On the Vorticity Budget and Vertical Velocity Distribution Associated with the Life Cycle of a Monsoon Depression

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ABSTRACT

This paper discusses the results obtained from a diagnostic study of a monsoon depression which formed in the northern part of the Bay of Bengal. The depression, while intensifying, progressed westward across India with a speed of about 5° longitude per day. The computed vertical velocity is in good agreement with the observed asymmetric distribution of rainfall around the depression. The presence of a low level of non-divergence (i.e., around 850 mb) is found to have a significant role in the dynamics of the monsoon depression.

The important result of the computed vorticity budget over the period of the intensification of the depression is the detection of a middle and upper tropospheric cyclonic vorticity depletion due to large-scale dynamics in the western sector of the depression. This result is rather unexpected because of the fact that the depression’s observed cyclonic vorticity increases, not only in the lower troposphere but also in the middle and upper troposphere while progressing westward. It has been shown that the presence of deep convective cloud activity in the western sector provides the necessary process to compensate the negative vorticity tendency in the middle and upper troposphere. Through a simple parameterization it has been shown qualitatively that the transport of subgrid-scale vorticity by deep convective clouds in the western sector is significant.

This mechanism of vertical transport of extremely rich boundary layer cyclonic vorticity by deep convective clouds is found to be essential for the intensification as well as for the westward movement of the monsoon depression.

1. Introduction

Monsoon depressions are the most important of the rain-producing disturbances of the Indian southwest monsoon. In a general classification of tropical disturbances, the monsoon depressions come in priority next to hurricanes, in the viewpoint of intensity and associated rainfall. These depressions form over the north Bay of Bengal and move in a west to northwest direction along the seasonal monsoon trough. The number of depressions that pass through the Indian subcontinent during each monsoon season bear an important implication on the total rainfall of the southwest monsoon season. The average frequency of these depressions is about 1.8 per month (Ramage, 1971, p. 45), but it is not unusual to have a higher frequency over a shorter period of time. For instance, a sequence of depressions formed in 1968 over the head of the Bay of Bengal which were approximately a week apart (i.e., on 26 July, 2 August and 10 August). The present study concerns the life history of the depression that formed on 2 August. The general synoptic features of monsoon depressions were studied before by several workers (Koteswaram and George 1958, 1960; Koteswaram and Rao, 1963; Pisharoty and Asnani, 1957; Ananthakrishnan and Bhatia, 1960). Recently, prognostic and diagnostic studies were also made utilizing numerical models (Rao and Rajamani, 1970; Das et al., 1971; Krishnamurti et al., 1975, 1976; Singh and Saha, 1976).

Generally tropical depressions over other parts of the world form and intensify over oceans and weaken as they travel over land areas. But one of the peculiarities of monsoon depressions is that in most cases they form very close to land and move inland while intensifying. Generally they attain peak activity over land. The case chosen for the present study is that of the low-pressure area which had formed over the northwest Bay of Bengal on 2 August 1968 and concentrated into a depression on 3 August. It deepened on 4 August and moved west across central parts of India, reached a peak activity on 6 August, and weakened over western parts of Gujarat on 8 August. In this paper, we will present the synoptic structure and dynamical characteristics of the depression as it propagated across India during the period 3–6 August. We will also attempt to explain the reasons for its intensification and westward movement with the help of the diagnostic nonlinear balance model.
2. Data set

The area of interest for this study is contained within 6–34°N and 56–108°E. Wind and temperature fields are analyzed over a larger area lying between 5°S to 45°N and 40°E to 120°E for levels 1000, 850, 700, 500, 300 and 200 mb. The charts are analyzed with 0000 GMT observations for 4 days from 3 August to 6 August 1968. Sea level pressure charts are analyzed for 0300 GMT, since this is the main synoptic hour for Indian stations. The data coverage is maximum at this time. The 24 h accumulated rainfall for the previous 24 h is also measured from 0300 GMT. Sea surface temperature data are obtained from the monthly mean charts. Data of smoothed terrain height above sea level are taken from Berkofsky and Bertoni (1960). The wind field is analyzed in the form of streamlines and isotachs. Rawinsonde data are augmented with the available data from pilot balloons and aircraft as well as making judicious use of satellite cloud-picture-derived winds. The moisture field is analyzed in the form of relative humidity up to 500 mb and nephanalysis is used as a help in this step. Over the extensive orographic region of the Himalayas and the Tibetan plateau, available data above 500 mb are supplemented with August climatic data. Below the 500 mb level the data are fictitiously fitted to maintain continuity with the surroundings; hence, all the results to be interpreted bear reality south of 28–30°N only. Grid-point data for the above-mentioned meteorological variables are subjectively interpolated at 2° latitude and longitude intersections from the analyzed charts. This forms our original data set for the numerical model.

The above-mentioned data set, some of the analyzed charts and some cross sections for the period 4–6 August have been used in the studies by Krishnamurti et al. (1975, 1976).

3. Synoptic features of the depression

Here we will deal only briefly with the synoptic description and bring out some salient features.

A low pressure area formed at the eastern end of the monsoon trough over India in the Bay of Bengal and adjoining land area on 2 August 1968. This low pressure area extended as a cyclonic circulation up to
500 mb. Examination of the data did not suggest movement of either the low pressure system from the Burma region or the passage of any wave disturbance in the upper tropospheric easterlies. During the 24 h period prior to the 2 August, there was extensive and heavy rainfall also associated with some thunderstorms reported over west Bengal and adjoining regions. Figs. 1–6 give the analysis of sea level pressure, circulation features through streamlines and isotach analysis for 850, 500, 300 (wind vector plots of grid-point data) and 200 mb levels, and recorded rainfall and weather code for the previous 24 h for a period of four days, starting from 3 August 1968. The following are some of the noteworthy features:

1) The depression is continuously intensifying over the 4-day period, while it moved westward across the Indian subcontinent approximately along 22°N latitude (Fig. 1). The westward speed of propagation is about 5° longitude per day.

2) The horizontal size of the depression or the half-wavelength of the disturbance is about 1000 km. The circulation covers an extensive region, ranging between $10^5$ and $10^6$ km². At the developed stage, there is some strengthening of the 850 mb winds over peninsular India and often the easterly winds to the north of the depression center also became strong. On 6 August, the lower level winds at Bombay became gusty and at 900 mb a 70 kt strong wind is reported (Fig. 2).

3) The depression extended in the vertical up to the 300 mb level as a cyclonic vortex on 3–5 August and up to the 250 mb level on the 6th. There is no noticeable tilt in the vertical of the cyclonic vortex up to 500 mb. On 3 August, there is no tilt in the vertical up to 300 mb, whereas on the 4, 5 and 6 August there is marked southward tilt of about 3–4° latitude (Figs. 2–5). Table 1 gives the position of the cyclonic vortex from our analysis for the four map times.

4) Associated with the cyclonic circulation at 300 mb, a wave formed in the easterlies on the 4, 5 and 6 August (Fig. 4). The trough and ridge axes have an orientation from NE to SW. This particular circulation pattern

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**Fig. 2.** 850 mb streamline and isotach (kt) analyses at 0000 GMT on (a) 3 August, (b) 4 August, (c) 5 August and (d) 6 August 1968.
has some significance for the vorticity advection at this level.

5) Except on 3 August, the subtropical anticyclonic belt is north of its normal position. On 2 August, under the influence of an extratropical westerly wave, the anticyclonic belt was 3–5° south of its normal position of 30–32°N. Thus there was some short period of oscillation in the planetary-scale circulation during the formation of the monsoon depression (Fig. 5).

6) The depression does not seem to extend up to the 200 mb level but its reflection as a wave in the easterlies can be seen. This wave is quite marked on 5 August with a wavelength of about 2000 km.

7) The observed rainfall pattern (Fig. 6) shows an asymmetric distribution with respect to the depression center (marked as Θ). The significant zone of rainfall always lay in the western and southwestern sectors of the depression. This is known to be the common feature for monsoon depressions (Fisharoty and Asnani, 1957). Generally, the rainfall amounts recorded during 24 h increased with the intensification of the depression. During the study period, the majority of the stations in the western sector of the depression reported rainfall amounts ranging from 5 to 15 cm day⁻¹. However, some individual stations in that sector reported rainfall amounts as high as 19, 21, 26 and 38 cm day⁻¹ on different days of the period of the depression's intensification. The convective phenomenon is seen as an integral part of the depression's history.

a. Moisture structure

The moisture content below 850 mb is generally high all over the country and the specific humidity values are around 18 g kg⁻¹. The northwestern desert regions are very dry with values around 5 g kg⁻¹. The main variations appear in the middle troposphere. The regions of high moisture content at these levels are purely associated with boundary layer convergence zones. The moisture is converging all over the region of the large-scale monsoon trough. Its maximum concentration is in the vicinity of the depression. The moisture convergence into a unit column of the atmosphere has increased threefold from 3 August to 6 August within the zone of the depression. On all four map times, the Arabian Sea and the extreme southern Bay of Bengal are the regions of moisture divergence.

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**Fig. 3.** As in Fig. 2 except for the 500 mb level.
b. Thermal structure

In order to view the thermal structure within the immediate proximity of the depression, as a general feature for the 4-day period, we composited the temperature anomaly; this is shown in Fig. 7 for the 900 and 500 mb levels. The temperature anomaly is the deviation from the area-averaged value (6–30°N, 56–108°E) at a pressure surface. The compositing with respect to the depression center is done for the four map times over a 6° radius surrounding the center. The basic thermal gradient is from south to north. At the 900 mb level, the depression has a marked cold core of about 2°C and is slightly to the east. The gradient is strong in the immediate vicinity to the north and northwest and very feeble toward the south and southeast. Pronounced warm advection is expected in the western sector and a feeble cold air advection to the east of the depression center. At the 500 mb level, the depression center is 1°C warmer than its surroundings and the gradient around it is very weak.

c. Distribution of the vorticity field

The absolute vorticity pattern is shown for the four map times along 22°N in the vertical plane in Fig. 8. It may be noted that the earth's vorticity at this latitude is $55 \times 10^{-8}$ s$^{-1}$. One will notice the strong core of absolute vorticity on all four map times coinciding with the depression's position at the surface. The pattern shows a strong concentration of isopleths of absolute vorticity in a zone of 4–6° of longitudinal width. The absolute vorticity is maximum at the surface and its magnitude varies between 120–150 $\times 10^{-8}$ s$^{-1}$. This strong core of absolute vorticity continues straight up to the 300 mb level. This suggests that the depression is a very intense cyclonic vortex up to the 300 mb level and its surface magnitudes are 1–2 times larger than the earth's vorticity at this latitude. On 3 August, the variation of vorticity with height is very small and it is maximum on 6 August. The core of strong cyclonic vorticity up to 300 mb systematically moves westward, while intensifying in association with the depression. Yanai and Nitta (1967) found, in the
Table 1. Approximate position of the depression center.

<table>
<thead>
<tr>
<th>Time/Date</th>
<th>Surface to 500 mb</th>
<th>At 300 mb</th>
</tr>
</thead>
<tbody>
<tr>
<td>0000 GMT 3 Aug 1968</td>
<td>23.0°N, 89.0°E</td>
<td>22.0°N, 89.0°E</td>
</tr>
<tr>
<td>0000 GMT 4 Aug 1968</td>
<td>22.5°N, 84.5°E</td>
<td>19.0°N, 83.5°E</td>
</tr>
<tr>
<td>0000 GMT 5 Aug 1968</td>
<td>22.5°N, 79.0°E</td>
<td>18.0°N, 78.0°E</td>
</tr>
<tr>
<td>0000 GMT 6 Aug 1968</td>
<td>23.5°N, 74.5°E</td>
<td>18.5°N, 73.5°E</td>
</tr>
</tbody>
</table>

The case of the easterly wave trough zone of the Caribbean Sea region, that the cyclonic vorticity varies in the vertical with a value of $10 \times 10^{-6}$ s$^{-1}$ at the surface to a maximum value of $40 \times 10^{-6}$ s$^{-1}$ at 600 mb and changes to an anticyclonic vorticity of $-20 \times 10^{-6}$ at 200 mb. Williams and Gray (1973) have noted for the western Pacific cloud clusters that the relative vorticity is of the magnitude $5-9 \times 10^{-6}$ s$^{-1}$ throughout the layer from the surface to the 400 mb level. The maximum anticyclonic vorticity is at 200 mb and its value is about $-2 \times 10^{-6}$ s$^{-1}$. In the case of the monsoon depression the respective magnitudes are $90 \times 10^{-6}$ s$^{-1}$ at the surface and $-4$ to $-10 \times 10^{-6}$ s$^{-1}$ at the 200 mb level. Thus, in conclusion, one may say that the monsoon depression, in the realm of large-scale to medium-scale disturbances, comes next to hurricanes.

4. Diagnostic model

Most of the previous numerical studies on monsoon depressions were done with the help of quasi-geostrophic models and without the diabatic heating effect especially due to convective clouds. As one can note from Figs. 2–5, the motion field associated with the depression is significantly nongeostrophic in character and the presence of convective clouds is an integral part of the cloud cover of the depression. Hence, to bring out the dynamical character of the depression, we have used a nonlinear diagnostic balance model developed by Krishnamurti (1968a).

The nonlinear balance model equations are as follows:

\[
\nabla \left( \frac{\partial \psi}{\partial t} \right) = -J(\rho, \psi) + \nabla \cdot \nabla \psi + \eta \nabla^2 \psi - \frac{\partial}{\partial \rho} (\nabla \psi) - \nabla \cdot \left( \frac{\partial \psi}{\partial \rho} \right) - \frac{1}{\rho} \frac{\partial}{\partial x} \left( \frac{\partial \tau_x}{\partial x} + \frac{\partial \tau_x}{\partial y} \right). \tag{1}
\]

Fig. 5. As in Fig. 2 except for the 200 mb level.
NONLINEAR BALANCE EQUATION

\[ \nabla^2 \phi - \nabla \cdot j \nabla \psi - 2J \left( \frac{\partial \psi}{\partial x}, \frac{\partial \psi}{\partial y} \right) = 0. \]  

EQUATION OF CONTINUITY

\[ \nabla^2 \chi = -\frac{\partial \omega}{\partial \rho}. \]  

THERMODYNAMIC EQUATION

\[ \frac{\partial \theta}{\partial t} = -\pi J(\psi, \beta) + \pi (\nabla X \cdot \nabla \theta) + \omega + \frac{R}{\rho C_p} H. \]  

The \( \omega \) equation can be derived from the above set of equations, and it is given by

\[ \nabla^2 \omega + f^2 \frac{\partial}{\partial \rho} \left( \frac{\partial \tau_x}{\partial x} + \frac{\partial \tau_y}{\partial y} \right) - 2 \frac{\partial}{\partial \rho} \left( \frac{\partial \psi}{\partial y} \right) \]  

\[ - \rho C_p \frac{\partial}{\partial \rho} \left( \nabla \cdot \nabla \psi \right) + \frac{\partial}{\partial \rho} \left( \nabla \cdot \nabla \eta \right) - \pi \nabla^2 (\nabla \cdot \nabla \theta) \]  

\[ - \beta \frac{\partial}{\partial \rho} \left( \frac{\partial \psi}{\partial \rho} \right). \]  

The symbols used above are defined in the Appendix. The \( \omega \) equation (5) contains 11 internal forcing functions which were described in detail by Krishnamurti (1968a). Some of the forcing functions which are of particular interest in this study are the first five terms on the right side of (5). They are, respectively, (i) differential vorticity advection, (ii) Laplacian of thermal advection, (iii) effects of frictional stresses, (iv) differential deformation effect, and (v) diabatic heating effects.

The solution of (5) is obtained by successive approximations which involve a three-dimensional relaxation procedure for \( \omega \) and a two-dimensional relaxation for \( X \) and \( \partial \psi / \partial t \). Certain boundary conditions are necessary in solving Eqs. (1), (3) and (5). At each level along the northern and southern boundaries, \( \omega \), \( X \) and \( \partial \psi / \partial t \) are set equal to zero and \( \omega \) is assumed to be zero at the upper boundary (100 mb level). A surface terrain effect is treated as an external forcing function at the boundary by defining \( \omega \) at the 1000 mb surface as

\[ \omega_{1000} = \frac{1000}{RT} [f(\psi,h) - \nabla X \cdot \nabla h], \]  

where \( h \) is the smoothed terrain height. This particular term is effective in the production of upward vertical velocity on the west coast. In addition to these boundary conditions, we had a cyclic boundary condition on the east-west direction.

Frictional stresses are defined at 1000 mb level and are assumed to vanish above the 900 mb level. The horizontal components may be written

\[ \tau_x = \rho C_p \left( \frac{\partial V}{\partial x} \right); \quad \tau_y = \rho C_p \left( \frac{\partial V}{\partial y} \right), \]

where \( C_p \) the drag coefficient is assumed to be 2.0 \times 10^{-3} for this study.

The forcing function due to heating [fifth term of Eq. (5)] is composed of three types of diabatic heating effects. Thus \( H \), the rate of heating per unit mass, is composed of

\[ H = H_S + H_L + H_C. \]

\( H_S \) is the divergence of the sensible heat flux over the sea. We used a bulk aerodynamic formula to parameterize this effect. During this case study, the surface temperature was 2-3° C warmer than the surface air temperature over the northern end of the Bay of Bengal and over the west coast of India. The flux of sensible heat is assumed to vanish at the 900 mb level. \( H_L \) is the diabatic heating effect due to latent heat of condensation released through stable clouds and \( H_C \) is the parameterized diabatic heating effect due to convective cloud phenomena. The procedure used in this study for parameterizing \( H_L \) and \( H_C \) was developed by Krishnamurti and Moxim (1971). Tests are made at each grid point for 1) rising motions at the top of the boundary layer due to the frictional forcing function, 2) large-scale moisture convergence in the vertical column, and 3) conditional instability in the boundary layer. If these three conditions are satisfied, then a heating function for large-scale motions is specified by parameterizing cumulus convection and is defined by the following equations

\[ H_C = a \left[ C_p(T_S - T)/\Delta t, \right. \]

\[ \left. - g^{-1} \int_{P_B}^{P_T} \left( \nabla \cdot qv + \partial q\omega / \partial \rho \right) d\rho \right. \]

\[ a = \frac{1}{(g \Delta t)^{-1} \int_{P_B}^{P_T} \left[ (C_p/L)(T_S - T) + (q_s - q) \right] d\rho}. \]

The parameter \( a \) represents the fractional area of the synoptic-grid-scale area covered by deep and active convective clouds. In Eq. (8) the denominator is the amount of moisture supply needed to cover the entire grid-scale area by a model cloud whose temperature \( T_S \) and humidity \( q_s \) are arrived following the moist adiabatic from the cloud base level. The numerator in Eq. (8) is the net amount of moisture supply actually available in the unit column. In our case, most of the moisture
Fig. 6. Past 24 h rainfall amounts (cm) and past 6 h weather reported at 0300 GMT on (a) 4 August, (b) 5 August, (c) 6 August and (d) 7 August 1968.

Convergence into the depression area is supplied through boundary layer convergence; hence the numerator can be simplified as $-(1/g)\omega_R \cdot b$. It is of interest to note that $H_C$ is independent of the cloud life period, whereas $a$ is dependent on it. The assumed value for $\Delta t$ in this study is 15 min. For further physical explanation on $H_C$ see Kuo (1965, 1974), Krishnamurti (1968b) and Krishnamurti et al. (1973). If the condition of rising motion is satisfied in stable layers and the relative humidity $\geq 80\%$, then the latent heat release due to nonconvective or stable type of clouds is calculated using

$$H_L = -L_o \frac{\partial q_s}{\partial \rho} \quad (9)$$

Here $q_s$ is the saturated specific humidity for the stable air at temperature $T$ of the level and the grid point in question. Formulation of the diabatic heating $H_C$ for the large-scale motion used in the present study is somewhat simple. But the scheme has received considerable application and is successfully utilized in tropical prediction and diagnostic experiments (Ceselski, 1973;
Krishnamurti, 1969; Ellsberry and Harrison, 1972; Rao and Hassebrock, 1972; Astling et al., 1972). In the present study heating due to radiative processes is neglected.

5. Results

The data set discussed in Section 2 is used as input for the numerical model of the earlier section.

The vertical motions are computed at five levels of the model, i.e., 1000, 900, 700, 500 and 300 mb. This computational procedure is repeated for the four map times which comprise the period of the early stage of the developed monsoon depression at 0000 GMT 3 August 1968, to its peak activity at 0000 GMT 6 August 1968.

a. Vertical velocity distribution

In the following we will refer to the vertical motion in the units of $10^{-5}$ mb s$^{-1}$, a value approximately equal to 0.01 cm s$^{-1}$ below the 500 mb level and to 0.02 cm s$^{-1}$ above the 500 mb level.

1) Horizontal distribution of $\omega$

At the 1000 mb surface, the vertical motion is only due to the terrain effect and in all four map times there is ascent over the western Ghats and over the Assam hills; the former is associated with westerlies and the latter with southerlies of the general monsoonal flow. At the 900 mb level, the total vertical motion, i.e., due to all forcing functions of Eq. (5), shows general
ascent over the whole region of the large-scale monsoonal trough zone and small pockets of descent over the Arabian Sea and over the southern Indian Peninsula (south of 16°N). The magnitudes are around −50 units in the general ascending area not associated with the depression and around +20 to +50 units in the descending region. This seems to be the normal monsoonal vertical circulation at lower levels (i.e., the lower limb of the reverse Hadley cell, an ascent in the monsoonal trough zone and descent in the southern regions). At the next level of our computation, the 700 mb surface, there is an additional descending region over northwest India associated with the intense thermal seasonal surface low. The general pattern of the horizontal distribution of \( \omega \) remains the same but the magnitudes are changing in such a way that horizontal divergence appears at 700 mb and continues up to at least 300 mb. In our computations 300 mb is the last level of computation for \( \omega \). This pattern suggests, in general, that a level of nondivergence exists between 700 and 900 mb over the most of the monsoon region. This particular characteristic seems to have some dynamical importance over the Asian monsoon region. The vertical velocity distribution associated with the monsoon depression is such that the upward motion is at least three times larger than the maximum ascending motion anywhere else on the map at the 900 mb surface. Its maximum ascent is always on the western or southwestern region of the depression center, except for the first map time on 3 August.

The figures for the horizontal distribution of \( \omega \) are not presented in this study.

2) Vertical cross sections for \( \omega \)

We present in Fig. 9 the total \( \omega \) distribution in the vertical for the four time periods of our study along latitude 22°N. We selected this particular zonal plane because the monsoon depression is progressing westward approximately along this latitude. The \( \omega \) distribution in the lower troposphere shows a center of maximum upward motion around the 900–850 mb level for all four periods. The maximum isopleth of upward motion is always present on the west side of the depression center, except for 3 August. The surface position of the depression is marked on the cross sections for easy reference. Only on 3 August is there another maximum upward motion zone, this one centered around 78°E. This is due to the presence of another depression which is weakening over western Uttar Pradesh. As the depression intensifies, the upward motion in the western zone is significantly increased in magnitude. It reached a peak magnitude of −320 units on the 6th from a mere −110 units on the 3rd. There is, however, an exception on 5 August, where this trend is not seen. There is no zone of significant downward motion in any of these cross sections. On 3 August there is another maximum of upward motion at 500 mb, right above the surface center of the depression. On 4 August at the 700 mb level, there are two pockets of upward
motion, both on the far eastern and western sides of the depression position. The significant phenomenon noticed associated with the depression is the shallower layer of convergence below the 850 mb level and a deeper layer of weak divergence above it. This results in a very characteristic feature of the monsoon depression with a low level of nondivergence approximately around the 850 mb level. Though our \( \omega \) distribution is strictly

Fig. 10. Vertical cross section of omega \((10^{-5} \text{ mb s}^{-1})\) along 22\(^\circ\)N due to Laplacian of thermal advection for (a) 5 August and (b) 6 August; and due to frictional forcing for (c) 5 August and (d) 6 August.

Fig. 11. Vertical cross section of omega \((10^{-5} \text{ mb s}^{-1})\) along 22\(^\circ\)N due to differential deformation forcing for (a) 5 August and (b) 6 August; and due to differential vorticity advection for (c) 5 August and (d) 6 August.
dependent on the physical processes in the numerical model, this level of nondivergence is in agreement with other studies (Rao and Rajamani 1970; Das et al., 1971). Moreover, in the kinematic method, following O'Brien's (1970) technique, we have noticed that the level of nondivergence is around 700 mb. Krishnamurti et al. (1975) also noticed the same feature with the kinematic method.

b. Contributions by various forcing functions

We now examine the physical processes that are responsible for the above-mentioned total $\omega$ distribution. In Figs. 10, 11 and 12, we present the contribution to $\omega$ due to individual forcing functions. We find, in general, that the lower tropospheric upward motion is chiefly contributed by the frictional forces. Frictional forces produce upward motion in the areas of cyclonic vorticity and hence around the depression the effect is to produce approximately equal upward motion in both the eastern and western zones. The warmer air advection by the northerly and northwesterly winds in the western zone brings upward motion comparable to the magnitude of $\omega$ due to friction. As the depression intensifies from 5 to 6 August, the contribution due to friction increased (Fig. 10) and it overpowered the descending motion due to cold air advection on the eastern zone of the depression. We have noticed from the composite of temperature anomaly (Fig. 7) that the thermal gradient is significant in the lower troposphere and has weakened at the 500 mb level. Hence, this is reflected in the weaker $\omega$ distribution at upper levels due to the thermal advection term.

The next significant contributory processes, in order, are differential deformation and vorticity advection effects. Because of the presence of the strong confluence zone west of the depression center on the 6th, the contribution due to this significant nongeostrophic process is well marked by $-75$ units of upward motion as a result of the deformation effect. Vertical motion due to the diabatic heating effect is the main contributor to the upward motion in the middle and upper troposphere (Fig. 12). The intense upward motion around 500 mb on 3 August (Fig. 9) is solely due to the latent heat effect. Examining the diabatic heating profile for the 3rd, we noticed that the stable type heating is more than the convective type and its maximum is centered around the 500 mb level. The convective heating rate is half of the stable type and its maximum is around 300 mb. Whereas on the 6th, though the depression is at its peak activity, the vertical motion due to the heating effect is not as significant as on the 3rd. On 6 August the diabatic heating rate is mainly due to convective type and its magnitude is twice that of the 3rd at 300 mb. We feel, in the case of the 3 August, that the combination of both types of heating, especially that of the stable type, is responsible for the greater upward motion at the 500 mb level. Synoptic observations support the existence of both stratiform and convective type clouds on the 3rd over that region. On the 4th, $H_L$ is more than $H_C$ and the $\omega$ is maximum ($-75$ units) at the 700 mb level. On 5 August, $H_C$ is more than $H_L$ and the $\omega$ is maximum ($-50$ units) at the 300 mb level. Since the parameterization of $H_C$ does not include the effects of evaporative cooling due to falling rain drops or detrainment of cloud droplets, we presume that the values of $H_C$ and $\omega$ are slightly overestimated in our study.

In other tropical disturbances, the effect of latent heat release due to convection is the main contribution to large-scale vertical velocity at upper levels. At lower levels, the other forcing functions do not contribute significantly enough to produce larger $\omega$ at low levels (Astling et al., 1972). Hence there exists a deep layer of convergence and a higher level of nondivergence. But in the case of the monsoon depression, in spite of the role of convection to provide a source of significant upward vertical velocity at the upper level, the maximum $\omega$ is found around the 850 mb level. This is mainly due to the stronger influence of the baroclinic and dynamical effects, rather than to the latent heat effect associated with the depression.

c. Computed vertical motions vs. the observed cloud and rainfall pattern

During the well-developed states of the depression on 4–6 August, the observed asymmetric distribution
of rainfall with respect to the depression center (Fig. 6) is in very good agreement with the computed vertical velocity pattern (Fig. 9). On 3 August, the distinct patches of rainfall are only marginally correlated with the two centers of maximum vertical velocity in the zonal plane, but the asymmetry in the rainfall zone is not brought out with the computed vertical velocity pattern. The reason is difficult to evaluate. It may be due to the data paucity on the east side of the disturbance or to the fact that the depression is itself in an organizational period.

Fig. 13 shows the cloud pictures taken by the ESSA 5 satellite for the 3, 5, 6 and 7 August 1968. There is no satellite picture for the 4th. The observed cloud pictures are at approximately 1200 GMT on each day. As our map times are on 0000 GMT, the computed results lag the cloud pictures by 12 h. The common feature of the pictures is the presence of an extensive overcast to the west and southwest of the center of the depression and a relatively clear region to the immediate east. Generally, the center is located at the northeastern edge of the overcast area. On 5 and 6 August, we can
see the spiraling of convective bands toward the center of the depression.

The fractional area of deep and active convective clouds occupied in a grid-scale area is computed by using Eq. (8); the results for the four map times are depicted in Fig. 14. The values are expressed in percentage. On 3 August, about 2% of the grid-scale area is occupied by deep convective clouds in the region of the weakening depression over Uttar Pradesh. Over the region of the developing depression around 90°E, 22°N, the value is around 0.5%. On the next map time, the fractional area shifts from the earlier weakening depression area to the newly developed depression on 4 August. From the surface charts the depression center is known to be at 84°E, and 22°N. As the depression intensifies and progresses westward, the fractional area increases in magnitude and the general envelope of the area contours moved westward accordingly. The increase of a to 4% on the 6th also agrees with the higher intensity of rainfall recorded at a few stations on that day. As the satellite pictures are 12 h ahead of our computed cloud cover figures, one can visually shift the satellite pictures by 2–3° longitude toward the east while comparing. The distribution of a did not show the expected asymmetric pattern with respect to the depression center. This is because the moisture convergence used to compute a utilizes only fractionally derived \( \omega \). This will be remedied in future experiments.

6. Vorticity budget

Along with the problem of intensification, the movement of the depression is also an interesting dynamical problem. The analysis of the vorticity budget helps to resolve some important characteristics peculiar to monsoon depressions.

In quite a few budget studies, the sum of all the large-scale terms in the vorticity equation is evaluated to find the residual which is presumed to represent the effects of convection. Here we first compare the observed local vorticity tendency with the computed tendency due to the large-scale terms of the vorticity equation and then assess the necessity of invoking convective cloud effects. Then we explicitly include the effects of convective clouds in the budget equation. We will examine the behavior and the contribution due to these effects to the large-scale vorticity tendency in the latter part of this section.

The complete vorticity equation can be written in the form

\[
\frac{\partial \zeta}{\partial t} = -\mathbf{V} \cdot \nabla \eta + \mathbf{\omega} \cdot \nabla \frac{\partial \tau_v}{\partial p} - \nabla \left( \frac{\partial \tau_v}{\partial x} \frac{\partial \tau_z}{\partial y} \right) - \nabla \left( \frac{\partial \tau_z}{\partial x} \frac{\partial \tau_v}{\partial y} \right) - \nabla \left( \frac{\partial \tau_z}{\partial x} \frac{\partial \tau_z}{\partial y} \right)
\]

\[
-\frac{\partial \zeta}{\partial p} - \mathbf{k} \cdot \nabla \left( \frac{\partial \mathbf{V}}{\partial p} \right) \frac{\partial \zeta}{\partial p} - \left( \frac{\partial \zeta}{\partial t} \right)\;\;\;\; (10)
\]
Table 2. Areas used for averaging the computed terms of vorticity equation following the passage of the depression.

<table>
<thead>
<tr>
<th>Time/Date</th>
<th>Western sector</th>
<th>Eastern sector</th>
</tr>
</thead>
<tbody>
<tr>
<td>0000 GMT 3 Aug</td>
<td>82–90(^\circ)E</td>
<td>92–98(^\circ)E</td>
</tr>
<tr>
<td>0000 GMT 4 Aug</td>
<td>78–86(^\circ)E</td>
<td>88–94(^\circ)E</td>
</tr>
<tr>
<td>0000 GMT 5 Aug</td>
<td>70–78(^\circ)E</td>
<td>80–86(^\circ)E</td>
</tr>
<tr>
<td>0000 GMT 6 Aug</td>
<td>68–76(^\circ)E</td>
<td>78–84(^\circ)E</td>
</tr>
</tbody>
</table>

where \( \mathbf{k} \) is the unit vector in the vertical. The last two terms on the right side of Eq. (10) are the effects of the convective cloud ensemble on the large-scale vorticity budget. The other terms are all well known. The computed \( \omega \) from our numerical model is utilized in evaluating the terms that involve \( \omega \) in Eq. (10). We will present the results of the effects of each term in the western and eastern sectors of the depression following its westward passage.

The zone occupied within the latitudinal belt 16–24\(^\circ\)N is divided into western and eastern sectors with respect to the center of the depression by choosing 8\(^\circ\) of longitudinal width on either side. Table 2 gives the exact areas utilized in averaging the computed vorticity terms following the westward passage of the depression.

Fig. 15 shows the vertical distribution of the rate of change of relative vorticity due to each one of the first five forcing terms averaged over the respective sector areas. These are presented for all the 4 days of the depression's history.

We will discuss the vorticity tendency profiles in two stages: 1) lower tropospheric features up to 850 mb and 2) middle and upper tropospheric features up to 200 mb.

a. Lower tropospheric features

The total vorticity tendency (i.e., the net effect of the first five forcing terms) is positive in the lower troposphere and is maximum at the surface in the western sector of the depression at all four map times. It is negative in the eastern sectors throughout the period. Thus, we note the continuous intensification of the cyclonic vorticity in the western sector and the decay of the same in the eastern sector. This explains the observed westward movement of the monsoon depression at the surface level. Since the lower tropospheric flow during this period is a strong westerly current, some explanation is needed for the movement of the depression against the zonal flow.

Let us examine the role played by each term in the vorticity equation. The horizontal advection of the
absolute vorticity is negative in all four cases, both in the western sector as well as in the eastern sector. In the western sector, the monsoonal westerly flow is advecting lower values of absolute vorticity and it also reveals that the $\beta v$ term is overpowered by the horizontal advection of relative vorticity. Incidentally, this explains that the $\beta$ effect is not the cause for the westward movement in the case of the monsoon depressions, whereas for other tropical disturbances this is found to be effective. In the eastern sectors, the southerlies are advecting lower values of vorticity (Fig. 2) and hence the cyclonic vorticity is decaying in those zones. As the depression intensifies from 3 to 6 August, the westerlies become stronger in the western sector and as a result the effect of the first terms reaches a higher negative value. The strength of the westerlies and correspondingly the effect of negative vorticity advection decreases with the height.

The second term of the Eq. (10), the so-called divergence term, is computed using the $\omega$ distribution from our numerical model. This term is effectively a production term for the cyclonic vorticity and is positive throughout the period in both sectors of the lower troposphere. The effect of the divergence term is such that it not only compensates the negative tendency due to horizontal advection but it also significantly predominates the latter by 2–3 times in magnitude. The magnitude of the term generally increases as the depression intensifies and reaches a high magnitude at the peak stage of the depression. Physically this is due to the strong convergence and its development in the western sector. The reasons for stronger vertical motion and its associated convergence in the lower troposphere have already been discussed in an earlier section. The positive tendency for the vorticity due to this term even in the eastern sector, though small in magnitude, can be easily understood by referring to the vertical distribution of $\omega$ east of the depression center (Fig. 9). The upward vertical velocity is increasing with height in the lower troposphere. The cyclonic vorticity production due to this term steadily decreases with height, reflecting the nature of the convergence associated with the depression. Due to the absence of a deep layer of convergence and to the presence of a lower level of nondivergence, the effect of convergence on the vorticity tendency changes its positive contribution to a negative one above 800 mb. The other significant term for the vorticity tendency in the lower troposphere is due to the third term, the frictional stresses. The frictional stresses are assumed to vanish at the 900 mb level in our model. Hence the nature of the differential frictional stress in the vertical always produces negative vorticity tendency at the surface. We extrapolated arbitrarily its contribution exponentially in the vertical to a zero value around the 900 mb level. In both the sectors the cyclonic vorticity decreases with time due to this effect. The negative vorticity tendency is maximum at the time of peak activity of the depression. This is simply due to the fact that the vorticity tendency is generally proportional to the vorticity itself through the type of frictional stress formulation. But as we noted earlier the frictional force is one of the main contributors for upward motion in the lowest layer. Thus one may say that the frictional force indirectly provides the mechanism through the divergence term for the production of the cyclonic vorticity.

The vertical advection and tilting [fourth and fifth terms of Eq. (10)] terms have very insignificant contributions to the vorticity budget. In most of the cases their magnitudes were $10^{-10}$ to $10^{-13}$ s$^{-2}$. Only on 5 August (Fig. 15c) did these two terms show some contribution.

Utilizing a quasi-geostrophic model Rao and Rajamani (1970) computed the effect of the horizontal advection term and the effect due to the corresponding divergence term ($f \partial \omega / \partial p$) at a single 850 mb level for another synoptic situation. Although they achieved the correct direction of movement of the depression, the intensity of the vorticity tendency must have been underestimated. The reason is that in reality, the divergence term, the dominant one, is at least 2–3 times larger than its quasi-geostrophic counterpart. From a quasi-geostrophic prediction experiment Krishnamurti et al. (1976) found that the westward propagation of the depression is extremely slow.

Now summarizing the features in the lower troposphere, we note that the divergence term is the significant term for production of cyclonic vorticity in the western sector. It not only compensates the negative vorticity tendency due to the advective term but predominates in magnitude over the total contribution of all the other terms of the equation. In the eastern sector horizontal advection of vorticity individually dominates the effect of the divergence term in producing the negative vorticity tendency. One interesting feature that is well recognized is that generally with stronger westerly monsoon current the depression’s western sector is observed to have heavier rainfall rates than when the monsoon current is not so strong. This is observed during the period 5–6 August. The westerlies and the rainfall rates on the 6th are observed to be stronger and heavier than on the 5th. This phenomenon can be explained very clearly from our computations. Since the system is moving westward the divergence term must dominate the combined effects of the advective and the frictional terms. Due to the increased strength of the westerlies on the 6th, the advective and the frictional terms are greater than they were on the 5th. Since the depression is moving at a constant speed the divergence term on the 6th must be greater than it was on the 5th. Thus the required upward motion at lower levels is greater than that of the 5th and hence produces the heavier rainfall rate.
The monsoon depression exists on all the four map times as a strong cyclonic vortex well beyond the middle tropospheric level (Fig. 8). Hence it is of vital importance to know the dynamical behavior of the vorticity tendency at these levels.

b. Middle and upper tropospheric features

The divergence term provides a negative vorticity tendency during the whole 4-day period in both sectors in the middle and upper troposphere. This is because of the presence of the deep layer of divergence in this region of the atmosphere. The negative contribution due to this term is stronger in the western sector than in the eastern sector.

The lower tropospheric monsoonal westerlies steadily decrease with height and are replaced by the easterlies over southern India around 400 mb. In the western sector the process of positive vorticity tendency due to the effect of the horizontal advection term during 4 and 5 August (Figs. 15b,c) and negative vorticity tendency on 3 and 6 August (Figs. 15a,d) can be understood by examining the circulation features at 500 mb (Fig. 3). At 300 mb the depression has a more southerly position (about 18°N) and the circulation is dominated by strong easterlies (Fig. 4). As noted earlier in Section 3 there is a short-wave pattern in the easterlies with its trough axis passing through the cyclonic circulation center and the axis of the ridge to the east of it. The axes of both trough and ridge are inclined northeast to southwest. This particular circulation pattern is the cause for the positive vorticity advection into the western sector and negative vorticity advection into the eastern sector during 4, 5 and 6 August. The resultant total vorticity tendency is such that there is destruction of cyclonic vorticity in both sectors during this 4-day period in the layer of the middle and/or the upper troposphere. As explained earlier the large-scale circulation features and the \( \omega \) distribution ensured this property during the study period. But this result is rather unexpected in the western sector. Now if we examine the observed vorticity cross sections along 22°N (Fig. 8) during all four phases of the depression, there is a strong core of absolute vorticity extending up to the 300 mb level. In other words, in the western sector there should be strong continuous development of cyclonic vorticity, not only in the lower troposphere but up to 300 mb. The continuous depletion of cyclonic vorticity both at the surface and at the upper levels in the eastern sector is very consistent with the observed vorticity pattern. Hence we have to invoke a physical process which provides a compensatory mechanism for the vorticity loss in the middle and upper troposphere in the western sector. This particular physical process should be active and dominant in the western sector.

Our computation in the previous section on the fractional area of the deep and active convective clouds, the observed extensive cloud cover and the reported rainfall rates in the western and southwestern regions of the depression strongly support the following process.

The vertical transport of boundary layer cyclonic vorticity by convective cloud phenomena is the only conceivable process that explains the necessary production of vorticity and compensates for the loss of cyclonic vorticity in the middle and upper troposphere. This sort of mechanism has been put forward as necessary by Holton and Colton (1972) for the observed quasi-stationary vorticity pattern of the 200 mb level planetary-scale summer tropical circulation. A similar physical process is suggested for the required vorticity balance around convective cloud clusters associated with the equatorial waves in the study by Williams and Gray (1973). Reed and Johnson (1974) identified this and attempted to diagnose the cloud mass flux and the cloud vorticity distribution from the large-scale vorticity budget. Both studies were made with composited data sets. Our study seems to be the first attempt to show the necessity of such a vertical transport by convective phenomena in the case of a transient medium-scale disturbance in the monsoon region, i.e., the monsoon depression.

Before we discuss the parameterization of the effects of convective cloud phenomena on the large-scale vorticity budget, we substantiate our finding by showing the observed and the computed vorticity tendencies in Table 3. For the computation of tendencies we chose the grid point in each case such that it is on the west side of the depression center at the initial time and at the center of the depression at the final time. The table is self-explanatory. Briefly, the observed tendency showed significant vorticity generation at almost all levels; nevertheless, the calculated tendency caused by large-scale forcing terms showed vorticity depletion at middle and upper tropospheric levels.

c. Subgrid-scale contribution

We now examine the last two terms of the vorticity tendency equation. Following Reed and Johnson (1974) and Fraedrich (1973) the last two terms can be
written as

\[ \frac{\partial \zeta'}{\partial \xi} = -a \omega \frac{\partial}{\partial \eta} (\zeta_e - \zeta), \]  

(11)

\[ \frac{\partial \zeta_e}{\partial \eta} = -a (\zeta_e - \zeta). \]  

(12)

Some of the assumptions involved are as follows: 1) the contribution by subgrid-scale convective clouds through the tilting term is neglected; 2) the net horizontal eddy flux divergence of vorticity over a cloud ensemble average area (i.e., grid square area) is neglected; and 3) the parameter \( a / \Delta t \) is assumed to be constant in the vertical.

Eqs. (11) and (12) express, respectively, the contribution to the large-scale vorticity tendency through the statistical average of the vertical transport of vorticity by a group of convective clouds and the transient cloud effect which is also known as the buildup of vorticity within the cloud ensemble during its life cycle. Here \( a \) is the fraction of the grid-scale area occupied by deep convective clouds, and \( \omega \) and \( \zeta_e \) are the mean vertical \( p \) velocity and the mean relative vorticity of the whole group of convective clouds at a level respectively. \( \Delta t \) is the average half-life span of the convective clouds and it is generally of the order of 15–30 min. The parameterizability of the vorticity transport by deep convective cloud elements crucially depends upon the condition that the large-scale variation of vorticity does not change significantly over the period of the cloud life cycle. In other words, it is assumed that the convective activity is in statistical equilibrium with the large-scale motion. This criterion is generally met in this transient monsoon depression case, because the period of the large-scale changes is of the order of more than a day. Now these new variables \( \omega \) and \( \zeta_e \) must be determined as a function of the large-scale dynamic and thermodynamic variables either empirically or through cloud modeling. The determination of cloud mass flux \( -a \omega \) through heat and moisture budgets is not favored because of inadequacies in data and high empiricism involved with the stable type of heating. The evaluation of cloud ensemble vorticity and its vertical variation is a difficult theoretical problem. Use of \( \omega \) from cloud models and the application of conservation of potential vorticity equation to the cloud and its environmental motions may be the rational approach to this problem. In the present study we do not intend to compute these terms [Eqs. (11) and (12)] systematically, but we would like to see their approximate behavior in our case study.

Reed and Johnson (1974) evaluated cloud vorticity by determining subgrid-scale effects as a residue of the computed large-scale vorticity budget equation. They found in the case of deep cloud activity in the easterly wave trough region that the cloud relative vorticity increases from the cloud-base level to a maximum at 350 mb and then decreases on up to a level of 100 mb.

The cloud vorticity remained cyclonic throughout with a magnitude 2–3 times larger than the absolute magnitude of the environmental relative vorticity over a large part of the atmosphere. The vertical distribution of \( \zeta_e \) is such that it increases with height in the deep layer of large-scale convergence and decreases with height in the layer of divergence. Since \( \omega \) is always negative, Eq. (11) contributes to large-scale cyclonic vorticity depletion in the layer where the excess vorticity is increasing with height and to large-scale cyclonic vorticity generation in the layer where the excess vorticity is decreasing with height. If we suppose in our case study that the vertical distribution of \( \zeta_e \) follows that of Reed and Johnson (1974), i.e., \( \zeta_e > \zeta \) and the excess vorticity increases in the region of convergence and decreases in the region of large-scale divergence with height, then we will have negative vorticity tendency in the layer up to 850 mb and positive vorticity tendency throughout the layer up to 200 mb. This simply explains the removal of cyclonic vorticity from low levels and its deposition above the level of nondivergence by convective cloud phenomena.

The right-hand side of Eq. (12) shows that the accumulated excess of vorticity during the growing phase of convective clouds is distributed to the environment through the process of mixing in the period of their decay. Hence this term contributes to positive vorticity tendency in the layers where convective clouds exist with an excess of vorticity. Although \( \zeta_e \) is found to be more than \( \zeta \) in the earlier discussion, we assumed as a first approximation that \( \zeta_e \) is of the same magnitude as \( \zeta \) at the cloud-base level (i.e., \( \sim 950 \) mb) and it is assumed to remain constant throughout the atmosphere. In Table 4, we give the magnitude of the contribution to large-scale vorticity tendency due to this term. The magnitudes are all significant. However, the effective tendency due to subgrid-scale phenomena is the net effect of the two terms in Eqs. (11) and (12). The contributions due to these terms discussed above are highly suggestive and speculative in nature. But we have shown the definite necessity of the parameterized convective transport of vorticity in the large-scale dynamics for the westward movement and for the intensification of the cyclonic vortex associated with the monsoon depression.

<table>
<thead>
<tr>
<th>Level (mb)</th>
<th>3 August</th>
<th>4 August</th>
<th>5 August</th>
<th>6 August</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>84°E, 22°N</td>
<td>80°E, 22°N</td>
<td>76°E, 22°N</td>
<td>70°E, 22°N</td>
</tr>
<tr>
<td>a = 0.7%</td>
<td>0.0</td>
<td>0.31</td>
<td>0.55</td>
<td>2.41</td>
</tr>
<tr>
<td>700</td>
<td>0.8</td>
<td>0.14</td>
<td>2.03</td>
<td>3.29</td>
</tr>
<tr>
<td>500</td>
<td>0.54</td>
<td>0.16</td>
<td>1.91</td>
<td>3.21</td>
</tr>
<tr>
<td>300</td>
<td>0.57</td>
<td>4.4</td>
<td>2.26</td>
<td>4.71</td>
</tr>
<tr>
<td>200</td>
<td>2.3</td>
<td>8.6</td>
<td>7.55</td>
<td>5.96</td>
</tr>
</tbody>
</table>
In other tropical disturbances through the process of CISK in the disturbed regions, the net effective role of convection is to produce a deep layer of large-scale convergence and through the effect of large-scale convergence, the large-scale vorticity increases in the lower troposphere and in the layers of convergence (Charney, 1973). But in the case of the monsoon depression, convection is ineffective in producing the deep layer of large-scale convergence, and large-scale dynamical forces seem to be inadequate to generate cyclonic vorticity in the middle part of the troposphere. Hence convection must be involved but through a different mechanism, that is, by transport of rich boundary layer cyclonic vorticity to the layers in need of it.

These phenomena need further research with a systematically better parameterization of the vertical transport of momentum by deep convective cloud activity. The forthcoming monsoon experiment (MONEX), a part of the Global Atmospheric Research Programme, with more reliable data sets and higher resolution of the horizontal and vertical dimensions will be most helpful in diagnosing the bulk properties of convective clouds and their effect on the vorticity budget of any monsoon depression during all phases of its intensity.

7. Conclusions

The computed vertical velocity distribution is in agreement with the region of overcast cloudy area. The maximum intensity zone of upward motion coincides with the heaviest rainfall region in the western sector of the depression.

Absence of a deep layer of convergence and, hence, the presence of a low level of nondivergence around 850 mb is a significant characteristic of the depression.

The presence of upward motion at lower levels is mainly due to frictional forces and other baroclinic and dynamical forces.

The role of friction is found to be very important. By nature it impedes the movement of the depression toward the west and dissipates the cyclonic vorticity in the forward sector. But it indirectly helps through its significant contribution to $\omega$ at lower levels and thereby provides positive vorticity contribution through the convergence effect. It also provides the mechanism for the maintenance of convection through moisture convergence.

From the computed vorticity budget, it has been shown that in the lower troposphere (i.e., below 850 mb) the convergence effect provides the positive vorticity tendency in the western sector.

The most important and significant result of this study is the detection of the middle and upper tropospheric cyclonic vorticity depletion with time due to large-scale dynamics in the western sector of the depression in its day-to-day progression. This result is rather unexpected because of the fact that the depression's observed cyclonic vorticity increases, not only in the lower troposphere but also in the middle and upper troposphere, day after day in the western sector.

It has been hypothesized that the presence of deep convective cloud activity in the western sector provides the necessary physical process to compensate the negative vorticity tendency and provides the source of required cyclonic vorticity in the middle and upper troposphere. With a simple parameterization, we have shown qualitatively that the transport of subgrid-scale vorticity by deep convective clouds in the western sector is significant.

In other tropical disturbances, the net result of the role of convection is to increase the large-scale upward motion with height and thereby increase cyclonic vorticity with time at lower levels (Charney, 1973). But in the case of monsoon depression the role of convection is quite different. The convective phenomenon has to transport vertically the intense boundary layer cyclonic vorticity to middle and upper tropospheric levels against dissipation by large-scale forces. This mechanism is found to be very essential, without which there is no physically conceivable process at present, to keep the monsoon depression intensifying and to provide the process for its westward movement during its life cycle.

In the recent monsoon circulation simulation experiments (Washington and Daggupaty 1975; Hahn and Manabe 1975) the development and movement of monsoon depressions are not properly simulated. We feel that one of the reasons for the failure is probably the lack of the above mentioned physical process in their general circulation models.

It is now possible to postulate from our study that the parameterization of cumulus ensemble effects for diabatic heating and moisture supply to the large-scale system alone is not sufficient for proper prediction of monsoon depression. However the incorporation of those effects seems to be sufficient for the prediction of other tropical disturbances. Hence, more emphasis should be given to the parameterization of the effects of momentum transports by deep convective clouds. A numerical prediction experiment for the monsoon depression with the inclusion of the above mentioned physical process is in progress and the results will be reported in due course.

Acknowledgments. This whole project was carried out in different stages at three places: Florida State University, the National Center for Atmospheric Research, and the University of Toronto.

We thank Prof. T. N. Krishnamurti for giving us the opportunity to use his diagnostic model program. One of the authors (S.M.D.) thanks Prof. R. List for his encouragement in completing this project. The portion of the study carried out at the University of Toronto was supported by a NRC Negotiated Development Grant on Atmospheric Dynamics.
APPENDIX

List of Symbols

\( x, y, p, t \)  
independent variables—zonal, meridional, vertical pressure and time coordinates, respectively

\( a \)  
fractional area covered by deep and active convective clouds

\( C_p \)  
specific heat at constant pressure

\( f \)  
Coriolis parameter

\( \beta \)  
northward variation of \( f \)

\( g \)  
acceleration due to gravity

\( h \)  
terraheight above sea level

\( H \)  
rate of heating per unit mass of air

\( L \)  
latitude of heat of vaporization of water

\( P_B \)  
pressure at cloud-base level

\( P_T \)  
pressure at cloud-top level

\( q \)  
specific humidity per unit mass of air

\( q_s \)  
saturation specific humidity

\( R \)  
gas constant of air

\( T \)  
air temperature

\( T_s \)  
saturation temperature of air

\( V \)  
horizontal wind vector (\( u, zonal \); \( v, \) meridional)

\( \omega \)  
vertical velocity in pressure coordinate

\( \psi \)  
streamfunction

\( \chi \)  
velocity potential

\( \eta \)  
vertical component of absolute vorticity

\( \zeta \)  
vertical component of relative vorticity

\( \phi \)  
geopotential

\( \theta \)  
potential temperature of air

\( \pi \)  
\( RT/\rho \)

\( \tau_{xx}, \tau_{yy} \)  
frictional stresses

\( \nabla \)  
del operator

\( \nabla^2 \)  
Laplacian operator

\( J \)  
Jacobian operator

\( \Delta t \)  
half-life period of convective clouds

\( \sigma \)  
static stability parameter \( = -\pi \partial \theta/\partial p \)

\( \rho \)  
density of air

Overbars represent averages of a variable over a grid square, primes represent the deviation of a variable from that average.

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