.
The NCAR GCM is not a full climate model because ocean tempera-
tures are held fixed; thus, the experiments were not intended to
obtain definitive answers on changes in climate. The historical
importance of these experiments is that they focus attention on
the problem of measuring significance of experiments when the
effect is relatively small. By assuming per capita energy use and expected population growth to the year 2000, we added to the surface energy balance equation the amount of thermal pollution proportional to the present population density. This amount of thermal energy led to temperature increases in the tropical land areas of order of 1°C and several degrees in the north pole area. These experiments, of course, must be repeated with a coupled atmosphere-ocean model to measure the response to additional energy input without the fixed ocean temperature constraint.

3.7 Transport Characteristics--Ozone and Particle Diffusion

The second-generation stratospheric model has been used by London and Park (1973, 1974) to study the transport and photochemistry of ozone. They also performed experiments on the possible contamination of the stratosphere by future supersonic transport fleets. The transported ozone was constrained not to interact with the radiation field. They found that ozone was transported poleward out of the equatorial stratosphere and then downward in both hemispheres. The principal mechanisms were the Hadley circulation in the tropics and large-scale eddies in the mid-latitudes similar to the transports of momentum and sensible heat.

Because of the lack of observed data readily available on the transports of ozone in the stratosphere, a direct comparison of the
computed ozone transports with observation is difficult. The general pattern of the computed ozone transports is consistent with what is presently known, although the absolute values are not (London and Park, 1973). For example, the model produced too much ozone in the Southern Hemisphere subtropical stratosphere and transported too much ozone into the Northern Hemisphere subtropical troposphere. One computational problem noted in this study is that the ozone budget was not well maintained, even accounting for sources and sinks, suggesting that future experiments with trace constituents should use a special numerical scheme to assure high accuracy.

In a second series of experiments, London and Park perturbed the oxygen-hydrogen-nitrogen photochemical calculation of ozone from supersonic transport exhaust emissions. These experiments are finished, but the analyses of the results have not been completed.

In the transport of particles (Kao and Chiu, 1975), clusters of marked air particles were inserted into the model-generated wind fields at various heights and latitudes. The three-dimensional trajectories of these particles were followed for 15 days. The analyses included computations of auto- and cross-correlations of large-scale velocity of particles, rate of diffusion, mean trajectories of particles and the ensemble averages of the square of particle distance.
3.8 Time-Averaging of GCM Statistics (Global, Zonal and Geographic)

Chervin et al. (1974) examined the time-averaged response of the GCMs to random perturbations in the initial conditions while leaving all boundary conditions fixed. The rms difference between the perturbed and unperturbed control cases—noise level—showed a nearly monotonic decrease with increasing averaging interval. However, in view of the smallness of many observed climatic changes in relation to the 30-day noise levels computed from the GCM, time averaging over longer periods may be necessary for certain variables. For example, in Figs. 3.12 and 3.13, the global rms for NCAR and UCLA-Rand models of the zonal wind component and precipitation starts to level out at 5-10 days. A more complete survey has been made of sufficient time-averaging intervals for atmospheric variables in terms of global, zonal and geographical averages (Chervin and Schneider, 1975a,b; Chervin et al., 1975). This study is crucial for the proper interpretation of climatic sensitivity studies.

The time required to arrive at equilibrium in the stratosphere is much longer. Figure 3.14 shows the equatorial mean zonal temperature at various heights as a function of time from the stratospheric model. Figure 3.15 shows the same from various cases with the tropospheric model. Case 331 has sea-level pressure correction and 156 does not (Section 2.1). The troposphere is in equilibrium after
30-40 days, while the stratosphere takes as much as 120 days (19.5 km) to approach equilibrium (Fig. 3.14).

3.9 Variability of GCM-Generated Januaries

Chervin and Schneider investigated the variability of various Januaries of the GCM by computing the standard deviation of five independent simulated monthly means. Figure 3.16 shows the geographic distribution of the standard deviations of 1.5-km temperature, which agree reasonably well with the observations in Fig. 3.17 of Crutcher and Meserve (1970) except over the oceans, where the GCM values are too small by a factor of 2. The reason they are small can be attributed to the fact that the ocean temperatures are the same in all five GCM cases, whereas year-to-year changes are in the observed data. Other variables such as cloudiness, precipitation, winds etc. are being studied and compared to the observed variability.
Fig. 3.1 Changes in zonally averaged sea-level pressure vs. time at various latitudes.
Fig. 3.2 Changes in zonally averaged 9-km pressure vs. time at various latitudes.
Fig. 3.3 Infrared cooling rates at the start of radiation experiment: solid lines are Fels-Kaplan method; dashed are Sasamori method.
Fig. 3.4 Infrared cooling rates at the end of radiation experiment: solid lines are Fels-Kaplan method; dashed are Sasamori method.
Fig. 3.5 Zonal temperature distribution: test is Fels-Kaplan method; control is Sasamori.
Fig. 3.6 Changes in zonally averaged cloudiness caused by $\pm 2^\circ$K ocean surface temperature changes.
Fig. 3.7(a)  Left: mean January Northern Hemisphere (N.H.) geopotential distributions (dkm) at 200 and 50 mb. Center: computed 30-day mean pressure distributions (mb) at 12 and 18 km in the N.H. for Case M. Right: same as center but for Case NM.
Fig. 3.7(b) Left: mean January Northern Hemisphere (N.H.) geopotential distributions (dkm) at 30 and 10 mb. Center: computed 30-day mean pressure distributions (mb) at 24 and 30 km in the N.H. for Case M. Right: same as center but for Case NM.
Fig. 3.8 Difference in 30-day means of $T_{1.5 \text{ km}}$ between random-control, anomaly-control, superanomaly-control for Days 31-60.
Fig. 3.9  Difference in 30-day means of $T_{1.5\text{ km}}$ between random-control, anomaly-control, superanomaly-control for Days 61-90.
Fig. 3.10 Difference in 30-day means of $T_{1.5 \text{ km}}$ between random-control, anomaly-control, superanomaly-control for Days 91-120.
Fig. 3.11 Difference in 90-day means of $T_{1.5 \text{ km}}$ between random-control, anomaly-control, superanomaly-control for Days 31-120.
Fig. 3.12 Determination of the global rms noise level change as a function of averaging interval: zonal wind component $u$. 
Fig. 3.13 Determination of the global rms noise level change as a function of averaging interval: precipitation.
Fig. 3.14 Time change of zonally averaged temperature at the equator for various levels.
Fig. 3.15 Time change of globally averaged temperatures for various levels.
Fig. 3.16 Standard deviation of 1.5-km temperature of five computed Januaries.
Fig. 3.17 Standard deviation of 850-mb temperature of 14 observed Januaries.
4.0 One- to Seven-Day Forecasts of Large-Scale Atmospheric Flow Generated from the GCM

4.1 Introduction

This chapter is divided into four major sections: (1) Data Analyses, (2) Predictability, (3) Forecast Verification and (4) Forecast Sensitivity. Forecast Sensitivity is further divided into initial data, numerical techniques and physical processes. An attempt is made in each subsection to outline past research at NCAR and to emphasize future plans where pertinent. Unless otherwise noted, all comparisons discussed are valid only for mid-latitude, wintertime synoptic-scale waves for periods extending to 1 week. Many results are based on only one integration from a single data set; hence, the conclusions drawn from these experiments are tentative at best and await confirmation with DST and FGGE data sets.

4.2 Data Analyses

4.2.1 Subjectively analyzed data sets

Preparation of high-quality global data sets to provide initial states for GCM forecasting began in 1967, and since that time we have assembled two subjectively analyzed data sets. The first for 15-19 January 1958 utilized the existing IGY data and analyses and was the first coherent global input to a global GCM. The analyses consisted of 3 levels of geopotential
and 2 layers of wind for each 24 hrs. The analyses for 15 January were expanded to include 6 layers of geopotential and wind fields and the subjective analyses were interpolated by hand to a 5° grid.

Another data set collection was started in 1969 to utilize the satellite information then being provided, as well as to improve the quality of the IGY analyses. The period 6-18 December 1967 was selected for two reasons. The available satellite data were at a maximum and a substantial index cycle occurred during the 2-week period (Fig. 4.1). The geopotential was subjectively analyzed from sounding data from the USAF archives for the surface and 500 mb every 24 hrs for the entire period to provide verification and four-dimensional continuity over the globe. Daily satellite pictures of the visual cloudiness were used extensively in the Southern Hemisphere. Seven levels of geopotential extending to 70 mb were then analyzed for 6, 10 and 14 December. These dates were used as initial conditions for forecast experiments with the GCM. The analyses were objectively digitized on a 2.5° global mesh; the wind fields were analyzed for 6 December only. Most predictability and sensitivity experiments were performed using the December 1967 data set.

With the advent of high-resolution LAMs, a subset of the December data was re-analyzed incorporating fine-scale features as well as vertical consistency. Thirteen levels of geopotential
for 10-14 December were re-digitized using a 5/8° grid over the Northern Hemisphere only.

4.2.2 Objective analysis methods

Because of the time-consuming nature of subjective analysis, we have developed a global objective analysis scheme for use with the NCAR GCM. The objective method will allow us to assemble larger data bases for computing error climatology than are feasible with subjective methods. Development has been in three distinct steps. (1) The Cressman (1959) successive correction scheme which analyzed geopotential and wind on a global latitude-longitude grid was programmed (this necessitated ignoring wind observations in the tropics). (2) A multivariate statistical analysis scheme for single-point testing was then programmed (Schlatter, 1975). (3) Based upon the success of experiment (2), which produced rms differences between analyzed and observed 500-mb heights of ~ 12 m and wind component differences of ~ 4 m sec\(^{-1}\), the scheme is currently being expanded for use with the 2.5° GCM grid.

A brief description of the multivariate analysis scheme follows. The geopotential height \(h\) and the \(u\)- and \(v\)-components of the wind are analyzed simultaneously based upon observations of \(h\), \(u\) and \(v\) and a "first guess" provided by the NCAR GCM. The scheme is statistical in that the data weights depend upon covariances among "forecast errors" in \(h\), \(u\) and \(v\). Covariances involving wind
components use geostrophic approximations in mid- and high latitudes
and the nondivergent assumption in the tropics. We analyze only the
wind in the tropics and obtain the geopotential heights from the
balance equation. Poleward of 70N and 70S, we analyze data on a
polar stereographic projection. The analysis method is applied at
sea level and 10 mandatory pressure levels.

We have started testing this scheme with the subjectively
analyzed December 1967 data set described earlier and will compare
our objective analyses with Baumhefner's subjective analyses, NMC's
analyses, Bleck's isentropic analyses and with actual data. We also
plan to run tests with preliminary DST data from May 1974 which
include temperatures retrieved from radiance measurements and winds
estimated from cloud motions. We can also compare these analyses
with NMC's new global "Flattery" analyses which use Hough functions.
It is planned to couple the analyses in the vertical through the use
of interlevel statistics. The analysis scheme will be adapted for
use in high-resolution limited areas in the near future.

4.3 Predictability

In response to the First GARP Objective, we have performed a
large number of experiments to assess the predictability of large-
scale atmospheric flows. Two basic approaches are used: first,
to evaluate the present forecast skill, global data sets are used
as initial conditions for GCMs and the subsequent integrations are
compared with the actual evolution of the atmospheric systems. These experiments are aimed at testing the actual forecast skill of the data-analysis-model system. Second, the theoretical predictability of the atmosphere is examined by the use of GCM simulated data. A simulation control experiment is generated and compared to a modified experiment in which the data or model is systematically changed. However, if the simulated data are not as realistic as the actual atmosphere, the results may be misleading.

4.3.1 Forecast skill

The first experiments using the GCM to predict the global atmosphere were conducted with a 5° 2-layer model (Baumhefner, 1970). Overall skill was noted out to 2 days at the surface and to 4 days at 6 km (Figs. 4.10 and 4.2). The decrease in amplitude of the synoptic-scale waves was quite evident in both hemispheres near the end of the forecasts.

Global forecasts were also carried out with the 2.5° 6-layer GCM, using the December 1967 data set for initial states. An example of one of these forecasts is displayed in Fig. 4.3. The error distribution indicates a mass loss occurring in the model (differences primarily of one sign), as well as problems near the pole. A forecast error climatology based on integrations from 12 hemispheric initial states selected from the 6-18 December 1967 data set is currently being evaluated; the forecasts again were
made with a 2.5° 6-layer GCM. Figure 4.4 shows the range of rms error for these cases. The actual predictability of the 2.5° GCM is roughly the same as that of the 5° model, except for a slight improvement at sea level. The amplitudes of the synoptic-scale waves are more accurately calculated; however, the verification scores are insensitive to this improvement (Section 4.4). A breakdown of the error by two-dimensional wavenumber (Fig. 4.5) shows the most predictable waves to be the large-scale baroclinic modes (wavenumbers 4 to 8). The very long waves (wavenumbers 1-3) and shorter waves (wavenumbers over 9) have less skill. The forecasts approach an equilibrium climate different from the observed within 10-14 days indicating that certain systematic biases exist in the model. Figure 4.6 indicates the changes in the zonal wind and temperature after 4 days of integration. The overall skill of the NCAR GCM is equivalent to other models of the same complexity (Table 4.1). The figures, other than NCAR's, are taken from Druyan (1974) and do not cover the same data periods. The numbers in parentheses indicate the latitudinal averaging interval. It is quite apparent that no one model has consistently better skill over the others at all levels and all times. The variation in the statistics can be attributed to different averaging intervals in space and time, the smoothing applied to the model output and differences between standard deviation which removes mean bias and rms
error. A "Monte Carlo" combination (Leith, 1974a) of 6 GCM-produced forecasts with slightly perturbed initial conditions exhibited only a 5% reduction in rms error when compared to the error of a single forecast.

4.3.2 Theoretical predictability

Predictability experiments have been carried out with simulated data to consider the effect of resolution on predictability (Williamson, 1973) by comparing a 5° model forecast with data generated by the 2.5° model. An experiment was also performed to consider the effect on predictability of the physical processes included in the models (Williamson, 1974b), by comparing a forecast from a 5° model, in which the total heating is set to zero, with data generated by a 5° model in which the heating is computed as usual. In the experiment comparing the 5° and 2.5° models, the error reaches 25% of its asymptotic value after 12 hrs even though the initial error is zero (Fig. 4.7). The error is very close to its asymptotic value after 2 weeks. The rate of error growth for forecasts from real data described in Section 4.3.1 is somewhat slower for the first few days. In the case of zero total heating, the error reaches 10-15% of its asymptote value in the first 12 hrs and after about 1 week becomes very similar to the error of the 5° and 2.5° experiments. These
curves indicate that both numerical truncation errors and systematic errors in the physical processes may be major sources of forecast error in actual practice.

Since lack of quality global data sets has hindered the proper assessment of actual predictability, we have embarked on an ambitious program in conjunction with the DST to analyze several periods throughout the seasons. These data sets will be used to identify and correct the major problems associated with medium-range (~7-day) forecasts and hopefully lead to an extension of the present skill.

4.4 **Forecast Verification**

Three types of scores have been used to verify GCM forecasts—the rms error, the $S_1$ scores (a gradient measure used by NMC) and correlation coefficients. A normalized rms error has been used recently (Baumhefner and Julian, 1972, 1975) in which the actual rms is divided by the natural variance of either the model or the atmosphere. The correlation coefficients were calculated from both the actual patterns and the differences of the forecast and verification from the respective mean value. A comparison of harmonic analyses (Fourier or spherical) from the forecast and verification fields has also been used as a method of verification.

It has been found that verification scores such as the rms error and $S_1$ score are biased toward smooth patterns (Baumhefner,
A 5-10% reduction in error was achieved by applying a smoother to the model output. The correlation coefficients were judged insensitive to small changes in the patterns and, therefore, are no longer used. To date, the best way to judge the overall skill of a particular forecast is to examine the actual difference maps in relation to the verification fields along with the conventional skill scores. The reduction of one particular score does not guarantee an increase in forecast skill.

The limit of skill has been defined in several ways. The most widely used method—the rms persistence crossover—tends to overestimate the skill of the forecast. We prefer to evaluate the limit of usefulness with several criteria. If the difference between forecast and verification exceeds the natural variance of the atmosphere over more than 50% of the verification region or the phases of the baroclinic disturbances are different by more than 90°, the limit of usefulness has been reached. This usually corresponds to a normalized rms error of 0.8.

Forecast comparisons are being contemplated with the use of spherical harmonic expansions (Leith, 1974b). Orbits (phase-amplitude diagram) of various forecast and verification harmonics can be plotted and differences examined.
4.5 Forecast Sensitivity

Sensitivity studies can be divided into experiments (1) testing the accuracy, analyses and initialization of the observed variables, (2) examining the effects of grid resolution, coordinates, accuracy and precision on the forecasts and (3) testing various physical parameterizations in the model.

4.5.1 Initial data

4.5.1.1 Observational error

In this category, various types of observational errors have been introduced into an analyzed initial state of the atmosphere and a forecast made from the modified data and compared to a control forecast with unmodified initial conditions. Large-scale random and systematic errors were examined (Baumhefner and Julian, 1972). The systematic error was introduced only in longitudinal wavenumber 6. The rate of error growth is faster for a systematic error than a random error and is sensitive to the horizontal resolution of the model. Further studies, which realistically simulated the error associated with a proposed observing system of remotely sensed temperature profiles (Baumhefner and Julian, 1975), showed that systematic errors due to cloudiness contamination grow catastrophically at the end of a 1-week forecast. However, an experiment containing a constant
error in the horizontal with a monotonic increase in the vertical 
and another including a 2° error imposed on wavenumber 1 in the 
stratosphere indicated very slow error growth.

If a reference level is used to determine, by vertical inte-
gration, the geopotential from a known temperature profile, the 
best location of the reference is at sea level. This location 
results in a lower error level being reached at the end of an 
integration compared to other locations (Baumhefner and Julian, 
1975). The accuracy of data at the reference level, if the errors 
are not systematic, does not seriously affect the error growth 
rate.

Another way in which the model can be used to measure the 
sensitivity of observing systems is through the use of simulated 
data. In response to the request of the Joint Organizing Committee 
for GARP (Report of the Ninth Session of the Joint Organizing 
Committee for GARP, Canberra, 8-12 January 1974), we performed 
a series of Observing Systems Simulation Experiments with simulated 
data to aid in the design of special First GARP Global Experiment 
observing systems. These experiments differed from earlier ones 
in that the perfect model assumption was not made, i.e., the data 
were generated by a different model from that used for the updating. 
Our main results agree quantitatively with those from the United 
Kingdom Meteorological Office, Bracknell, but result in much larger
absolute errors due to the lack of optimum interpolation. The major conclusions of these and earlier experiments, as summarized in the Report of the Tenth Session of the Working Group on Numerical Experimentation, Copenhagen, 19-22 August 1974, are:

"(i) It is imperative that wind observations in the tropics are obtained from a network of carrier balloons and/or ships as there is unlikely to be an adequate network of wind observations obtained by tracking clouds. In addition, the carrier balloon and/or ship network will provide temperature information which will also be essential unless the RMS error of the satellite information is reduced to well below the present 2.5°C.

(ii) Significant reductions in the rms errors of both surface pressure and zonal wind were obtained in the simulation experiments when data from constant level balloons in the southern hemisphere were added to the basic data set. It is therefore recommended that a network of constant level balloons be maintained in the southern hemisphere with a spacing of about 1000 km.

(iii) The simulation experiments indicate that it is important to know the surface pressure (or some low-level reference level in the southern hemisphere). The recent experiments indicate that better overall results can be obtained with the same number of buoys, by using a loose network (1000 km spacing) of buoys between 20°S and 65°S than by a dense network (500 km spacing) between 50°S and 65°S. As far as a possible gap (80° - 140°W) in the buoy array is concerned, the numerical experiments do not show any significant influence of this gap on the results."

Proposed observing systems to be used for GARP data sets have error characteristics that are different from the conventional observing systems and will affect the actual predictability if not treated properly. We are attempting, by working with DST data, to
assess these characteristics and minimize their effect in both the data analyses and model forecasts.

4.5.1.2 Analysis difference

Several experiments comparing the analysis of atmospheric variables were examined. The subjectively analyzed December 1967 NCAR data set was compared to the operational NMC analyses for the same period. The initial rms difference is 2-3 times larger than the typical instrument error with most of the difference over data-sparse areas. However, the forecast difference was not large for the period of 1 week. Other experiments comparing different analysis techniques (Section 4.2.2) are currently underway.

A subjectively analyzed large-scale moisture field was used as an initial condition and compared with forecasts which used a zonal climatological average or values diagnostically calculated from the temperature field. The modified versions produced virtually no error growth. This result is probably sensitive to the scales of motion that are forecast by the model and, therefore, further testing is necessary.

4.5.1.3 Lateral boundary ("wall") effects

The exclusion of certain initial data by introducing lateral boundaries in the global GCM has been studied intensively. A southern boundary ("wall") was placed in the 5° and 2.5° models
at various latitudes north and south of the equator (Baumhefner, 1970, 1972). The resulting integrations from the January 1958 and December 1967 data sets were compared to the global forecasts. The effect of these boundaries on Northern Hemisphere mid-latitude waves was substantial for wall positions north of the equator. Figure 4.8 shows the rms difference between the forecasts for various boundary positions. Positions at or south of the equator yielded acceptably low error growth for periods up to 1 week.

Experiments with an LAM, in which forecasts are integrated over a 90° longitudinal window with inflow boundary conditions provided by a global model (Williamson and Browning, 1974), have produced acceptable results. Forecasts using both simulated and real atmospheric data show little deviation out to 2 days for large-scale flow when compared to their global counterparts.

4.5.1.4 Initialization

Once the observed data have been analyzed, they must be initialized to eliminate unwanted wave modes which may interfere with meteorologically significant waves. Considerable effort has been spent in examining the relative effect of various initializations on the actual predictability of the GCM.

The conventional forms of initialization, such as the geostrophic, linear balance and complete balance equations with
nonlinear terms were tested with the 5° model (Baumhefner, 1970) and the 2.5° model. Calculation of the wind near the equator involved simple interpolation across the equator or the use of a constant f value in the geostrophic equation. The various forms of the balance equation were tried with and without the Jacobian term. The complete balance equation was solved over the entire globe (Houghton and Washington, 1969) and over only the mid-latitudes. In the latter case, an observed wind field was used in the tropics and the geopotential was derived from the balance equation. For the complete balance case, several initial vertical motion fields were calculated (Houghton et al., 1971). A forecast from independently analyzed pressure and velocity fields was also made to assess the types and amplitudes of waves produced from an unbalanced initial state and to compare them with initialized forecasts.

The results indicated that the geostrophic equation was slightly better than the others, mainly due to lack of smoothing of the initial velocity maxima. The numerical solutions of the second-order equations inherent in the other initializations produced the unwanted smoothing. None of the initializations tested produced a significant gain in forecast skill. An unfortunate characteristic of the geostrophic wind initialization is the generation of large Lamb waves with amplitudes and
wavelengths comparable to Rossby waves, but with much higher frequency. Both a divergence filter imposed on the wind field (Washington and Baumhefner, 1975), which sets the vertical integral of divergence to zero, and the nondivergent geostrophic wind eliminate the unwanted waves. During the integration, we obtained smoother and more realistic pressure distributions; however, there was no significant gain in forecast skill. These studies seem to indicate that the gravity waves appearing from unbalanced initial states do not interact strongly with the mid-latitude Rossby waves over a time period of less than 1 week.

A procedure has been developed theoretically to filter unwanted gravity waves from initial data using the normal modes of a particular forecast model (Dickinson and Williamson, 1972). Experiments are being initiated to test this procedure in practice. The technique may also be used for diagnostic studies of both models and analyses.

4.5.1.5 Four-dimensional assimilation

Another way of initializing atmospheric data—four-dimensional assimilation—has been tested with both simulated and real atmospheric data. The method uses the model directly to establish a balance between the mass and velocity fields.
Early experiments using the perfect model assumption with simulated data (Williamson and Kasahara, 1971; Kasahara and Williamson, 1972; Kasahara, 1972) and an accompanying theoretical study (Williamson and Dickinson, 1972) demonstrated that if the prediction models are sufficiently accurate and if data are homogeneous and accurate, four-dimensional data assimilation involving simple updating is sufficient to determine the mass field from the wind field and vice versa.

More recent experiments using simulated data with imperfect models (Williamson, 1973, 1974b), along with a theoretical study (Blumen, 1975a,b), indicate that errors in current operational prediction models and observed data are probably large enough to foil the simple updating procedure and require an optimal interpolation scheme to aid in determining winds from temperatures.

Updating experiments using NMC analyses as data were performed. Twelve-, 6- and 2-hr intervals were used to update the pressure fields only. A comparison of the results with updating simulated data and atmospheric data is shown in Fig. 4.9. It was found that the wind error actually increased with time during the update, contrary to the perfect and imperfect model results. Forecasts made from the updated variables were inferior to those using the conventional geostrophic initialization, mostly due to the large residual error left by the updating scheme.
Because of the large spatial inhomogeneities as well as inaccuracies in current observations, we await a more complete data set, such as the DST series, to test the more sophisticated methods of four-dimensional assimilation.

4.5.2 Numerical techniques

Two major factors influencing the results obtained from finite-difference methods of integration are grid resolution and the differencing algorithm. Extensive testing on increased grid resolutions in both the horizontal and vertical has produced results closer to the observed.

4.5.2.1 Vertical resolution

Changes in vertical resolution were made with 3 versions of the GCM by simply interpolating real atmospheric data from the coarse mesh to the finer versions. Experiments with the 5° global model, in which the number of layers was expanded from 2 to 6, yielded a reduction in rms error of 5-10% (Baumhefner, 1969) (Fig. 4.10). The vertical resolution in the third-generation GCM, which used a mesh of 2.8°, was varied from 6 to 18 layers, holding the height of the top fixed. A comparison of the differences (Fig. 4.11) shows a slight improvement for the 18-layer version. The resulting rms differences deviated by less than 5%.

The model sensitivity to the vertical resolution becomes more significant when the horizontal resolution is increased. The LAM
version of the GCM was integrated at 1.25° with 6 and 18 layers of information. Forecasts of surface features were somewhat better with the 12-layer LAM. These experiments show that the actual model predictability is sensitive to vertical resolution; however, the magnitude of the effect is relatively small. Further tests are being made with the initial data analyzed at the highest possible resolution with the aid of isentropic analysis.

4.5.2.2 Horizontal resolution

Horizontal resolution experiments have produced the largest changes in integration comparisons. A 2.5° 6-layer global model was compared to a 5° 6-layer version. The forecasts of synoptic-scale wave amplitudes were much improved; unfortunately, the rms errors did not show a significant change. It was judged, however, that the higher resolution forecast was closer to the real atmosphere. Further experimentation with the LAM, in which the horizontal resolution was reduced from 5° to 5/8°, also indicated a strong sensitivity to grid spacing. In fact, the 1.25° and 5/8° forecasts begin to pick up the fine-scale details associated with the large-scale waves (Fig. 4.12 (a) and (b)). These resolution experiments suggest that an asymptotic forecast solution from the initial data used has not been reached, even at 5/8°, for scales of motion on the order of 1000 km.
Since all experiments above were based on initial data originally gridded at 2.5°, it is important to verify these results with a high-resolution data set such as the December 1967 case. Utilizing the DST data archives, we are also planning to analyze a data set at 1/2° horizontal mesh and 24 layers in the vertical.

4.5.2.3 Accuracy

The accuracy of numerical integrations of atmospheric flow has been investigated for medium-range forecasts with real atmospheric data. An experiment was conducted with the third-generation model in which the horizontal derivatives were approximated with second- and fourth-order accuracy; comparison of the 2 forecasts produced little change in the first 2 days. A small change was noted at the end of 3 days with the fourth-order scheme showing slightly higher skill. A comparison between the second- and third-generation models, including more accurate numerical techniques near the poles, indicated that the skill of the current form of the third-generation model has not surpassed the second-generation version (Fig. 4.11). A possible explanation of this result, having to do with the horizontal diffusion, is discussed in the next section.

The importance of precision in numerical prediction was examined by truncating the arithmetic with a 48-bit mantissa to
21 bits (Section 3.3). The forecast error growth due to truncation was much smaller than a typical predictability error growth. For a 1-week forecast, the effect is minimal.

4.5.3 Physical processes

Discussion of GCM forecast sensitivity of various physical processes included in the model can be divided into (1) horizontal diffusion, (2) boundary effects and (3) parameterization of heat sources and sinks.

4.5.3.1 Horizontal diffusion

Three types of horizontal diffusion techniques have been applied to the GCM when using real atmospheric data. The second-generation model uses a diffusion form proportional to the deformation and having an empirical constant $k_0$ that must be determined. Values of $k_0$ lower than .3 produced computationally unstable results, whereas values higher than .45 yielded overly smooth features. The forecast skill was not sensitive to values of $k$ inside the range .3-.45.

The third-generation model was tested with real data using two different versions of horizontal diffusion. The first was a spectral chopping technique in both latitudinal and longitudinal directions and the second used a fourth-order smoother in latitude and chopping in longitude. These diffusion techniques are discussed
in Williamson (1974a). The spectral chop consisted of setting to zero all waves smaller than $4\Delta x$ at a specified time interval. Chopping at 30- and 90-min model time intervals gave similar results. Longer time intervals between chopping produced computational instability. Considerable alteration of the initial energy spectra occurred during a 3-day forecast with the chopping method. Substituting the smoother in the latitudinal direction produced little change.

The large difference in the evolution of the energy and enstrophy spectra between the second- and third-generation models suggests that the type of diffusion used is responsible for the lower predictability of the new model. Further testing of the third-generation GCM with both simulated and real data is necessary to determine the proper horizontal diffusion (Section 3.5.8).

4.5.3.2 Lower boundary conditions

The sensitivity of the model to lower boundary conditions, such as orography, sea surface temperature and ice-snow locations, has been examined. Forecasts of real data in which the actual orography was replaced by 10-m values were compared to standard forecasts with realistic mountains. Deviations between the forecasts at the end of 7 days are quite large with differences appearing in all scales (Fig. 4.13). Tests with artificial mountain terrain inserted in the model, such as very steep slopes and large mountain ranges, produced
similar sensitivities to the forecast. The skill of the forecast with mountains is not, however, substantially better than the skill of a no-mountain case.

A January mean ice-snow line and an actual December analysis of these values were introduced as boundary conditions for two forecasts. The response of the integration to these changes was quite small for a 1-week forecast.

4.5.3.3 Latent heating

Several forms of stable and convective latent heating have been tested in the GCM using real data. Two variations of large-scale stable rainfall, in which the amount of latent heat released varied by a factor of 2, produced small changes for the first 3 days. At the end of 1 week, the zonally averaged temperature had changed about 3°C in the mid-latitudes; the skill of the 2 forecasts remained almost the same.

The simple convective adjustment, used in early versions of the GCM, was compared with a more sophisticated convective parameterization scheme. The parameterization developed by Kuo (1965) and Krishnamurti and Moxim (1971) relates cumulus convection to large-scale moisture convergence. Several variations of the formulation were tested in the global 2.5° GCM (Washington and Baumhefner, 1974) and 1.25° and .625° LAMs. For the one case used as initial conditions, the effect of the parameterization on the evolution of
large-scale flow was minimal. In the higher-resolution forecasts, intense rainfall developed and the effect of convective heating was overestimated. A large number of forecasts of weather systems that appear sensitive to convective processes need to be performed and examined to quantify the previous results.

4.5.3.4 Longwave cooling

The sensitivity of forecasts to the radiation parameterization was examined with the 5° model using simulated data (Section 3.5.1). Two initial states were forecast, one with a 10% random error imposed in the longwave cooling calculation. The forecast differences at the end of 1 week were indistinguishable.

4.5.4 Summary

It should be pointed out, again, that most sensitivity studies presented here were performed with only one initial state. It is our goal to provide a broad data base, from the DST, to evaluate more quantitatively the results presented. Some tentative conclusions, however, can be drawn from these forecasting experiments for medium range (up to 7 days). The model appears most sensitive to certain types of observational error, horizontal resolution and mountain formulations. The least sensitive areas have been initialization, numerical accuracy and precision
and physical parameterizations of heat sources and sinks. The importance of other topics, such as analyses differences, four-dimensional assimilation, vertical resolution and horizontal diffusion, is not yet known.
Table 4.1  rms errors of NCAR model compared to NMC, GFDL, GISS and FNWC. ( ) indicate latitude interval of statistics.

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<th>NCAR (30-70)</th>
<th>NMC (18-P)</th>
<th>GFDL (0-P)</th>
<th>GISS (22-P)</th>
<th>FNWC (20-P)</th>
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500-MB VERIFICATION - RMS, STANDARD DEVIATION (METERS)

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RMS S.D. S.D. RMS RMS
Fig. 4.1 Longitude-time plot of 500-mb geopotential for December 1967 at 40°N. Contours every 60 m. Horizontal lines are dates from which global forecasts were made.
Fig. 4.2 96-hr forecasts of 6-km pressure. Contours every 5 mb. Skill scores at right of panels.
Fig. 4.3 48-hr global forecast of 6-km pressure (top) from 2.5° 6-layer model and verifying map at 12 December 1967, 12Z. Light lines are pressure-contoured every 10 mb. Heavy lines are pressure difference every 8 mb. Vertical cross-hatch is negative error; stipple is positive error.
Fig. 4.4 rms error from 30–70°N of 12 hemispheric forecasts made from December 1967 data. Error bars indicate maximum and minimum values.
Fig. 4.5 rms error of a single forecast for 3 days, normalized against the natural variance of the atmosphere as a function of two-dimensional wavenumbers (abscissa).
Fig. 4.6 Meridional cross-sections of zonal wind (top) and temperature (bottom) at Day 4 of a global forecast. Light contours show initial distributions with intervals of 5 m sec\(^{-1}\) and 5°, respectively. Differences are indicated by heavy line (every 2 m sec\(^{-1}\), 2°) from the observed values. Stippled areas are positive; shading is negative.
Fig. 4.7 rms error of temperature (upper) and zonal wind (lower) for various predictability experiments.
Fig. 4.8 Skill scores between control (global case) forecasts and versions with a southern boundary as indicated.
Fig. 4.9 rms error of zonal wind when temperatures are used as updating variable for various cases. RD are NMC analyses; others are simulated data.
Fig. 4.10 Comparison of 2- and 6-layer 48-hr surface forecasts with a 5° grid. Verification at the bottom.
Fig. 4.11 72-hr forecast of 6-km pressure and differences $\Delta p$ from the observed fields. Top: $2.5^\circ$ 6-layer, second-generation model. Middle: $2.8^\circ$ 6-layer, third-generation model. Bottom: $2.8^\circ$ 18-layer, third-generation model. Light lines are pressure every 10 mb; heavy lines are $\Delta p$ at 8 mb. Stippled areas are positive error (forecast $> \text{observed}$). Shaded areas are negative.
Fig. 4.12(a) Initial conditions (left) and verifications (right) for the horizontal resolution experiment shown in Fig. 4.12(b). Pressure contoured every 4 mb.
Fig. 4.12(b) 24-hr forecast of 6-km pressure starting from 12Z 10 December 1967 (see left side of Fig. 4.12(a)) using four horizontal resolutions—5°, 2.5°, 1.25° and 0.625°, as indicated. Pressure contoured every 4 mb.
Fig. 4.13 7-day forecast of 6-km pressure with and without mountains (from 1967 data). Verifying data (top).
5.0 Conclusions

Having reviewed past and present activities of general circulation modeling at NCAR, we now identify particular areas in the GCM needing improvement for application to numerical weather forecasting and climate studies.

5.1 Modeling

The GCM, an indispensable tool for studying the general circulation of the atmosphere and performing medium-range forecasts, must continuously undergo improvement. The dynamical framework and programming of the third-generation model are sufficiently flexible to allow varied experimentation on present and future computers.

Continued research is required to solve satisfactorily the numerical problems arising from applying the prediction equations near the poles. It is also important to develop accurate methods of dealing with the earth's orography in the model. The transformed coordinate in the third-generation model conserves the total mass so that the surface pressure adjustment, needed in the second-generation model (Section 2.4), is not required.

Although the evidence is not conclusive, it is desirable to investigate alternative upper boundary conditions. We recommend initially a separate study using a simpler model.
It is now apparent that to simulate eddy flow comparable to the real atmosphere, the horizontal resolution of the model should be less than ~ 2.5° with the current second-order scheme. More research is necessary to shed light on the trade-off between increasing the accuracy of a finite-difference scheme and reducing the horizontal resolution of the computational mesh. A similar effort should be made to determine the effect of increased vertical resolution. In terms of computation, doubling the vertical resolution is only one-fourth as costly as doubling the horizontal resolution. Again, a simple model can be used to test such trade-offs.

To judge model performance, it is desirable to compile better climatological statistics from observed data, particularly second- and third-moment quantities such as eddy kinetic energy and various energy conversion quantities. The seasonal characteristics of the model should be checked with reference to monthly and annual observed statistics and their month-to-month and interannual variability.

Examination of model performance using zonal mean statistics alone is not an effective means of comparison with observed data; we need to examine geographical distributions of climate. In this respect, with the GCM we should pursue investigations of world major regional climates such as the Asian-African summer monsoon now under study.
5.2 **Physical Processes**

Parameterization of physical processes in GCMs must continuously be revised to improve fidelity of the model. Since the model responds collectively to physical processes, it is essential to improve their overall quality. We must also be aware of feedback between computational methods (for dynamical equations) and physical processes. For example, the parameterization of subgrid-scale processes may be closely related to the horizontal resolution and accuracy of finite-difference approximations.

Even in short- to medium-range forecasts and climate simulation, in particular, heating and frictional terms in prediction models should be accurately formulated. Unfortunately, the details of physical processes in the atmosphere are often difficult to quantify or to treat within limited model calculations. We do not yet know how to parameterize adequately the collective effects of cumulus convection in prediction models and how to predict the distribution of clouds for radiative calculations. On the other hand, the method of computing solar and terrestrial radiation in the atmosphere is well known, but the computing time required to solve radiative transfer equations is prohibitive unless the calculations are simplified.

We should further improve parameterization of thermodynamic processes, namely moist convection, atmospheric and ground hydrological cycles, including their cryospheric coupling, and
cloud-radiation interactions. It is essential to treat accurately the planetary boundary layer.

It is well known that energy dissipation occurs mainly in the planetary boundary layer, but a significant part also takes place near the tropopause level, presumably a result of clear-air turbulence and breaking gravity waves. Although the importance of these mechanisms to weather prediction and climate simulation is yet to be determined, search for a better parameterization of subgrid-scale motion in prediction models should continue.

5.3 Prediction Experiments

The first objective of GARP is to extend the useful range of weather forecasts beyond the present limit of about three days. The development of objective analysis of meteorological data at NCAR is well on the way to meeting the schedule of the Data Systems Test (DST). Adaptation of second- and third-generation GCMs for medium-range prediction experiments is progressing. Although the decision must soon be made on whether to use the second- or third-generation model for DST prediction experiments, depending on the progress in tuning the third-generation model, the decision will not seriously jeopardize the immediate objective of DST--the real-time rehearsal for FGGE in 1977-1978.

As far as judging forecast skill, the conventional verification measures, such as root-mean-square errors and $S_1$ scores,
are not entirely satisfactory, since the improvement of one particular score does not necessarily reflect an increase in forecast accuracy. Until better objective verification methods are developed, subjective judgment on the actual difference maps should continue, along with conventional verification procedures.

Many sensitivity studies have been carried out at NCAR to examine the effects on forecasts of (1) the accuracy, analyses and initialization of observed variables, (2) grid resolution and accuracy of difference approximations and (3) various physical parameterizations in the models.

Systematic observational errors in initial data are certainly detrimental in forecasting, but if observed data are sufficiently dense and accurate, it appears that the general quality of forecasts is not strongly influenced by differences in initial data analysis. Changes in various physical parameterizations in the models generally do not strongly affect the short-range transient motions.

The most sensitive factor contributing to improvement in short- to medium-range forecasts is the horizontal grid resolution in the prediction models. Another sensitive factor is the formulation of the earth's orography.

Initialization schemes tested so far have not produced significant changes in forecast skill. If initial data are sufficiently
dense and accurate and if a four-dimensional data assimilation operates successfully, data initialization may not be required. Since some analysis methods adopt forecast values as initial guess, any improvement in the prediction model reflects a better analysis and contributes to an even better forecast. Over data-sparse areas, the data initialization remains a problem, particularly in the tropics where it is difficult to separate gravity waves from Rossby/Haurwitz-type waves. The application of normal mode analyses being investigated may offer a solution.

Since lack of adequate computational resolution in the models in both the horizontal and vertical is a prime suspect in forecasting failure, testing the impact of varying computational resolutions in the limited-area model should have high priority. The LAM is now capable of forecasting on a 5/8° horizontal mesh with 24 vertical layers.

With the DST data, NCAR's plan is to prepare high-density global data sets which will be the best ever available for global forecasting studies. We recommend experimenting with the LAM and a high-resolution global model using the DST data to clarify the factors contributing to degradation of short- to medium-range forecasts.

5.4 Climate Studies

We have already said that accurate formulations of physical processes in GCMs are crucial for a successful reproduction of
climate. When an appropriate energy balance is not maintained as in the real atmosphere, the model eventually seeks its own equilibrium climate different from the real one. Thus, improvement in the formulation of various physical processes in the model should be given priority in climate studies. Documentation of model climatology, particularly in seasonal variations, must be made to provide a basis for further consideration in modeling.

Just as the skill of medium-range forecasts is judged against the predictability of the atmosphere, climate simulation must be examined in light of sensitivity to climate variations. Since ocean surface temperature is specified in the present GCM, the model is not a climate model. Nevertheless, when experiments are properly formulated, the model can be useful, within its limitations, in gauging climate sensitivity to advertent and inadvertent changes in environmental forcing. Results of such experiments, however, must be examined carefully for statistical significance and viewed in light of mathematical shortcomings in modeling.

Taking into account the above limitations and imperfections, it is clear that many sensitivity experiments with the second-generation GCM during the last two years have provided useful information, otherwise unobtainable, on the intricate coupling of physical processes in the atmosphere.
Specification of ocean surface temperatures in the model bypasses the need for a delicate radiation energy balance—the primary regulator of climate. For this reason, it is desirable to couple an atmospheric model with an active ocean model to test the energy balance aspect.

Similarly, a more detailed treatment of physics is required in order to investigate the general circulation in the stratosphere. Photochemistry and transport of ozone will be the principal additions to the stratospheric model, and careful finite-differencing is needed to solve the transport equation for ozone and other chemical variables. Since the stratospheric circulation is apt to be influenced by the troposphere rather than vice versa, no stratospheric model will produce meaningful results without inclusion of a realistic tropospheric model.

Finally, we emphasize the utility of a one-dimensional model as a means of economizing computer time. Before an experiment is designed for the GCM, we should examine the possibility of using a one-dimensional model to estimate the response, for instance, to a change in the parameterization of a physical process. Along the same line, GCM users should use GCM history tapes whenever possible for evaluation of modifications in the model before executing changes in an interactive mode with the GCM.
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