Indian J. Met. Hydrol. Geophys. (1978), **29**, 1 & 2, 205-214

551·513·1 : 551·557 (213)

A numerical study of time-mean northern summer monsoon with steady and fluctuating heating

C. P. CHANG and R. J. PENTIMONTI

Naval Postgraduate School, Monterey, California, U.S.A.

ABSTRACT. In a study of the planetary-scale vorticity budget of the northern summer monsoon, Holton and Colton (1973) found that a strong vorticity sink is needed to balance the generation by horizontal divergence at 200 mb. In this work we use numerical experiments to examine the hypothesis that the fluctuation of the divergence forcing as observed by several studies serves as this sink. A 3-level numerical model is used with the thermal forcing specified by the horizontal divergence distribution observed by Krishnamurti (1971). Two experiments are carried out, one with heating steady in time and the other fluctuating in time with a period of 10 days. The time-mean fields of the quasi-steady state solution in both experiments show a significant westward phase shift of the Tibetan high, thus the fluctuation of heating cannot be a sufficient sink. On the other hand, the positions of the simulated tropical upper tropospheric troughs agree quite well with those observed. This suggests that damping due to cumulus transport is more likely to account for the vorticity sink. Transient wave activities are also found at the upper level in both experiments. They appear to be drawing energy from the planetary-scale flow and could serve as a vorticity sink. These waves are most active in the fluctuating heating case, suggesting that local barotropic instability at upper levels may be more important than that implied by the time-mean wind.

1. Introduction

In the study of the northern summer planetary-scale monsoon circulation, Holton and Colton (1973) discovered that very strong damping is necessary to balance the budget of time-mean planetary-scale vorticity at the upper levels. Specifically they found that without this strong damping to counterbalance the generation of vorticity by horizontal divergence the amplitude of the long waves of the monsoon circulation would be too large at the 200 mb level and the phase would be shifted westward by up to 5000 km. This same phase problem was also apparent in a numerical study by Abbott (1973), who used a 3-level quasi-geostrophic model with prescribed heating to study the monsoon.

In looking for a source of this strong damping mechanism, Holton and Colton suggested two hypotheses. The first is the vertical mixing of vorticity and momentum by deep cumulus convection. The second is related to the observed time fluctuation of the monsoon system. If the large scale divergence contains a strong fluctuation then a correlation term of the divergence and vorticity fluctuations will appear in the time-mean vorticity equation and may reduce the generation of the time-mean vorticity by the time-mean divergence.

Murakami (1974) used a three-component, quasi-spectral model with prescribed heating which has a fluctuating component. He found that it can reasonably simulate the amplitude and the phase of the long waves associated with the northern summer monsoon circulation. But his model suffers from the lack of the zonal basic flow which means that advection is absent. Also, his resultant vertical profile of the divergence field differs considerably from observations. Therefore, it is not clear which mechanism causes the seemingly correct phase and amplitude of his model.

The purpose of this study is to examine the second hypothesis of Holton and Colton using a nonlinear, finite difference primitive equation model. We will carry out two controlled experiments. One will have the heating specified to be independent of time and the other will have a fluctuating heating component similar to that observed by Murakami (1972) and Krishnamurti and Bhalme (1976). We hope that by comparing these experiments it will be possible to evaluate the effects of this fluctuation in the monsoon heating.

A second purpose of this study is to simulate the barotropic instability of the tropical easterly jet associated with the northern summer monsoon. This instability arises as a result of the strong easterly jet in the upper troposphere and has been studied by Colton (1973). It represents an important scale-interaction mechanism for the monsoon circulation and may also serve as an energy for the planetary-scale waves.

2. The numerical model

The numerical model used was formulated by Monaco and Williams (1975) and is similar to the UCLA General Circulation Model described by Arakawa and Mintz (1974) who used the primitive equations in spherical coordinates with "sigma" as the vertical coordinate. In this study the domain of calculation is restricted to the band between 18°S and 46°N for the sake of computing economy, and a free slip condition is applied at the lateral boundaries. The horizontal grid spacing is 4 degrees in both latitude and longitude. The troposphere is divided into seven equally spaced sigma layers in the vertical with three main reporting levels. The surface pressure is set at 1013 mb initially. No topography is included although its thermal effect is implicitly contained in the specified forcing function.

In the two controlled experiments carried out the heating function, H, contains two components:

$$H = H_S + H_{AS}$$

where H_S, the zonally symmetric heating, is given by a continuous adjustment to a specified equilibrium temperature, T^*, using the following formula:

$$H_S = -\left[\frac{T - T^*(y, \sigma)}{\tau}\right]$$

Here T is the temperature and τ is a specified adjustment time set to two days in this study. The equilibrium temperature, T^*, is prescribed to maintain the zonal mean north-south temperature gradient so that the mean zonal flow can be properly included.

The horizontal distribution of diabatic heating is derived from the 200 mb time-mean divergence field observed by Krishnamurti and Rodgers (1970) as follows :

$$Q\ (°\mathrm{K\ day^{-1}}) = \mathrm{Divergence\ (sec^{-1})} \times 10^6 \qquad (1)$$

The field is smoothed on the boundaries by reducing the strength of the heating by 50 per cent at the points just inside the boundaries and 80 per cent at the boundary. The entire field was also smoothed by a five point smoother in order to remove small scale features. In the vertical, the full value of heating derived by (1) is specified at the middle level (Fig. 1) and a vertical distribution is assumed such that the heating at the top and bottom levels are 50 per cent and 25 per cent, respectively, of the value at the middle level.

The zonally asymmetric component, H_{AS}, takes the following form :

$$H_{AS} = Q \times \left((1 + A\,(\sin \frac{2\pi}{T}\ t)\right),$$

where A is the amplitude of the heating fluctuation whose period is T. In the first experiment $A=0$; in the second, $A=0{\cdot}5$. A 10 to 15-day period was observed by Murakami (1972) and Krishnamurti and Bhalme (1976) in the northern summer monsoon. Here 10 days is chosen for the periodicity T for the sake of computation economy.

Linear damping terms are included in the momentum equations to represent the effect of scale-independent friction. The drag coefficients are $0{\cdot}3\times10^{-6}$ sec^{-1}, $1{\cdot}7\times10^{-6}$ sec^{-1}, and 3×10^{-6} sec^{-1} at the top, middle and bottom levels, respectively. Horizontal diffusion is also introduced to include the effect of scale-dependent friction for it damps the small scale noise generated by the finite difference scheme of the model. The total damping time at the upper level for the planetary scale is $\geqslant$20 days, much weaker than the value used by Holton and Colton (1973).

The time integration is comprised of continuous sections of one Matsuno step and four leapfrog steps, with a time increment of six minutes. At the onset the pressure levels are initialized at standard heights and temperatures are also specified by the zonally averaged climatology. The model is started without the diabatic forcing function until a quasi-steady zonally symmetric circulation is established. At this point the diabatic heating is introduced and the integration continues until it reaches another quasi-steady state. For both experiments the model runs for 25 more days from the respective quasi-steady states with the data of the last 10 days used for computing the time-mean fields. All data presented is interpolated to constant pressure surfaces of 250 mb 500 mb and 850 mb.

3. Results

The principal circulation features of the northern summer monsoon include the Tibetan high, the Pacific tropical upper tropospheric trough (TUTT), the tropical easterly jet, the Mexican high and the Atlantic TUTT. The simulation of these features by the model are compared with observations, mainly those by Krishnamurti and Rodgers (1970).

3.1. *Time-mean field*

(a) *Steady heating* — The 10 day time-mean fields of wind, geopotential height and temperature at 250 mb and vertical velocity at 400 mb are shown in Figs. 2-5. In Fig. 2 the 250 mb wind field shows an anticyclonic wind centre near 30°N, 52°E which resembles the observed Tibetan high although the position of the generated feature is shifted 30° to the west. The calculated Pacific TUTT extends from 120°W, 40°N to 180°W, 20°N, approximately the same position at the observed TUTT. The TUTT has a northeast-southwest tilt and the amplitude is slightly greater than that observed. The strongest easterly jet is developed by the model in the southern periphery of the Tibetan anticyclone. The jet maximum is located at approximately 60°-90°E, 10°N, with a speed of $22 ms^{-1}$ which is an underestimate of the observed maximum speed of $\sim 30\ ms^{-1}$.

The Mexican anticyclone is centred near 130°W, 25°N. This feature is again shifted to the west of the observed position by 25°. The amplitude of the disturbance compares well with observed.

The Atlantic TUTT, situated in the mid-Atlantic in the area of 15°N-40°N, 20°-60°W is well developed but again the amplitude is slightly larger than that observed. As for the Pacific TUTT, there is no apparent phase shift problem when compared to observations.

The 250 mb geopotential field (Fig. 3) indicates that the Tibetan and Mexican highs and the Pacific and Atlantic TUTT's are all well defined, but the TUTT's are somewhat weaker in comparison to observations. The 250 mb temperature fields (Fig. 4) shows that both highs are warm core systems and both TUTT's are in the cold regions. These temperature anomalies are present throughout the vertical in both the 500 mb field and the 850 mb field.

As expected, the 400 mb vertical velocity (Fig. 5) closely resembles the horizontal diabatic heating distribution. This is consistent with the warm core energetics of the monsoon in which the long wave kinetic energy is converted from the available potential energy generated by diabatic heating. The resultant mean divergence at 250 mb (not shown) is very similar to the horizontal distribution of the vertical velocity. It is also a reasonable reproduction of the 200 mb divergence field as observed by Krishnamurti, which indicates that the specified heating function was properly chosen for simulating the correct forcing of the vorticity field.

(b) *Fluctuating heating* — The mean wind field at 250 mb of the fluctuating heating forcing function is shown in Fig. 6. A close comparison with the mean wind field for the steady heating case could not reveal any significant differences between the two cases except that the amplitude of the long wave appears to be slightly smaller. The phase shifts of the Tibetan and Mexican highs for the steady heating are almost exactly duplicated by the fluctuating heating. Other fields of the fluctuating heating case also resemble closely those of the steady heating case.

The above result suggests that the effect of fluctuating divergence forcing in the summer monsoon hypothesized by Holton and Colton is not prominent enough to explain the phase problem in the diagnostic simulation of the time-mean field. This conclusion is true if the amplitude of the fluctuating component in our model is representative of the actual observed amplitude. Since limited observational knowledge (based on Krishnamurti 1971) does indicate that the actual amplitude of fluctuation may not be much larger than what is specified here, Holton and Colton's hypothesis seems unlikely to be valid.

3.2. *Transient field*

(a) *Steady heating* — Fig. 7 is a time-longitude section for the 12-hr interval 250-mb vorticity field at 12°N from day 15 to day 25. In the western hemisphere from 0°-29°W there is evidence for strong synoptic-scale westward propagating wave activity. The wavelengths of these disturbances vary from 3500 km at 30°W to 2200 km at 60°W.

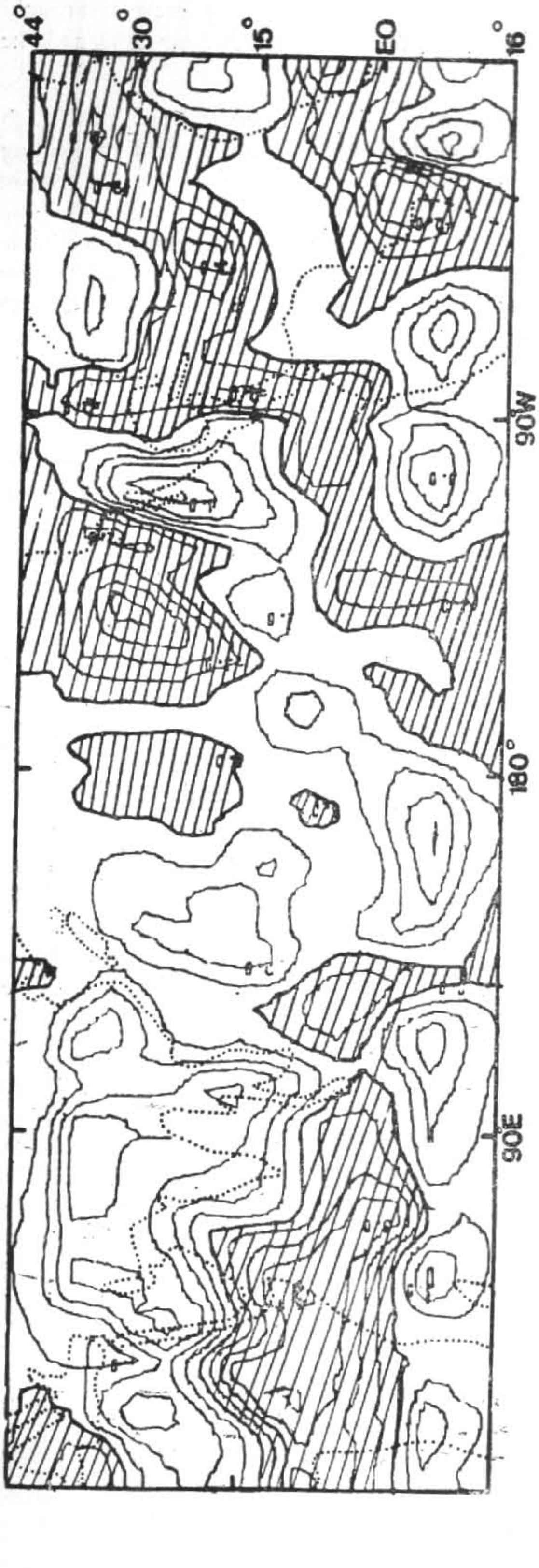

Fig. 1. Horizontal distribution of heating. Hatched area indicates negative heating. The interval is 1°K day^{-1}

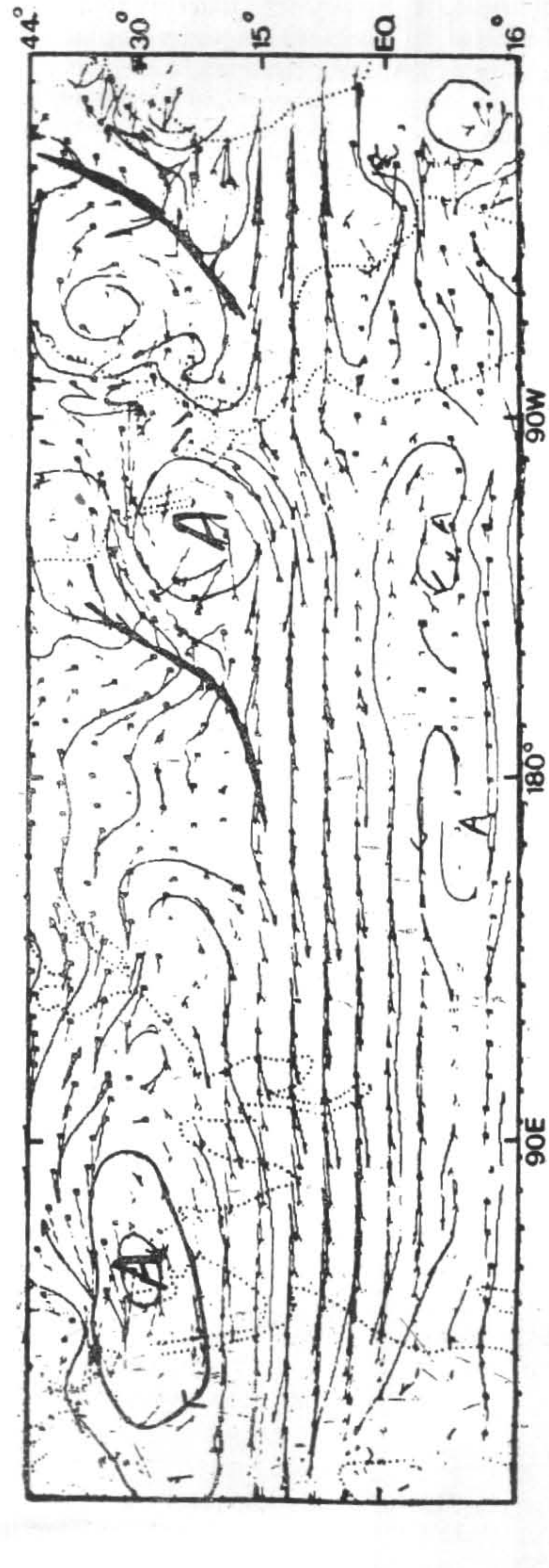

Fig. 2. 250 mb time-mean wind, steady heating

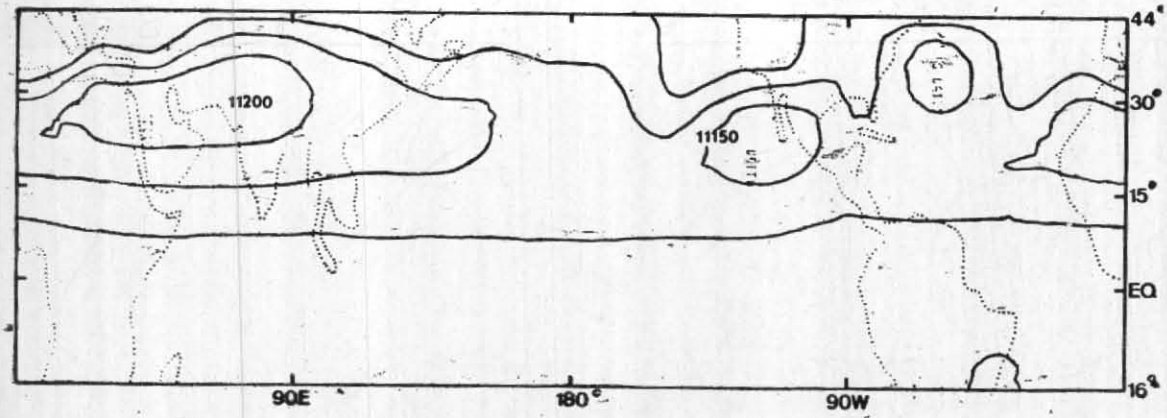

Fig. 3. 250 mb time-mean geopotential field, steady heating. Interval is 50 m^2s^{-2}

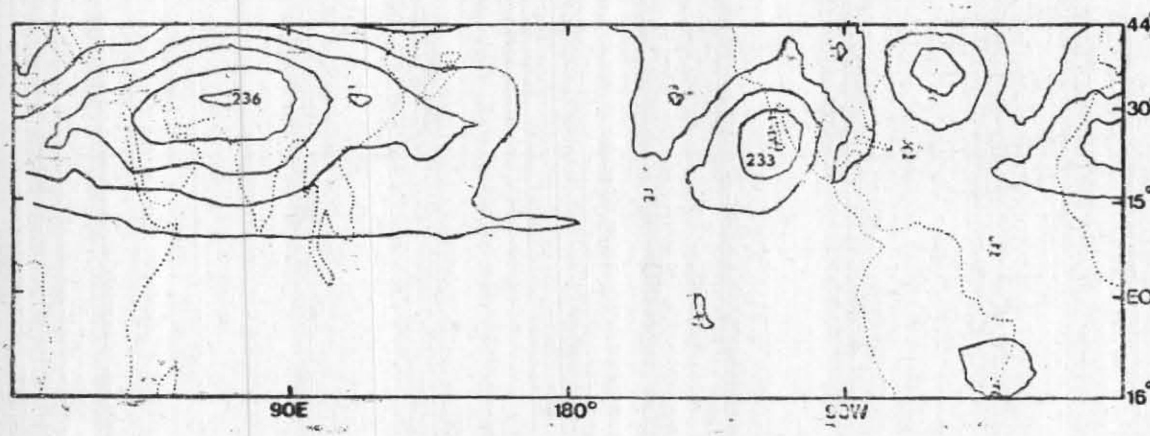

Fig. 4. 250 mb time-mean temperature field, steady heating. Interval is 1°K

In some cases the wavelength can be as short as 1000 km at 180°W to 90°E where the disturbances appear to be much more irregular with long wavelengths at the beginning of the time series and sporadic, short wavelengths toward the end.

Upstream from the jet maximum near 50°E the disturbance activity is very minimal with small amplitude. The waves begin to grow in the area of the easterly jet maximum and continue to grow as they propagate westward. Downstream of the easterly jet maximum the synoptic waves appear to be much more organized compared to upstream of the jet.

Krishnamurti (1971) and Colton (1973) have found that the upper-level synoptic waves are largely due to the barotropic instability associated with the easterly jet. To examine the possibility that this is the generation mechanism of the waves observed in this model, we plot the meridional gradient of the basic absolute vorticity,

$$\beta - (\partial^2 \bar{u}/\partial y^2,)$$

where β is the gradient of the coriolis parameter and $\bar{u}$ is the time-mean 250-mb zonal wind component, as a function of x and y (not shown). The result shows that the sign of this quantity changes within the north-south channel throughout the domain. Thus it is quite probable that the barotropic instability exists everywhere at the upper level. The important role played by the easterly jet in this regard is also reflected by the negative values in the entire latitude band of 12°N.

The result that the wave amplitude is minimum upstream of the easterly jet and increases to a maximum downstream further suggests that the mean kinetic energy is converted into eddy kinetic energy in the vicinity of the easterly jet maximum. There should be a stable area upstream of the easterly jet maximum where the amplitude

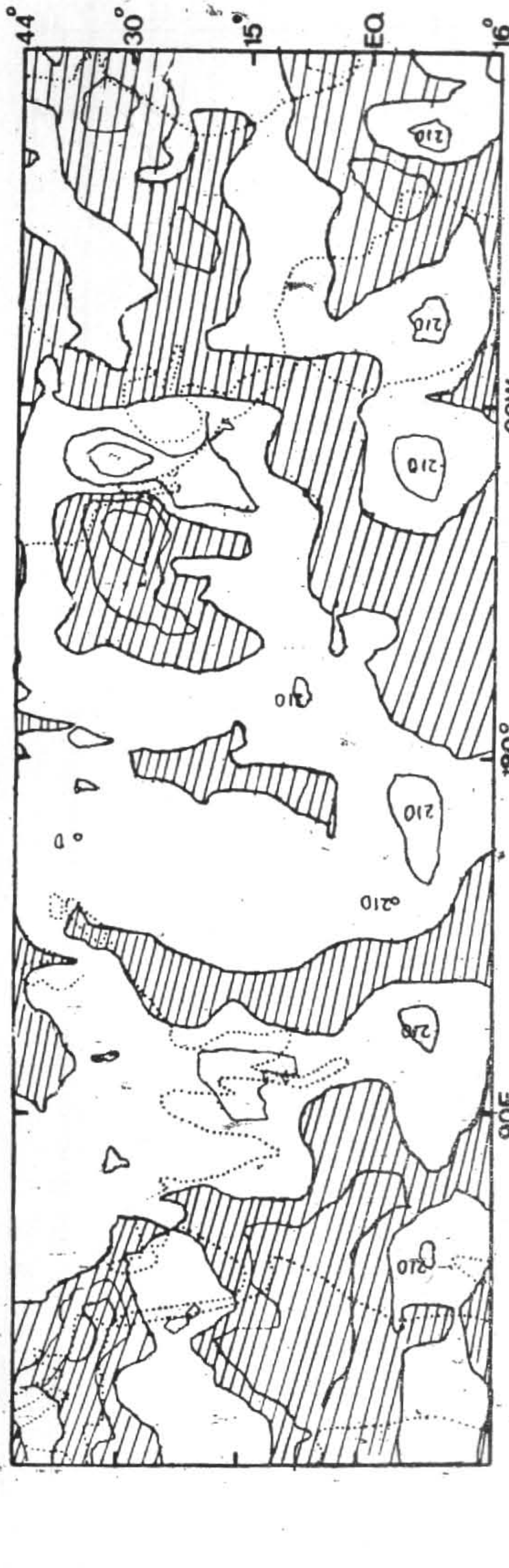

Fig. 5. 400 mb time-mean vertical velocity. Hatched area indicates positive value. Interval is 1 mb hr^{-1}

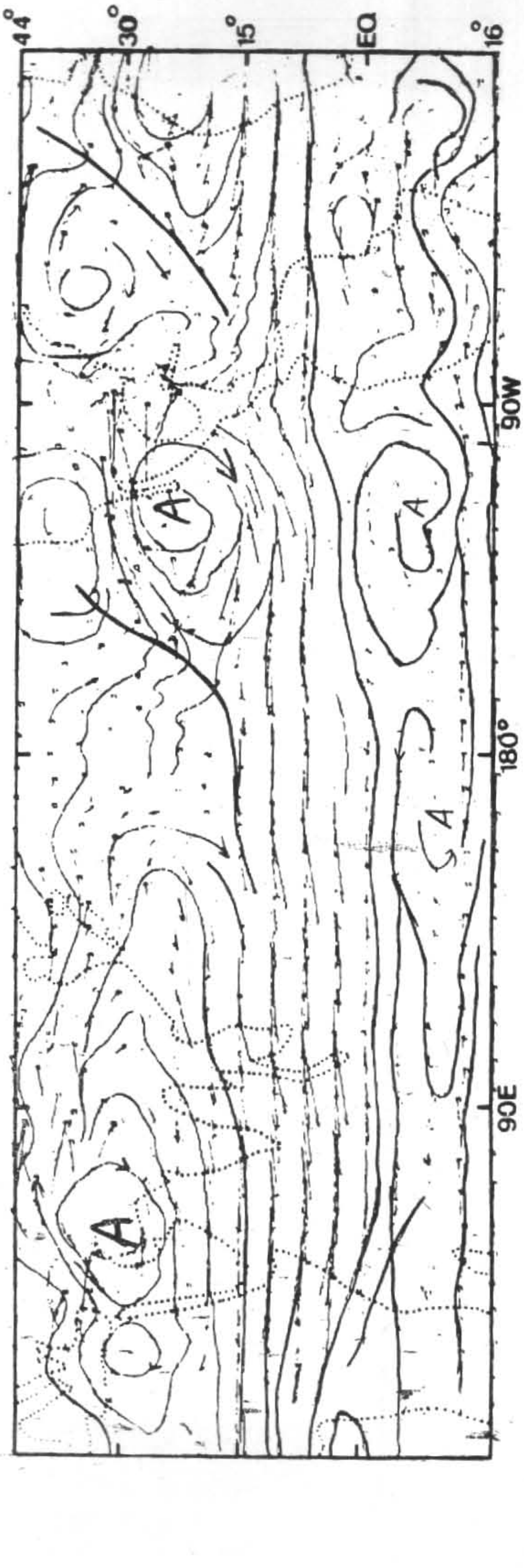

Fig. 6. 250 mb time-mean wind field, fluctuating heating

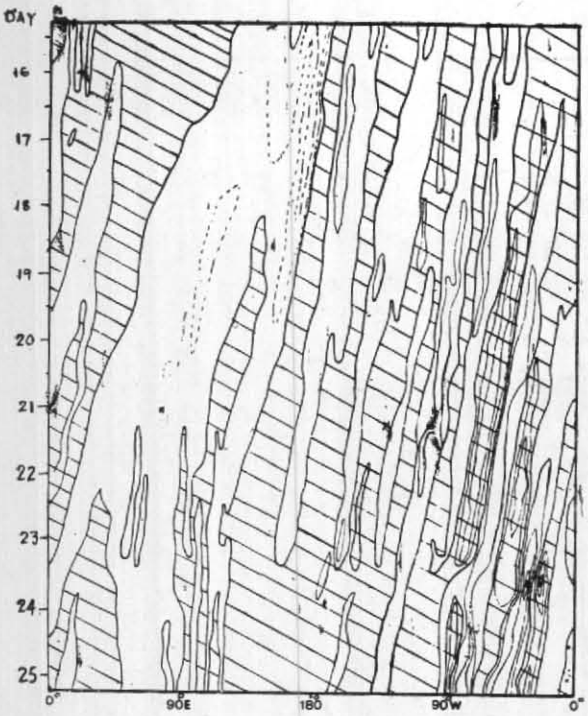

Fig. 7. 250 mb vorticity time-longitude section at 12°N, steady heating. Hatched area indicates positive values

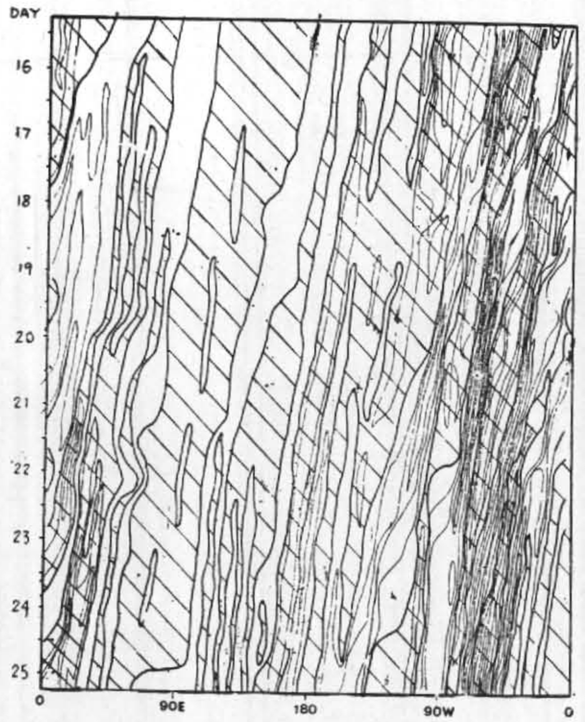

Fig. 8. 250 mb vorticity time-longitude section at 12°N, fluctuating heating. Hatched area indicates positive values.

of the disturbances is small. Near the jet maximum the change in amplitude is large as the wave propagates past the jet, indicating that the jet maximum is the most unstable region (Tupaz *et al.* 1978). The result that the most active area for synoptic-scale wave disturbances is in the Atlantic, Caribbean and central America rather than in the Pacific and Indian Ocean areas is probably related to the model position of the Tibetan high. As this high centre is shifted by 30° to the west from the observed position, a corresponding shift in the jet maximum from the Indian to the Arabian Sea also occurs.

The model contains a horizontal diffusion which tends to reduce the zonal wind shear and in so doing the barotropic instability. This diffusion was not used by Colton (1973). It is noteworthy, however, that even in this model the barotropic instability mechanism is still found to be prominent downstream from the jet maximum. An optimum use of the diffusion, to alleviate the problem associated with the finite differencing and at the same time avoid strong artificial smoothing of the basic wind shear, should give us a better simulation of this mechanism.

(b) *Fluctuating heating*—The time-mean meridional gradient of the absolute vorticity for the fluctuating heating case (not shown), as expected, strongly resembles that for the steady heating case. There is no question on the plausibility for barotropic instability to exist.

The time-section of the vorticity at 12°N (Fig. 8) also points out that the greatest wave activity is downwind of the maximum jet. The amplitude of the waves, however, appears to be greater in this fluctuating heating case, though the range of wavelengths remains the same as the steady heating case. There is also no noticeable change in the phase speed of $\sim 7{\cdot}6°$ day^{-1}. This increase in amplitude is readily apparent in the waves generated in the Atlantic area of 20°-90°W, 8°-16°N. This result indicates that the fluctuation of the total forcing field may actually enhance the barotropic energy conversion. Examination of 12-hr interval wind analysis (Fig. 9) clearly reveals that the short waves have a northwest-southeast tilt in the growing region north of the jet centre in opposition to the horizontal shear of the basic wind. As the amplitude reaches maximum downstream of the

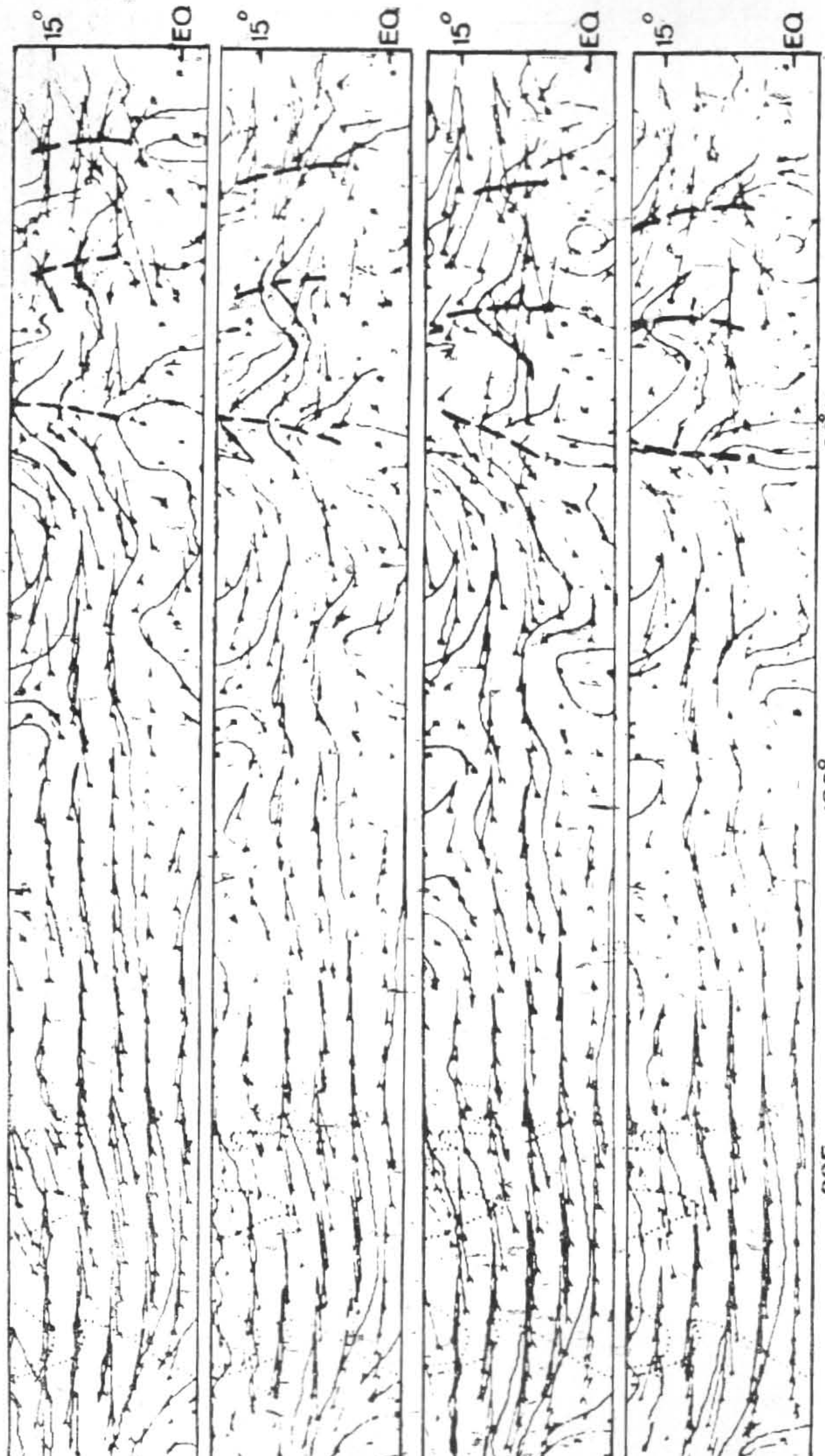

Fig. 9. Sequence (Day 22 to Day 23·5) of 12-hr interval wind fields at 250 mb, fluctuating heating

jet maximum, the tilt reverses and the waves begin to decay rather rapidly.

The stronger synoptic wave activity in the fluctuating heating case is consistent with the slightly reduced amplitude of the long waves because the synoptic scale waves draw energy barotropically from the planetary scale. Thus the fluctuating heating causes a damping of the planetary scale flow through enhanced non-linear interactions. This is different from the mechanism suggested by Holton and Colton (1973) and does not significantly alleviate the phase shift problem.

4. Conclusions

One major conclusion of this study is that the westward phase shift problem of the simulated Tibetan and Mexican highs in the absence of very strong damping is not likely to be explainable by the influence of the fluctuating heating. On the other hand, the tropical upper tropospheric troughs do not exhibit this phase problem. This may indicate the importance of damping due to cumulus transport of momentum and vorticity because cumulus convection is predominant in the upper level high pressure areas only.

In this model the upper level barotropic instability seems to be quite important for most of the area, even though the easterly jet was not developed to the fully observed strength, perhaps due to the inclusion of the somewhat strong horizontal diffusion. This instability seems particularly important when time fluctuations of the planetary scale are included through the imposed fluctuating heating. Thus the local barotropic instability at upper levels may be more important than that implied by the horizontal shear of the time-mean wind. Further investigations on this mechanism would be desireable to see whether the observed upper level disturbances, such as the cold core lows, are of this nature. Incorporation of cumulus parameterization and topography in the model should further enable us to study the possible implication of this upper level barotropic instability in the vertical development of other tropical disturbances.

In addition to the phase shift problem, Holton and Colton (1973) found very large amplitudes for long waves in the case of weak damping in their linear model. Our results do not have large amplitude discrepancy. This indicates that the transient waves due to non-linear barotropic energy conversion serve as an energy sink of the planetary-scale flow especially in the fluctuating heating case.

Acknowledgements

We wish to thank Professor R.T. Williams for valuable discussion and suggestions and Messrs A.V. Monaco, J.R. Bellis and S. Rinard for programming assistance. This research was supported by the Atmospheric Research Section, National Science Foundation, under Grant DE75-10719, and by the Naval Environmental Prediction Research Facility.

REFERENCES

Abbott, D. A. 1973 Scale interaction of forced quasi-stationary waves at low latitudes. Rep. 73-2, Department of Meteorology, Florida State Univ. Tallahassee, 190 pp.

Arakawa, A. and Mintz, Y. 1974 Notes of UCLA General Circulation Workshop, UCLA, Los Angeles.

Colton, D. E. 1973 *J. atmos. Sci.*, **30**, pp. 1287-1302.

Holton, J. R. and Colton, D. E. 1972 *Ibid.*, **29**, pp. 1124-1128.

Krishnamurti, T. N. 1971 *J. appl. Met.*, **10**, pp. 1066-1096.

Krishnamurti, T. N. and Rodgers, B. 1970 200 Millibar Wind Field June, July and August 1967, Rep. No. 70-2, Department of Meteorology, Florida State Univ., Tallahassee.

Krishnamurti, T. N. and Bhalme, H. N. 1976 *J. atmos. Sci.*, **33**, pp. 1937-1954.

REFERENCES (Contd)

Monaco, A. V. and Williams, R. T. 1975 An atmospheric global prediction model using a modified Arakawa differencing scheme. Naval Postgraduate School, Monterey, California.

Murakami, T. 1972 *J. atmos. Sci.*, **29**, pp. 1129-1137.

1974 *Ibid.*, **31**, pp. 340-357.

Sadler, J. C. 1967 The tropical upper tropospheric trough as a secondary source of typhoons and a primary source of tradewind disturbances. Final Rep., Contract —AF19 (628)—3860, Air Force Cambridge Research Labs.

Tupaz, J., Williams, R. T. and Chang, C. P. 1978 A numerical study of the locally unstable barotropic easterly jet. Proc. Int. Symp. Monsoons, *Indian J. Met. Hydrol. Geophys.*, **29**, 1 & 2.

DISCUSSION

(Paper presented by C. P. Chang)

PETER J. WEBSTER : Could you please indicate the phase speed and wavelength of your barotropic disturbances?

AUTHOR : The phase speed is 7·5° longitude per day with horizontal wavelength in the range of 2200-3500 km.