Indian J. Met. Geophys (1971), **22**, 331-336

551.576 : 551.576.4(54)

Diagnostic study of vertical motion *vis-a-vis* large scale cloud systems

P. K. DAS, R. K. DATTA and B. M. CHHABRA

Northern Hemisphere Analysis Centre, New Delh

ABSTRACT. In this paper we present results of vertical velocities computed for a monsoon depression over India. The vertical velocities were calculated at intervals of 100 mb from 1000 to 100 mb with the help of a 10-layer geostrophic model. The computed omega field has been compared with the composite nephanalysis received from APT bulletins from Bombay.

It is found that the field of vertical motion computed by the 10-layer model agrees fairly well with the configuration of large scale cloud systems.

The effect of surface friction and topography was included as a lower boundary condition on the 10-layer model. The vertical motion at the top of the friction layer was computed on the basis of the Ekman spiral. The paper discusses the results of the other types of frictional convergence and their effect on the omega field.

1. Introduction

If we use the conservation principle for vorticity and the first law of thermodynamics, then we can derive a single equation in which the vertical component of velocity (omega) is the only dependent variable. This can be solved as a boundary value problem for omega.

In this paper we present the omega field associated with a monsoon depression over the Gangetic plains of India. It is our object to see how far the computed omega values using quasi-geostrophic model discussed below, agree with the areas of large scale precipitation and cloud cover as revealed by the satellite pictures over the subtropical areas.

2. Basic equations of the model

The method of computation and the basic equations are similar to those outlined by Das (1962).

We express the conservation of vorticity and the first law of thermodynamics by the quasi-geostrophic equations :

$$\nabla^2 \phi_t + f\,\mathbb{V}\,\nabla \eta = f^2 \omega_p \qquad (2\cdot1)$$

$$\phi_{pt} + \mathbb{V}\,\nabla \phi_p = -\sigma \omega \qquad (2\cdot2)$$

where η stands for absolute vorticity, σ is the stability parameter

$$\left[\; -\frac{1}{\rho} \frac{\partial}{\partial p} \ln \theta \; \right],$$

θ represents potential temperature, f is the Coriolis parameter, ϕ represents geopotential and ω is the vertical pressure velocity. We have used subscripts to denote derivatives.

On eliminating ω we find

$$L\left(\phi_t\right) = -f\,\mathbb{V}\,\nabla\eta - \frac{\partial}{\partial p}\left(\frac{f^2}{\sigma}\,\mathbb{V}\,\nabla\phi_p\right) \qquad (2\cdot3)$$

$$\text{where} \quad L \equiv \left[\, \nabla^2 + \frac{\partial}{\partial p}\left(\frac{f^2}{\sigma}\frac{\partial}{\partial p}\right) \right] \qquad (2\cdot4)$$

The principal assumptions, which we have made in deriving $(2\cdot3)$ are :

(*i*) The vertical advection of vorticity and the twisting term in vorticity equation have been neglected, because they are smaller than the other terms.

(*ii*) Space derivatives of the Coriolis parameter, f, have been neglected.

(*iii*) We have not considered diabatic heating or cooling in the first law of thermodynamics.

It will be observed that equation $(2\cdot3)$ can be solved as a boundary value problem for ϕ_t. When the computed values of ϕ_t are substituted in $(2\cdot2)$ we obtain the corresponding field of ω.

The conditions which enable us to make the quasi-geostrophic assumption may be summarised as follows (Phillips 1963) :

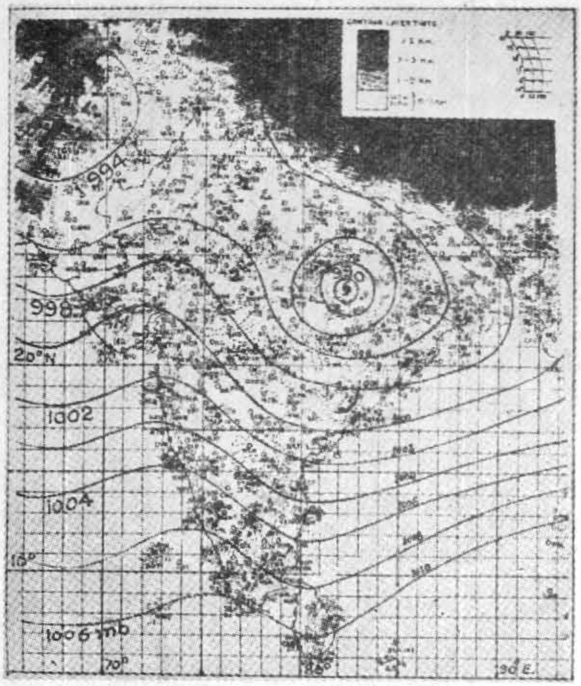

Fig. 1
Sea level chart at 00 GMT on 1 August 1969

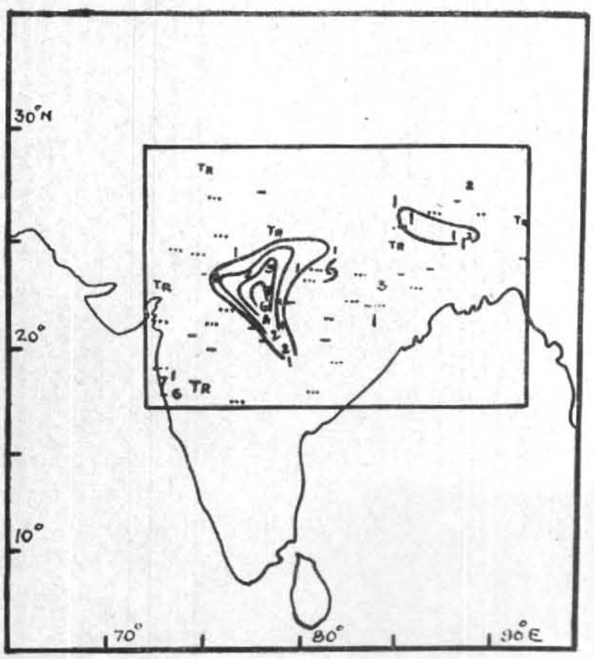

Fig. 2
Rainfall distribution as reported at 1200 GMT on 1 August 1969

(*i*) $R_o << 1$
(*ii*) $R_o^2 \; R_i \sim 1$
(*iii*) $L < a$

where R_o is the Rossby Number, R_i represents the Richardson Number, L is the characteristics length scale of the motion and a is the radius of the earth.

We have chosen a synoptic situation where the above conditions are fulfilled. If the characteristics velocity of the system (U) is taken as 10 m sec^{-1} and L is 1000 km then $R_o = 0 \cdot 16$, on using a value of f appropriate for 25° N.

Similarly, using the tephigram of Nagpur we estimate that R_i is of the order of 30. Hence

$$R_o^2 \; R_i \sim 0 \cdot 9$$

The length scale L is also much smaller than a.

3. Boundary Conditions

We assume that ω vanishes along the horizontal boundaries of the grid chosen for study (Fig. 1). Two more boundary conditions are needed at the top and bottom of the atmosphere. We assume that ω vanishes at the top of the atmosphere.

At the 1000 mb surface, which we assume to represent the base of the atmosphere, we put

$$\omega_o = g l_0 \left[\left(\frac{\partial \phi}{\partial t} \right)_0 - \mathbb{V}_0 \nabla H - L \, \zeta_0 \right] \quad (3 \cdot 1)$$

Here the subscript '0' represents conditions at 1000 mb, so that $\mathbb{V}_0$ and ζ_o represent the geostrophic wind and its relative vorticity at the 1000 mb pressure surface. The contribution from the second term is due to orographic effects. In this model, we have used values of ground elevations from Berkofsky and Bertoni (1955).

The third term in (3·1) represents the contribution to vertical motion due to frictional divergence within the friction layer, L represents a constant which has the dimensions of length, and is taken to be of the order of 200 m. We shall discuss the nature of L at a later stage, but it is to be noted that in regions of cyclonic vorticity (ζ_o positive) friction generates upward velocity at the lower boundary. On the other hand, in regions of anticyclonic vorticity, friction leads to downward motion at the top of the friction layer.

4. Computation procedure and results

Equation (2·3) was solved by three dimensional relaxation at each grid point. The relaxation was carried out on a 11×7 grid for the horizontal coordinates. Ten pressure levels from 1000 to 0 mb, with intervals of 100 mb formed the

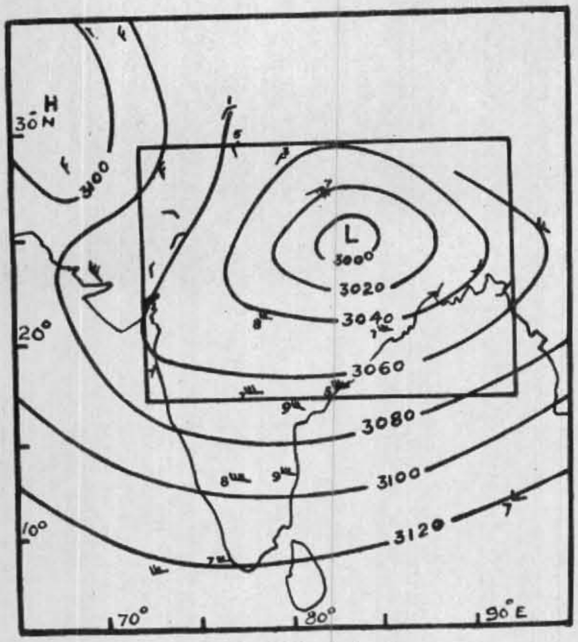

Fig. 3. 700-mb level

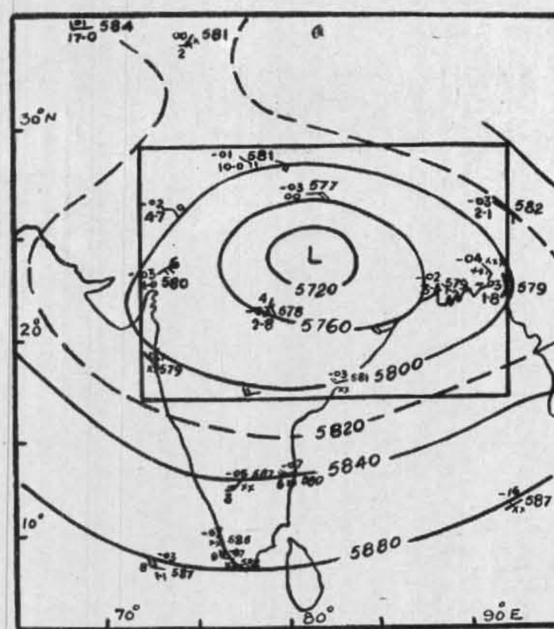

Fig. 4. 500-mb level

Flow patterns at 00 GMT on 1 August 1969

vertical grid. The programme for machine computation evaluated the height tendency (ϕ_t) for the levels 950, 850, 750................150 and 50 mb. From these values of ϕ_t the vertical velocity (ω) was computed for 1000, 900 800, ... 100 and 0 mb with the help of equation (2·2).

A grid increment of 220 km was used in the computations. The size of the grid is shown in Fig. 1. The stability factor σ was estimated for each of the ten layers from available radiosonde soundings in the vicinity of the depression. In this way, σ is a function of pressure only.

We used a factor of 1·7793 for over-relaxation by the extrapolated Liebmann method. We found that with this factor of over-relaxation convergence was achieved in approximately 35 iterations. Each residual was reduced to less than 2^{-14}.

The synoptic situation for which this study was made is shown in Fig. 1. It shows a monsoon depression centred approximately near 23°N, 82°E at 00 GMT on August 1, 1969. The precipitation recorded during the period from 03 to 12 GMT on August 1, 1969 is shown in Fig.2.

In Figs. 3 to 5 we present the upper level chart for 700, 500 and 200 mb. In preparing these charts we took care to adjust the spacing of the contours according to the available winds and the gradient wind equation. Since the radiosonde values are not generally very accurate, we have given more weightage to the winds and small corrections in the contour values have been applied. These corrections are of the order of 20-40 gpm. To save space we have only presented the chart for three different levels. In reality, we prepared charts for all the ten-layers for which ϕ_t was computed. The charts for the nonmandatory layers, such as, 400 mb were prepared by interpolation between the layers immediately above and below. A 50-mb chart was similarly prepared by extrapolation.

Figs 6 to 10 represent the computed values of omega for 1000, 800, 700, 500 and 200 mb respectively. To save space we have not shown the omega values at the other levels. On these charts we have also superimposed the composite neph analysis obtained from the APT unit, Bombay. It may be noticed that the maximum value of ω was obtained at 800 mb, and was of the order of 10×10^{-4} mb/sec (of the order of 1 cm/sec), this is comparable to values obtained by the other workers in tropical areas. Lateef (1967) using the kinematic technique found the value of ω in a depression field over the Caribbean area to be of the order of 2 to 4 mb per hour (10×10^{-4} mb/sec). This is comparable to our values of ω.

It is interesting to note that in the lower troposphere, areas of marked ascent occur to the west and the southwest of the depression centre. Although there exists an area of ascent to the east of the depression, but the vertical velocities

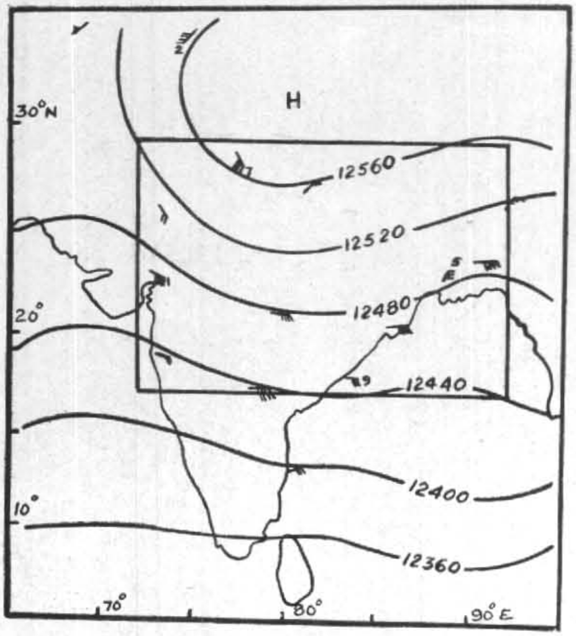

Fig. 5

Flow pattern at 200-mb level at 00 GMT on 1 Aug 1969

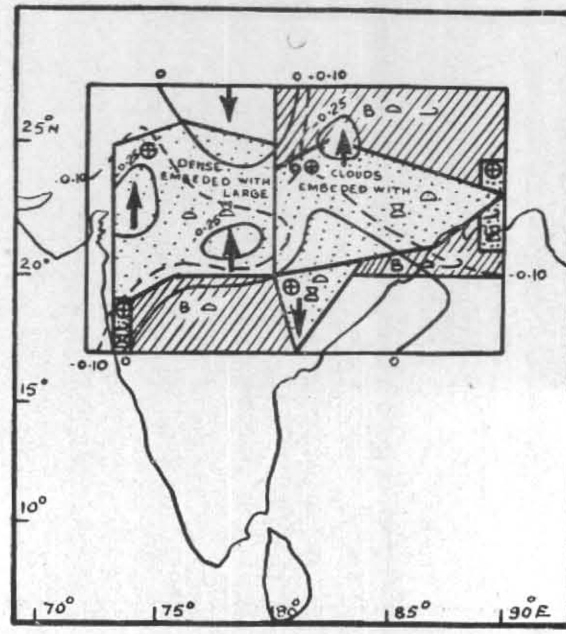

Fig. 6

Neph analysis of 1 Aug 1969 and superimposed computed 1000 mb vertical velocity (10^{-4} mb/sec)

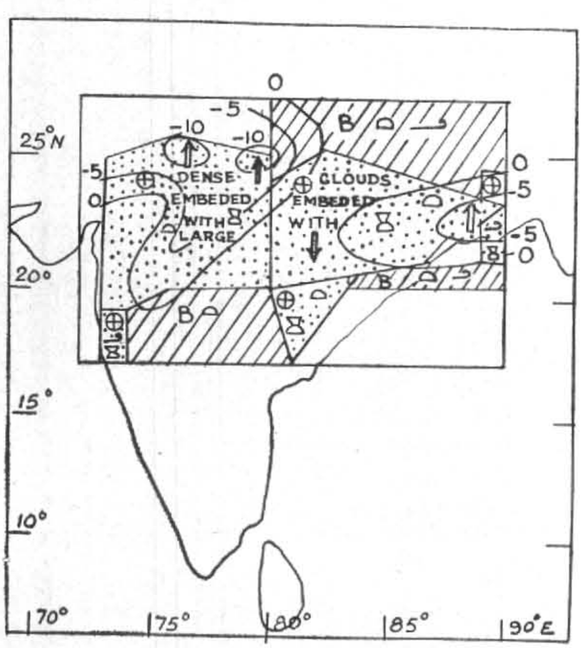

Fig. 7. 800 mb

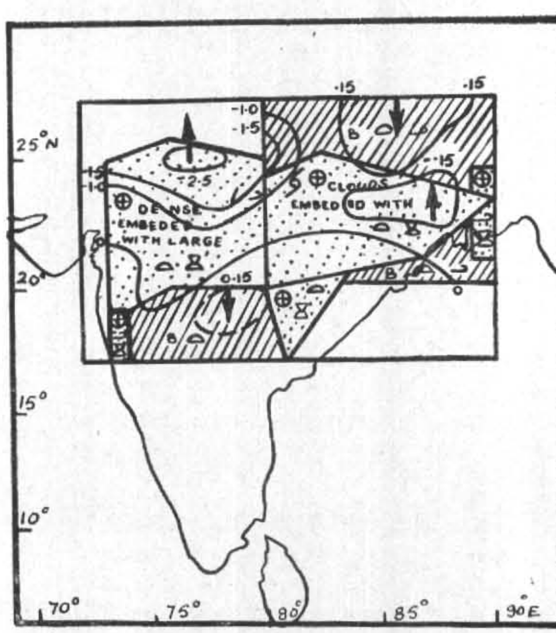

Fig. 8. 700 mb

Neph analysis of 1 Aug 1969 and superimposed computed 800 and 700 mb vertical velocity (10^{-4} mb/sec)

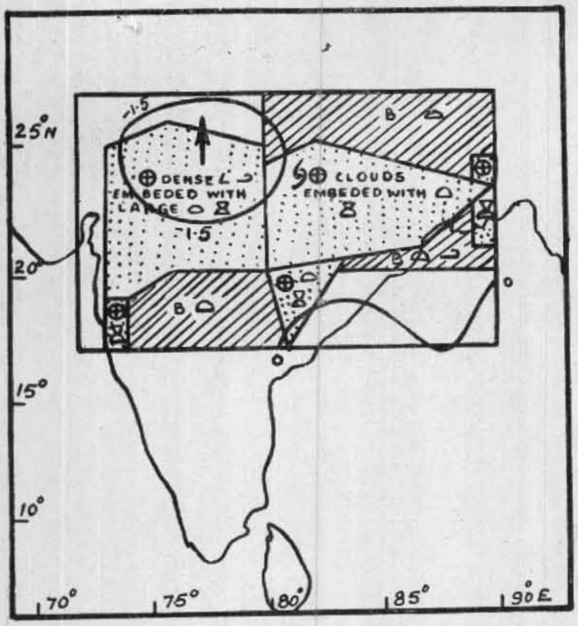

Fig. 9. 500 mb

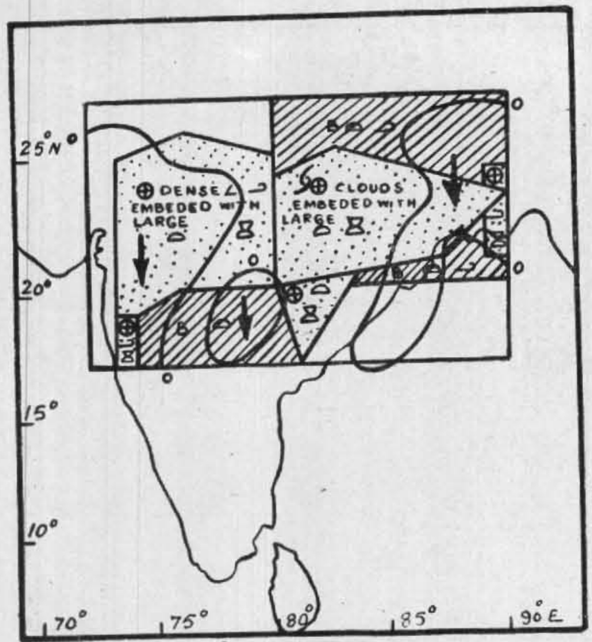

Fig. 10. 200 mb

Neph analysis on 1 Aug 1969 and superimposed computed 500 and 200 mb vertical velocity (10^{-4} mb/sec)

are weak. In the upper troposphere (200 mb) the field shows a reversal, indicating subsidence both to the east and west of the depression. At 500 mb, the computed values of ω were in general weak. It is also of some interest to note that the area of dense clouds, especially *Cb* development is associated with the area of ascent in the lower troposphere. It may be further seen that the centre of the up current at 1000 mb agrees well with the centre of heavy rainfall, however the tilt of the centre of maximum vertical velocity seems to be from southeast to northwest with height.

5. Discussion of results

From the right hand side of equation (2·3) we see that we get ascending motion in regions of upward increase of cyclonic vorticity advection, and near the maxima of warm temperature advection. In the present case study we could not find much synoptic evidence of warm temperature advection. It thus appears that the ascent of air is largely the result of upward increase in cyclonic vorticity advection.

There is one other aspect which requires further investigation. It has been pointed out by Phillips (1963) that geostrophic balance on which the ω equation is based is best suited for synoptic-scale circulations corresponding to hemispherical wave number 4 to 8. The clouds depicted by weather satellites are, however, much smaller. Consequently, we can not expect very great quantitative agreement between the details of the computed vertical motion and the observed cloud pattern. It is encouraging to note that despite this difficulty there is at least qualitative agreement in our case study between the cloud pattern and the ω field. This feature was also noticed by Sanders (1966).

We should also like to point out that the maximum precipitation recorded was 11 cm in 24 hours. This does not quite justify the neglect of latent heat in equation (2·2). We propose to include this effect in a later study.

The factor L depends on the eddy viscosity coefficient K and the angle of cross isobar flow. In the equation for ω_0, we have assumed that the eddy viscosity coefficient K remains constant with height. As is well known, this is not a realistic assumption. We are conducting further work on this aspect of the problem.

Acknowledgements

We are grateful to Dr. P. Koteswaram, Director General of Observatiories for his kind interest in the work and also for providing funds for computer time. We also thank members of Research Unit of NHAC for their valuable assistance.

REFERENCES

Berkofsky, L. and Bertoni, E. A. 1955 *Bull. Amer. met. Soc.*, 36, 350-354.

Das, P. K. 1962 *Tellus*, XIV, 2, pp. 212-220.

Lateef, M. A. 1967 *Mon. Weath. Rev.*, 95, 11, pp. 778-790.

Phillips, N. A. 1963 *Rev. Geophys*, 1, 2, pp. 123-176.

Sanders, F. 1966 Proc. Workshop Satellite Data in Met. Research, Boulder, Colorado, 25-31 Aug. 1965, pp. 103-111.

DISCUSSION

(Presented by R. K. Datta)

DR. P. KOTESWARAM : Has latent heat been taken into account ?

SHRI R.K. DATTA : In the present study we have not included the effect of diabatic heating but propose to do so in subsequent studies. However, as seen from the equation, the effect of diabatic heating is only to increase the rate of ascent in the precipitation areas.

DR. KOTESWARAM : What is the size of the grid ?

SHRI DATTA : It is 220 km.

DR. P.R. PISHAROTY : Was the vertical velocity areas at 1000 mb same as those at 900, 850, 700 mb etc ?

SHRI DATTA : The areas of vertical ascent at 900 mb are more comparable to 1000 mb. Those at 850 mb are comparable to 700 mb and agreed with the precipitation areas.

SHRI J. SHUKLA : The boundary conditions exert great influence on the ω in the interior of the computation domain. What is the justification for taking such a small area for computation and thereby making ω zero only a few grid points away from the centre? I suggest increasing the area.

SHRI DATTA : The main limitation is the memory of the computer ; working with 10 levels, near saturation is reached.

PROF. K R. RAMANATHAN : Will it make any difference if the upper limit of the atmosphere is taken to be, say at 200 mb.

SHRI DATTA : It would ordinarily bring in some discrepancy, since the ω field at 200 mb is unknown and as shown by our study is not a zero. However, it would be possible to take the upper limit at 200 mb or 100 mb by computing ω field at this level from adiabatic considerations as done by Lateef. For this the temperatures should be accurate.