THERMAL AND DYNAMICAL EFFECTS OF OROGRAPHY ON THE
GENERAL CIRCULATION OF THE ATMOSPHERE

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September 1968

1. INTRODUCTION

Although much literature is available on the thermal and dynamical effects of orography on large-scale motions, the question of the relative importance between the thermal and dynamic effects has not been adequately explained.

Charney and Eliassen (1949), Bolin (1950), and others put forth the hypothesis that the position of the semi-permanent high-level flow patterns in the atmosphere is determined primarily by the dynamic effect of the large-scale mountains such as the Himalayas and Rockies. Their argument is partially based on the observation that certain basic characteristics of the flow pattern at upper levels in the Northern Hemisphere do not change essentially from summer to winter, in spite of the reversal in the thermal contrast between the continents and oceans.

Sutcliffe (1951) objected to accepting the above orographic-dynamic hypothesis to explain semi-permanent tropospheric flow patterns. He argues qualitatively that a thermal-synoptic explanation, resting mainly on the direct thermal effect of land and sea modified by a baroclinic synoptic process, is satisfactory.

Up to now theoretical investigations of orographic-dynamic effects were based mostly on the solutions of barotropic equations of motion in either linearized or nonlinear form. Kasahara (1968) examined in detail the numerical solutions of barotropic nonlinear equations in spherical geometry for January and July global atmospheric conditions. Kasahara concluded that the orographic-dynamic effect is not likely to be a major factor in determining the mean circulation patterns in the mid-troposphere during the Northern Hemisphere summer season, and that the orographic-dynamic effect is small in the Southern Hemisphere except possibly for the Antarctic region. Therefore, if the orographic-dynamic effect is important, its effect must be dominant only during the Northern Hemisphere winter season, and possibly over Antarctica for all seasons. If this conjecture is correct, the thermal-synoptic effect may indeed be more important than the orographic-dynamic effect in determining the mid-tropospheric flow patterns.

In their studies of the influence of large-scale heat sources and sinks, Smagorinsky (1953) and Gilchrist (1954) formulated mathematical models which assume that the motion is dependent on the height and the east-west direction, but independent of the north-south direction. Döös (1963, 1968), Staff Members of Academia Sinica (1958), Saltzman (1963, 1965), and Sankar-Rao (1965) considered the same problem by dealing with the three-dimensional large-scale motions. However, only linearized models were considered in their studies.

Another shortcoming in the previous studies is that the distribution of heat sources is prescribed as a forcing function in the model. Ideally it is
It is thus clear that when both the thermal and dynamical effects are included simultaneously in a baroclinic and nonlinear model, the size of the computation becomes as large as a general circulation experiment. Thus, the use of a general circulation model is very appropriate to investigate the present problem.

2. INCORPORATION OF OROGRAPHY

The dynamical effect of orography is incorporated in the NCAR general circulation model in the following unique way. Figure 1 shows a typical cross-section in which mountains are indicated by shading. For the six-layer version, the height increment, $\Delta z$, is 3 km. Mountain height is indicated by $H$.

Surface pressure, $p_s$, is evaluated along the mountain surface. Pressure, $p$, and vertical velocity, $w$, are placed at $z = 3, 6, 9, 12$, and $18$ km, and wind components, $u$ and $v$, temperature $T$, density, $\rho$, and the mixing ratio of water vapor, $q$, are placed in the intermediate levels, i.e., $z = 1.5, 4.5$ km, etc. The variables $u, v, T, \rho$, and $q$, which correspond to variables in the surface boundary layer, are evaluated from the lower boundary conditions as described by Kasahara and Washington (1967), henceforth called Paper I.

As the upper and lower boundary conditions on the vertical motion, we assume that

\begin{equation}
\begin{aligned}
w &= 0 \text{ at the top of the atmosphere, } z = z_T, \quad (1) \\
w &= \nabla \cdot \nabla H \equiv v_s \text{ at } z = H. \quad (2)
\end{aligned}
\end{equation}

The prediction equations applied above the layers containing mountains are the same as described in Paper I. However, the computing procedure must be changed in the lower levels due to the presence of mountains. As indicated in the right-hand column of Fig. 1, the odd integer levels, shown by dashed lines, are always placed exactly in between the nearest even integer levels. We consider, therefore, that the variables at an odd integer level represent vertically averaged values of the variables within the layer bounded by the two nearest even integer levels. This consideration is used to formulate the equations of motion applicable to the levels $k = 1 \frac{1}{2}$ and $k = 2 \frac{1}{2}$.

By integrating the equation of motion (2.1) in Paper I from $z_k$ to $z_{k+1}$ and dividing by $(z_{k+1} - z_k)$, we have

\begin{equation}
\begin{aligned}
\frac{\partial}{\partial t} (\rho u)_{k+1 \frac{1}{2}} &= - \frac{1}{z_{k+1} - z_k} \left[ \nabla \cdot (z_{k+1} - z_k) (\rho u_l)_{k+1 \frac{1}{2}} + (\rho u w)_{k+1} \right] \\
&\quad - \frac{1}{a \cos \varphi} \frac{\partial p}{\partial \lambda} \bigg|_{k+1 \frac{1}{2}} + \left( f + \frac{u}{a} \tan \varphi \right) \rho v |_{k+1 \frac{1}{2}} + F_\lambda |_{k+1 \frac{1}{2}}.
\end{aligned}
\end{equation}

We get a similar equation for $\partial (\rho v) / \partial t$ by averaging Eq. (2.2) of Paper I. In deriving (3), we used the following useful formula:
(4)

\[
\Delta z \left[ \nabla \cdot \left( \frac{\partial w}{\partial z} \right) \right] dz = \nabla \cdot \left( \Delta z \int H W dz \right) + \nabla \cdot \nabla H + \nabla H \left|_{\Delta z} \right. - \nabla H \left|_1 \right.
\]

The last and the third term from the last of (4) cancel each other because of the lower boundary condition (1) on w.

Since the levels \(k = 1 \frac{1}{2}\) and \(2 \frac{1}{2}\) may not be horizontal over mountains, an interpolation procedure is used to evaluate the horizontal pressure gradient term in (3) and the \(k+\frac{1}{2}\) level. For details of the computation of the horizontal pressure gradient, see Oliger, Kasahara, and Washington (1968).

We shall now discuss modifications to the computations of the surface pressure \(p_k\) and the pressure tendency at the top level, and evaluation of the vertical velocity. The pressure tendency equation is given by (2.10) in Paper I. If level \(k\) is the mountain surface, then we have

\[
\frac{\partial p_k}{\partial t} = B + g(\rho w)_k - g \int_{z_k}^{z_T} \nabla \cdot (\rho \nabla) dz,
\]

where \(B = \partial p/\partial t\) at \(z = z_T\). Since the \(k+\frac{1}{2}\) level is, in general, not horizontal, the calculation of the horizontal divergence in (5) demands extra work. This difficulty can be removed by reversing the order of the two operations, integrating and differentiating. Using (4) we can rewrite (5) as

\[
\frac{\partial p_k}{\partial t} = B - g \int_{z_k}^{z_T} \nabla \cdot \rho \nabla dz.
\]

Since the last term of (6) is now expressed in the form of mass transport divergence, the divergence operator can be handled easily.

Similar modifications are made on the Richardson equation for \(w\) of (2.16a) in Paper I. A convenient alternate form is

\[
w = - \nabla \cdot \int_{z_k}^{z} \nabla dz - \int_{z_k}^{z} \left( \Delta z \right) \frac{1}{\gamma H} dz + \int_{z_k}^{z} \frac{1}{\gamma H} - dz.
\]

(7)

If level \(k\) is horizontal everywhere, \(J\) is evaluated as

\[
J = \nabla \cdot \nabla p - g \int_{z_k}^{z_T} \nabla \cdot (\rho \nabla) dz.
\]

(8)

However, if level \(k\) happens to be the mountain surface, the evaluation of \(J\) should be changed as follows:

\[
J = \nabla \cdot \nabla s - g(\rho w)_k - g \int_{z_k}^{z_T} \rho \nabla dz,
\]

(9)

where \(w_k = \nabla \cdot \nabla H\) by the lower boundary condition (2).
Substituting the upper boundary condition \( w = 0 \) at \( z = z_T \) into (7) yields the formula for \( B \),

\[
B = \left[ - \int_{z_H}^{z_T} \frac{1}{\gamma} dz + \frac{1}{\gamma} \int_{z_H}^{z_T} \frac{z_T}{p} \frac{Q}{T} \, dz - \nabla \cdot \left( \frac{1}{\gamma} \vec{V} \right) \right] \left/ \int_{z_H}^{z_T} \frac{1}{\gamma} \frac{dz}{p} \right. . \tag{10}
\]

A portion of the integration domain should be blocked out when the height of the mountains exceeds \( \Delta z \). This procedure is similar to the delineation of an irregular coastline for ocean circulation models. Details of the boundary treatment are described by Oliger, Kasahara, and Washington (1968).

From the fact that the thickness of the layer above the mountains is not necessarily equal to \( \Delta z \), minor corrections are necessary in the evaluation of the vertical derivatives of variables, limits of integrals, etc.

3. HEATING/COOLING FUNCTIONS

The heating/cooling rate \( Q \) in (7) consists of the following three parts:

\[
Q = Q_a + Q_c + Q_d. \tag{11}
\]

Radiative heating and cooling rate \( Q_a \). \( Q_a \) is further divided into two parts, \( Q_{as} \), the rate of heating due to absorption of the solar insolation by water vapor, and \( Q_{at} \), the rate of heating/cooling due to infrared radiation.

\( Q_{as} \) is evaluated using the formula by Mügge and Möller discussed in Paper I. It was later found that the scheme for computing \( Q_{at} \), presented in Paper I, was too time consuming. Therefore, Sasamori (1968) designed a simplified method by expressing the absorption functions of water vapor and carbon dioxide in the form of analytical functions. Sasamori's program computes the downward and upward fluxes of infrared radiation. The radiative cooling rate can easily be obtained by computing the flux divergence. At present we perform the radiation calculation once for every two hours of simulation time. The effects of clouds are taken into account in the radiation calculation by a method similar to one described by Manabe and Strickler (1964).

At present we assume that the distribution of clouds is a function of latitude only (Clapp, 1964). This means, for example, that the same cloudiness exists in a high over Siberia as in a low at the same latitude over the Atlantic. This assumption, therefore, effectively reduces the thermal contrast between lands and seas. For this reason, we are planning to use computed cloudiness in the radiation calculation.

Condensation heating rate \( Q_c \). In Paper I, we assumed that the atmosphere is completely saturated by water vapor and that the release of latent heat by condensation of water vapor takes place in regions of ascending motion. Since the actual atmosphere is not saturated by water vapor, the rate of condensation was reduced by a factor of \( E_f \), which is less than unity. Although the rate of condensation can be reduced artificially, this procedure can lead to an unbounded heating rate because of a physical instability. Thus, an upper limit of the heating rate had to be prescribed arbitrarily. To eliminate this undesirable constraint, we now compute the released latent heat of condensation based on the explicit prediction of water vapor field as follows:
\[ \frac{\partial (pq)}{\partial t} + \nabla \cdot (pq \mathbf{v}) + \frac{\partial}{\partial z} (pqw) = M + \rho E, \tag{12} \]

where \( M \) is the rate of condensation of water vapor and \( E \) is the vertical and horizontal diffusion of water vapor. When the predicted mixing ratio exceeds the saturation mixing ratio, we assume that the excess amount of water vapor condenses and releases latent heat. (We assume that saturation occurs in areas of 85 percent relative humidity with ascending motion.) Since the precipitation mechanism acts as a moisture sink, the evaporation of water vapor from the earth's surface is included in the model through the computation of the moisture flux by the lower boundary conditions, similar to the computation of sensible heat flux in the earth's boundary layer.

Heating rate due to thermal diffusion \( Q_c \). \( Q_c \) is further divided into two parts, \( Q_{dv} \), the rate of heating/cooling due to vertical diffusion of sensible heat, and \( Q_{dh} \), the rate of heating/cooling due to the horizontal diffusion of heat. The formulas for computing \( Q_{dv} \) and \( Q_{dh} \) are discussed in Paper I.

In previous experiments without the dynamical effect of orography, we prescribed a mean sea-level temperature distribution of the earth's surface as input data. In this procedure, there are three artificial constraints in addition to the lack of the dynamical effects of orography. One is that the sea-level temperature in mountainous regions is artificial since some means must be assumed in order to deduce a sea-level temperature from the surface temperature. The second is that the exchange of sensible heat between the atmosphere and the surface is poorly estimated. The third is that the influence of the daily variation of surface temperature is ignored.

In the present model, we divide the earth's surface into three parts: oceanic regions, continents, and regions where the surface temperature is below the freezing point (0°C) irrespective of whether it is over land or sea. The ocean surface temperature is prescribed as an observed climatological mean temperature for the appropriate season of the year. We compute the surface temperature over the land and ice areas based on the condition that there be no net heat flux at the atmosphere-ground interface. Over ice covered regions, we assume an upper limit of 0°C. The diurnal variation of the sun is also included.

4. RESULTS OF CALCULATIONS

We present here the results of two controlled experiments simulating January conditions, one with and the other without the dynamic effect of orography. The two-layer version of the model is used with a vertical height increment of 6 km. The model is global and has a horizontal increment of 5 degrees in both longitude and latitude. Both experiments were started from an isothermal atmosphere of 240K.

Figure 2(a) shows the mean sea-level pressure distribution averaged for the period of day 20 to day 35 in the case with no orography. Figure 2(b) is the same as Fig. 2(a) but for the case with orography. Without the dynamic effect of orography, we can see the tendency for highs over Siberia and North America to shift eastward, whereas with orography, the positions of both highs are confined only to the areas of the Himalayas and Rockies. Because of the eastward shift of the Siberian high, the Aleutian low is virtually nonexistent in the case with no orography. This latter finding seems to be a significant difference between the two cases. However, the
position of the Icelandic low is approximately the same in both cases, though
the center of the Icelandic low is also shifted eastward in the case of no
orography. In the Southern Hemisphere, the intensity of the high over
Antarctica is reduced by the effect of orography.

Figure 3 shows an observed mean January sea-level pressure distribution
prepared by Mintz (1965). By comparing Figs. 2(a) and (b) with Fig. 3, it
is clear that the inclusion of orography indeed improves the simulation of
a January climatology of the sea-level pressure distribution.

Figures 4(a) and 4(b) show the pressure distributions in the Northern
Hemisphere at 6 km in the cases without and with orography, respectively.
Figure 5 shows a mean observed 500 mb height contour over the Northern
Hemisphere for January, after Jacobs (1958). In Fig. 5, we see two dominant
troughs which lie along the east coasts of the North American and Asiatic
continents. There is also a minor trough over western Europe. Comparing
Fig. 5 with Figs. 4(a) and (b), we find that the trough over the east coast
of North America is correctly simulated in the orography case, but the
position of the trough over the Far East is shifted eastward about 40 degrees
in longitude for both cases. This failure to simulate the second trough over
Japan is difficult to explain at present. However, the third trough over
western Europe is correctly reproduced in the case of orography.

Figures 6(a) and 6(b) show mean distributions of east-west wind velocity
at 9 km averaged for day 20 to 35 for cases without and with orography,
respectively. The symbols F and S denote a local maximum and minimum,
respectively. A major difference between the two cases is found in the
vicinity of the Himalayas. In fact, as seen from Fig. 6(a), the intensity of
the zonal motion south of the Himalayas is stronger in the case of no oro-
graphy. The velocity distribution over the North and South Americas is almost
unchanged for both experiments.

In summary, we have shown here that the inclusion of orography indeed
improves the simulation of January climatology for the sea-level pressure
distribution. The simulation of a mid-tropospheric trough over the east
coast of North America is also better with orography. The dynamic effect of
orography appears to have little influence on the upper flow patterns except
in the vicinity of the Himalayas. It is a little bit puzzling that the
inclusion of orography does, as a matter of fact, reduce the intensity of the
zonal motion south of the Himalayas. It seems, therefore, that the thermal
effect is indeed a major factor in shaping the climatological mean state, and
the dynamical effect of orography helps bring the simulation of climatology
closer to reality.

In this paper, we hardly have space to discuss the simulation of
climatology in the Southern Hemisphere. We are also in the process of
simulating a July condition. We hope to discuss these aspects of simulation
at the meeting.

REFERENCES

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![Six Layer NCAR Model with Orography Diagram]

Fig. 1. Six-layer NCAR general circulation model with orography.
Fig. 2. Mean sea-level pressure distribution averaged for days 20 to 35 with a 5 mb contour interval. (a) is without orography and (b) is with orography.

Fig. 3. Observed mean sea-level pressure distribution for January (Mintz, 1965).
Fig. 4. Mean pressure distribution at 6 km over the Northern Hemisphere averaged for days 20 to 35 with a 5 mb contour interval. (a) is without orography and (b) is with orography.

Fig. 5. Mean 500 mb height contours over the Northern Hemisphere for January (Jacobs, 1958).
Fig. 6. Mean distribution of east-west wind velocity in m/sec at 9 km averaged for days 20 to 35 with a 5 m/sec contour interval. (a) is without orography and (b) is with orography.