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Moisture processes in the operational quasi-geostrophic model

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ABSTRACT. 5-layer quasi-geostrophic model with input data at 850, 700, 500, 300 and 100 mb is being run on an operational basis once a day. 24 hours forecasts of contour heights at the above levels and vertical velocity (ω) values at 200, 400, 600 and 800 mb are obtained, The model includes orography, cumulus scale heating in the tropical domain of the forecast area. The area of integration is from 40° to 120° E and 5·0° to 42·5°N at 2·5° grid interval. The model boundary values are generated from a bigger coarse mesh area from 0° to 140° and 0° to 60°N at 5° interval.

1 Introduction

Although quite successful forecasts have been made for middle and high latitudes with adiabatic quasi-geostrophic models, it has become increasingly evident, that the inclusion of cumulus scale heating in some form is necessary for the prediction of tropical flow patterns with the model. During the monsoon season, when the atmosphere is conditionally unstable and near saturation, the release of such latent heat as a direct result of the vertical velocity induced by conditional instability plays a vital role for the origin and maintenance of disturbances like the depressions.

The methods to achieve this may be classified under three categories : (i) those which depend on the convergence of moisture in an entire column of atmosphere (Kuo 1965, Estoque 1968, Rosenthal 1970, Krishnamurty 1967, 68, 69, 71, 73); (ii) those which depend on boundary layer convergence of moisture (Ooyama 1964 Charney and Eliassen (1964), Charney (1968), Ogura (1964) and Yamasaki (1968) and (iii) those which depend on lapse-rate adjustment (Manabe, Smagorinsky and Strickler (1965) and Benwell and Bushby (1970).

The difficulties in the application of any of these schemes which have no doubt, a satisfactory theoretical foundation are two-fold in an operational model especially for the Indian region. Firstly the specific humidity or mixing ratio values are not available over the vast oceanic areas in the region even on a sparse network, so that recourse has to be made to climatological values on a day to day basis. Secondly all these

parameterization processes usually contain a small number of empirical constants for the description of the cloud properties at their origins, their successive generation and the form of vertical distribution of heating. The motivation for this study is to test a few of the schemes at least in a crude way and compare the predictions with observations, to bring the computations into line with reality.

Accordingly, in Section 2 a brief description of the current version of the operational quasi-geostrophic model is given. Section 3 describes the parameterization processes tested. In the next section the results of an actual case study of a recent monsoon-depression is presented.

2. Model description

Details of the 4-layer operational quasi-geostrophic model have been discussed by Mukerji and Datta (1973) which we refer hereafter as version I. With the experience gained from the working of the model, the current version of the model (Ramanathan and Bansal 1975) was evolved, which differs from the previous one in the following aspects :

(a) Boundary conditions

(b) Orographic effects and

(c) Moisture processes (*see* Section 3)

A brief description of differences (a) and (b) are now given. The governing equations are

The vorticity equation

$$\frac{g}{f} \nabla^2 \frac{\partial z}{\partial t} + \frac{g}{f} J(z,\eta) = \bar{f} \frac{\partial \omega}{\partial p} \quad (2.1)$$

The thermodynamic equation

$$\frac{\partial}{\partial t}\left(\frac{\partial z}{\partial p}\right) + \mathbf{V}.\nabla \frac{\partial z}{\partial p} + \sigma_s\,\omega = -Q \tag{2.2}$$

And the diagnostic ω equation

$$\sigma_s\,\nabla^2\,\omega + \frac{\overline{f^2}}{g}\,\frac{\partial^2\,\omega}{\partial p^2} = \frac{\overline{f}}{g}\;\frac{\partial}{\partial p}\,(\mathbf{V}.\nabla\eta)$$

$$- \nabla^2\left(\mathbf{V}\,.\,\nabla\frac{\partial z}{\partial p}\right) - \nabla^2 Q \tag{2.3}$$

2·1. *Boundary condition*

In version I, the boundary values were kept constant so that ω and the height tendencies were zero during the integration. Though stable forecasts were achieved in this formulation, the forecasts were not realistic in a boundary zone of increasing width with integration. This posed a serious problem especially in the northern boundary ($42\cdot5°$N) when amplifying systems originally north of the domain extended south into the domain especially in the Chinese region east of the Himalayas, during 24 hours in reality, but effectively cut off by a constant boundary, in our model. The current version uses constant boundaries on an extended coarse grid model at 5° interval on a domain from 0° to 140°E and 0° to 60°N to compute boundary values for each step for ω and $\partial z/\partial t$ for the boundaries of a fine mesh model at $2\cdot5°$ interval from 5° to $42\cdot5°$N and 40° to 120°E. These are stored on a tape and used for the fine-mesh integration which is run subsequent to the coarsemesh integration.

2.2. *Orographic effects*

A number of grid-points in the lower levels (*e. g.*, 850 and 700 mb) are fictitious in the X, Y, P frame since the terrain pressure values at these grid points will be lower than the surface o analysis. The analysis is however completed for continuity and contour values at these fictitous points are included *prima facei* in the input. However the lower boundary condition for the omega equation is applied not at 1000 mb, as is done in version I, but at the pressure level nearest to the terrain pressure value at the grid point. Starting from a surface grid point the relaxation for the solution of the omega equation in the vertical *p*-surface proceeds to the grid-point above the boundary surface only. This procedure of relaxation along the vertical for each grid-point excludes the fictitious points. The height tendency is also not computed at these points.

3 Moisture Processes

3.1. *General considerations*

In the quasi-geostrophic theory the static stability parameter (σ) is a function of p only and cannot be varied from grid-point to grid-point in the pressure surface.

If the forecast domain includes both the middle latitudes and the tropics and a general overlay of these regions, choosing a single value of σ in an area of widely varying static stability from stable ($\sigma_E>0$) to conditionally unstable ($\sigma_E<0$), introduces certain empiricism. The σ_E values are negative or marginally positive for the tropical portion of the forecast domain. A single value of σ_E averaged overall the grid-points is about 2 orders less than σ even if it is positive. For a negative value of σ_E, the relaxation does not converge and for a very low positive value of σ_E the vertical velocities are very unrealistic. Hence the w equation presently is solved with a dry σ, with an additional forcing due to diabatic heating as the grid-points.

The forcing function of the heating term may be written

$$-\nabla^2 Q = -\frac{R}{C_p\,p}\,\nabla^2 H \tag{3.1}$$

where H is the rate of heating per unit mass of air per sec. expressed through the first law of thermodynamics.

In our model we have restricted H to include only the effect of release of latent heat H_L in the free atmosphere, omitting the effects due to sensible heat transfer from water surface and radiation effects. Our assumption that measures of the instantaneous tendencies of atmospheric variables for synoptic scale motions in short-range forecasting can be reasonably made without including these effects may still prove wrong.

Heating functions describing the effects of latent heat release can be easily prescribed if dynamic effects of absolutely stable air is producing condensation. This is generally true of the large sacle precipitation from stable non-convective cloud systems.

Danard (1966) has prescribed such a type of latent heat inclusion with considerable success. Version I also included effects of one such formulation. Krishnamurti did not include latent heat release from this process since it is probably less important in tropics than the latent heat release under a conditionally unstable vertical stratification. Riehl and Malkus (1958) mentioned that the cumulus scale "hot towers" are most effective in the latent heat generation.
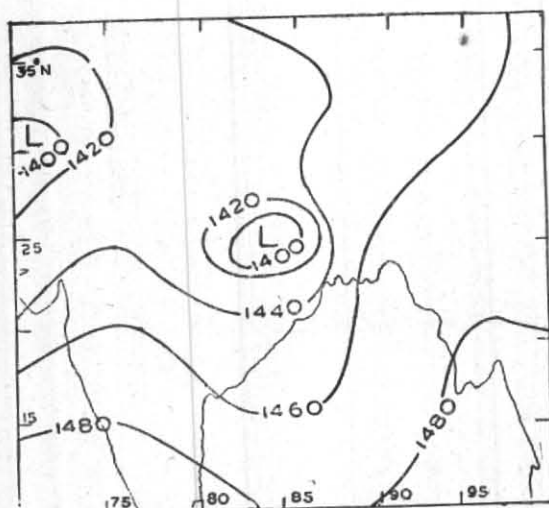
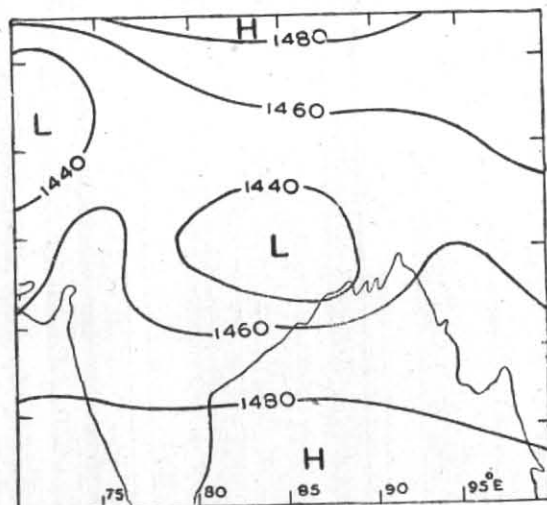Fig. 1. 850 mb (realised) valid for 00 GMT of
20 Aug 1975

Fig. 2. 850 mb forecast 24·hr valid for 00 GMT of 20
Aug. 1975 (without moisture)

In our model, we have adopted the method of Charney and Elliassen (1964) which is the simplest of all, resembling the stable region formulation and does not involve the actual calculations of the cloud area as required in more sophisticated formulations. This method was used by Krishnamurti (1967) with success in his diagnostic balance-models.

3·2. Method A

The convergence of moisture flux may be written

$$I = \frac{1}{g} \int_{900}^{0} \nabla . q \, \mathbf{V} \, d \, p - \frac{\omega_{900} \, q_{900}}{g} \qquad (3.2)$$

The first term on the right hand side is the horizotal flux above the boundary layer and the second is the moisture leaving or entering the top of the boundary layer, which is taken at 900 mb in the model.

The heating function

$$-\nabla^2 Q = - \frac{RLg}{C_p p} \nabla^2 \frac{1}{q_s} \frac{\partial q_s}{\partial p} a \, I \qquad (3.3)$$

where $a = ·01$ (percentage of the cloud amount) controls the (3·3) latent heat release.

Before the heating term is activated in the omega equation, it must pass the following requirements at each grid-point :

(a) The moisture flux should be positive

(b) w at the level should be rising

(c) regions of relatively dry air with relative humdity less than 60% are eliminated.

3.3. Method B

In a second experiment, following Charney (1964, 1968) and Yamasaki (1968), only the frictional boundary layer moisture convergence was considered

$$I = \frac{\omega_{900} \, q_{900}}{g}$$

The vertical distribution of heating was varied with pressure by a factor $(T_s\text{-}T) / \int (T_s\text{-}T) \, dp$ which is not calculated each time in the operational model but derived by empirical constants (0·7, 0·2, 0·1 at 800, 600 and 400 mb) calculated separately.

4. Case study

During the monsoon season in the lower level (e.g., 850 mb) there is invariably a high east of 100°E in the easterlies. During integration with version I this high moves to the adjoining Bay of Bengal and Indian subcontinent, completely filling up any low pressure systems and completely obliterating their existance. If the systems (e.g., monsoon depressions) were quite intense as represented by a strong gradient, they could retain their indentity during integration but were partially filled up and were stationary. This was the difficulty when version I was used and also when the current version was used without the present moisture formulations.

The case study presenred appropriately is that of one of the many monsoon depressions studied. Fig. 3 shows the location of the depression at 22·5°N, 86°E at 850 mb on 19 August 1975 at 00 GMT.
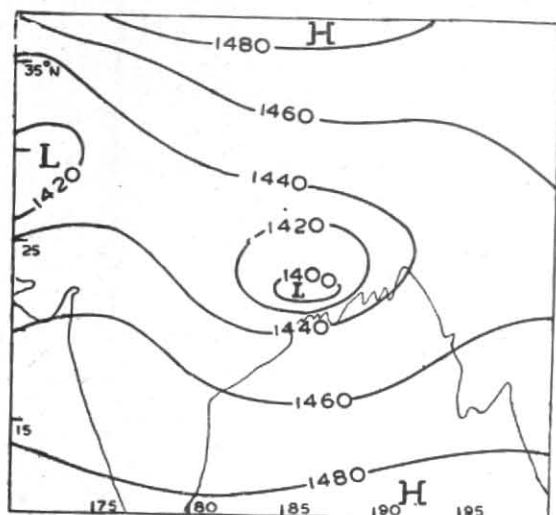
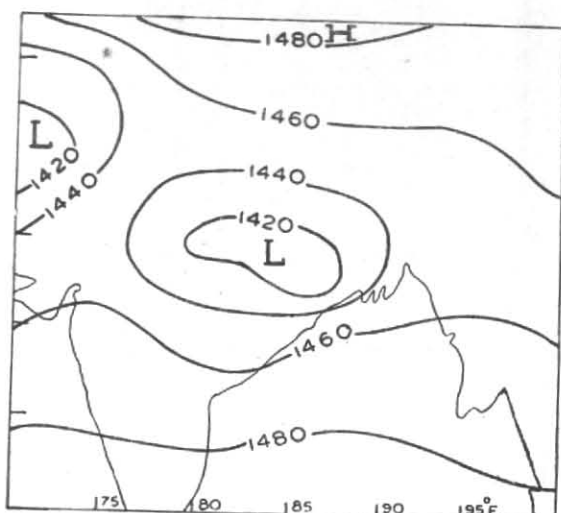Fig. 3. 850 mb initials for 00 GMT of 19 Aug 1975

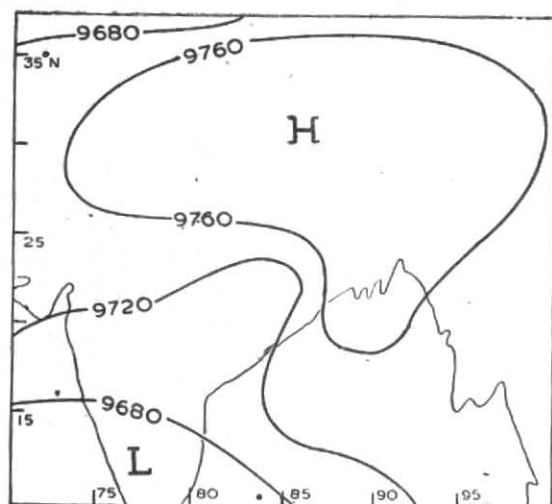Fig. 4. 850 mb forecast 24 hr. valid for 00 GMT of 20 Aug 1975 (Kuo's parameterisation)

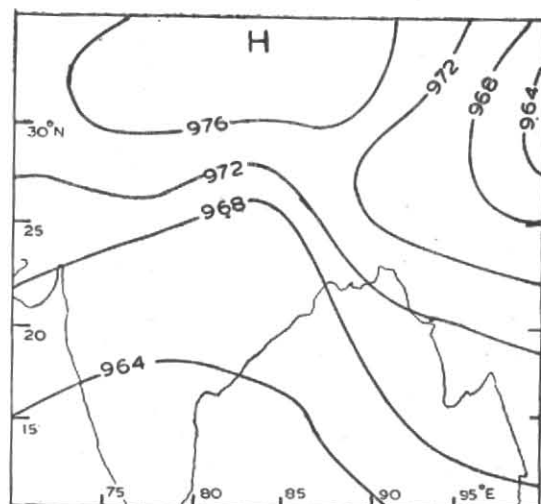Fig. 5. 300 mb forecast 24 hr valid for 00 GMT of 20 Aug 1975 (Kuo's parameterisation)

Fig. 6. 300 mb (re   ' valid for 00 GMT of 20 Aug 1975

Fig. 2 shows the forecast position after 24 hours by the model when moisture processes were not included. The depression though somewhat filled and became diffused retained its identity but showed little movement (see Fig. 1 for the actual position) at 24·5°N, 84·0°E.

Moisture processes were then included as in method A in 3·2. The depression showed movement in the right direction and is positioned at 25·0°N and 82·5°E slightly faster than in the verifying chart, Fig. 4 though it filled up slightly compared to the actual case.

The 300 mb flow pattern agreed well with verifying chart (Figs. 5 and 6). The moisture processes were formulated as in method B and the model integrated as before.

The forecast depression position is at 25·0°N and 82.5°E (Fig. 7) the movement agreeing well with the verifying chart. Also the depression shows no filling up. But in the 300 mb forecast chart (Fig. 8) the high over Himalayas had intensified somewhat unrealistically compared to the verifying chart (Fig. 6).
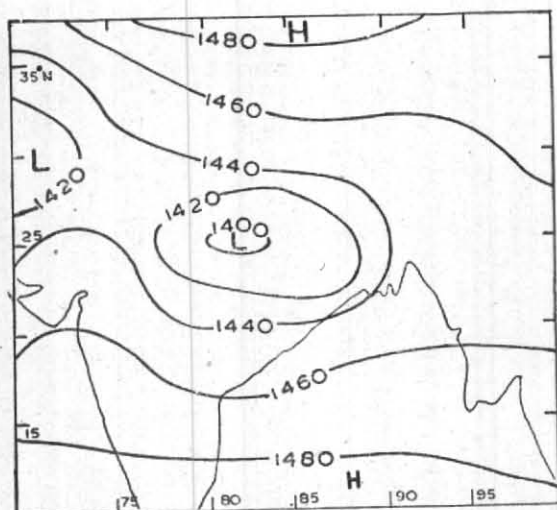
Fig. 7. 850 mb forecast 24 hr valid for 00 GMT of 20 Aug. 1975 (with boundary layer convergence)
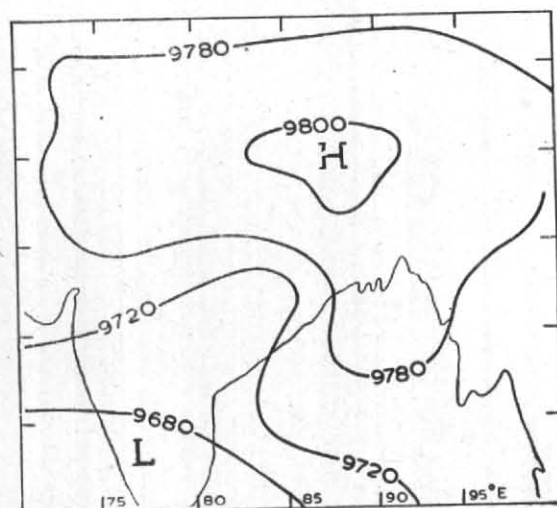
Fig. 8. 300 mb forecast 24 hr. valid or 00 GMT of 20 Aug. 1975 (with boundary layer convergence)

In the many cases we have studied, this tendency to develop anticyclogenesis at higher levels was found to increase, with the deepening of the system at lower levels.

5. Conclusions

Inclusion of cumulus scale heating in the prediction models is important for the monsoon region for realistic forecasts and the results of the study are encouraging. However observed values of mixing ratio in a close network for a few cases of disturbances like monsoon depressions are perhaps necessary, to determine the empirical constants on a more satisfactory basis. It may then be possible to find out the most suitable methods of cumulus parameterization for the

current model and for future primitive equation models. One of the objectives of the MONEX 1977 experiment is precisely this.

Acknowledgements

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APPENDIX

List of symbols units and typical magnitutes

Variables	Meaning	Unit	Typical magnitudes
X	Distance along zonal direction	cm	10^5
p	Vertical distance	mb	10^2
$\mathbf{V}$	Horizontal. velocity vector	cm/sec	10^3
ω	Vertical velocity	cm/sec	1
f	Coriolis parameter	sec^{-1}	10^{-4}
$\overline{f}$	Mean f	sec^{-1}	10^{-4}
t	Time	sec	3600
L	Latent heat of evaporation	Cal g^{-1}	600
σ	Dry static stability	$Dynes^{-2}cm^3$/cm MB^{-2}	$\begin{cases} 10^{-8} \\ 10^{-3} \end{cases}$
σ_E	Moist static stability	$Dynes^{-2}cm^3$	10^{-8} to 10^{-9}
T	Air temperature	Degree Kelvin	3×10^2
R	Gas constant	$\begin{cases} erg\ g^{-1}{}^\circ k^{-1} \\ cal\ g^{-1}{}^\circ k^{-1} \end{cases}$	$2 \cdot 87 \times 10^6$
c_p	Specific heat at constant pressure	cal $g^{-1}{}^\circ k^{-1}$	$0 \cdot 240$
g	Acceleration due to gravity	cm/sec^{-2}	$980 \cdot 6$
∇^2	Laplacian operator	cm^{-2}	10^{-14}
J	Jacobian operator	cm^{-2}	10^{-14}
d	Grid distance	cm	10^7
q	Specific humidity	gm/gm	10^{-3}
q_s	Saturated q	gm/gm	10^{-3}
$\dot{e}_s$	Saturated vapour pressure	mb	20
T_s	Temperature of a parcel raised along moist adiabat	$^\circ$degree	3×10^2