

Energetics of the monsoon circulation over south Asia : I—Diabatic heating and the generation of available potential energy

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सार — जुलाई 1963 के विशिष्ट मानसून माह के लिए दक्षिण एशिया पर उपलब्ध स्वेतिज ऊर्जा के डायबेटिक ऊष्मन तथा सृजन का अभिकलन किया गया है। इसमें मानसून परिभ्रमण पर स्थाई भंवरो (मानसून ट्रोणी आदि) और उन पर अध्यारोपित अल्पकालीन भंवरो (मानसून अवदावों, निम्न आदि) के साथ जोनल प्रवाह माध्य के रूप में विचार किया गया है। यह ω -समीकरण तथा ऊष्मा गतिक ऊर्जा समीकरण से अभिकलित ऊर्जाघन वेग से प्राप्त अभिकलनों पर आधारित है।

ऊपर की ओर ऊर्जाघन वेग के क्षेत्र अच्छी वर्षा के क्षेत्र सिद्ध हुए। क्षोभमण्डल में डायबेटिक ऊष्मन माध्य की मासिक दर 0.5° के कोटि की है और निम्न स्तरों की अपेक्षा 500 तथा 300 मिलिबार स्तरों पर उच्च है। इसके आधार पर जोनल और स्थाई दोनों प्रकार की भंवरो तथा उपलब्ध स्वेतिज ऊर्जा का अनुमान लगाया गया है।

ABSTRACT. Considering the monsoon circulation as the mean zonal flow with standing or stationary eddies (monsoon trough etc) and the transient eddies (monsoon depressions, lows etc) superimposed on it, diabatic heating and generation of available potential energy over south Asia have been computed for the typical monsoon month of July 1963, based on the computations of vertical velocity from omega-equation, and thermodynamic energy equation.

Regions of upward vertical velocity are found to be areas of good rainfall. Rate of monthly mean diabatic heating in the troposphere is of the order of 0.5 deg. C and is higher at 500 and 300 mb levels than at lower levels. Positive generation of both zonal and standing eddy available potential energy is inferred.

1. Introduction

Monsoon circulation envelopes India and its neighbourhood by July every year. The monsoon trough, a semi-permanent system, lies oriented almost east-west over the Gangetic plains from Ganganagar to Calcutta through Allahabad; its associated cyclonic circulation extends up to 500 mb level, sometimes even upto 300 mb level. To the south of this system there are west to southwest winds and to the north of this system and above 400 mb level there are easterlies. The upper level easterlies extend from the equator upto 30°N with maximum winds at 200/100 mb level over the region 10°N to 15°N .

The monsoon depressions/lows generally form over the head of the Bay of Bengal. Sometimes they form over the western Pacific Ocean and reach the Bay of Bengal after crossing Indo-China area. Although they weaken over the land, they get intensified again over the Bay of Bengal before entering the Indian mainland. These systems rarely reach the intensity of hurricanes. However, they cause widespread rainfall in the southwest sector as they move. On the average three

to four depressions form in a month and they have five to seven days of life span.

In this study an attempt is made to understand the maintenance of the monsoon circulation, the semi-permanent systems and their interactions with the transient systems, by computing their energetics. For this, the monsoon circulation is considered as the mean zonal flow with the standing eddies (monsoon troughs etc) and the transient eddies (monsoon depressions, lows) superimposed on it. It should be emphasised here that in this study the monsoon circulation itself is not considered as the perturbation of the general circulation, in which case the study should be on the global scale. In this paper which we hereafter call I, mean vertical velocity, mean diabatic heating and subsequently the generation of available potential energy are computed and studied. In the subsequent paper : Part II — Energetics of the monsoon circulation have been computed, the energy flow diagram has been constructed and discussed.

2. Computations

2.1. Symbols used (Appended)

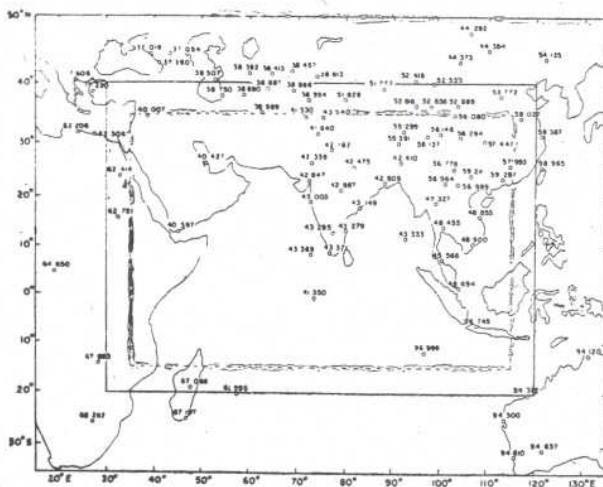


Fig. 1. Area of study with aerological stations

2.2. Data and computational procedure

Daily aerological data for July 1963 from about 90 stations (Fig. 1) covering the area from 30 deg. E to 120 deg. E and from 40 deg. N to 20 deg. S have been used in the study. July is the typical monsoon month when the summer monsoon is fully established over south Asia. In July 1963, the activity of the southwest monsoon in terms of rainfall was normal throughout India and neighbourhood. Hence the choice of July 1963 for this study. The data of morning (00 GMT) soundings were chosen in order to avoid the effect of local convection.

From the daily data for July 1963, the mean temperature \bar{T} , the mean u and v — components of the wind field, \bar{u} and \bar{v} respectively were calculated at standard isobaric levels, 1000, 850, 700, 500, 300 and 100 mb. These were plotted, analysed and grid point values were picked up at the intervals of 2.5 deg. latitude and longitude. The data were available from 30°E to 120°E and 40°N to 20°S. For the computation of vertical velocity the following ω -equation (Murakami 1957) was considered :

$$\nabla^2 \bar{\omega} + \frac{f^2}{\sigma} \frac{\partial^2 \bar{\omega}}{\partial p^2} = \frac{f}{\sigma} \frac{\partial}{\partial p} (\bar{v} \cdot \nabla \bar{\eta}) + \frac{1}{\sigma} \frac{R}{P} \nabla^2 (\bar{v} \cdot \nabla \bar{T}) \quad (A)$$

$$- \frac{f}{\sigma} \frac{\partial}{\partial p} \left(\bar{\zeta} \frac{\partial \bar{\omega}}{\partial p} - \bar{\omega} \frac{\partial \bar{\zeta}}{\partial p} \right) - \quad (C)$$

$$- \frac{f}{\sigma} \frac{\partial}{\partial p} \left(\frac{\partial \bar{\omega}}{\partial y} \frac{\partial \bar{u}}{\partial p} - \frac{\partial \bar{\omega}}{\partial x} \frac{\partial \bar{v}}{\partial p} \right) - \bar{\omega} \nabla^2 \ln \bar{\sigma} \quad (D) \quad (E)$$

$$- 2 \nabla \bar{\omega} \cdot \nabla \ln \bar{\sigma} - \frac{f}{\sigma} \frac{\partial^2}{\partial p^2} (\bar{\mathbf{k}} \cdot \nabla \bar{\omega} \times \bar{\boldsymbol{\tau}}) \quad (F) \quad (G)$$

$$- \frac{R}{c_p P \bar{\sigma}} \nabla^2 \bar{Q} - \frac{f}{\sigma} \frac{\partial^2}{\partial p \partial t} \nabla^2 \left(\bar{\psi} - \frac{g}{f} \bar{z} \right) \quad (H) \quad (I)$$

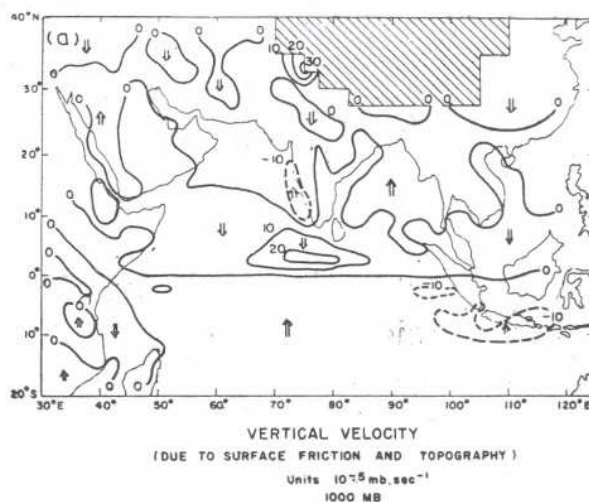


Fig. 2(a). Mean vertical velocity for July 1963 at 1000 mb

where the bar ($\bar{\quad}$) denotes the mean for the month July 1963. Other symbols are explained in Appendix. While arriving at the Eqn. (1), the terms like $\bar{\mathbf{v}} \cdot \nabla \bar{\eta}$, $\bar{\mathbf{v}} \cdot \nabla \bar{T}$ have been neglected. Although in the tropics this may not be negligibly small, but for the sake of simplicity these terms have been neglected. The terms (E) and (F) arise out of the variation in space of the monthly mean static stability. This variation in the mean static stability would be small. Also the term (I) contains time derivative, period being one month, makes this term small. Hence these three terms also are neglected. This simplification is justified in the extratropics (Murakami 1957, Asakura and Katayama 1964). In the present paper, it is extended to tropical latitudes also.

The term (G), the frictional term was included as lower boundary condition (Eqn. 2) assuming the friction in the free atmosphere to be small. In other words only surface friction and the effect of topography were taken into account.

$$\omega_{1000} = \omega_f + \omega_T \quad (2)$$

where,

$$\omega_f = \frac{g}{f} \left(\frac{\partial}{\partial y} (\rho_0 C_D u_0 V_0) - \frac{\partial}{\partial x} (\rho_0 C_D v_0 V_0) \right) \quad (3)$$

(Subscript '0' represents the values of the parameters at the surface) and

$$\omega_T = -g \rho_0 \mathbf{V} \cdot \nabla h \quad (4)$$

h is the height of the terrain obtained from Berkofsky and Bertoni (1960).

TABLE 1
Generation of Available Potential Energy
(Units : $\text{Jm}^{-2}\text{s}^{-1}$)

| Term | 1000- 850 | 850- 700 | 700- 500 | 500- 300 | 300- 100 | 1000- 100 |
|----------|--------------|-------------|-------------|-------------|-------------|--------------|
| $G(A_z)$ | 0.09 | 0.03 | 0.20 | 0.56 | 0.45 | 1.33 |
| $G(A_s)$ | 0.23 | -0.18 | 0.54 | 2.45 | 0.68 | 3.72 |

The final ω -equation after simplification becomes

$$\begin{aligned} \nabla^2 \omega + \frac{f}{\{\bar{\sigma}\}} \frac{\partial^2 \omega}{\partial p^2} &= \frac{f}{\{\bar{\sigma}\}} \frac{\partial}{\partial p} (\bar{\mathbf{V}} \cdot \nabla \bar{\eta}) + \\ &+ \frac{1}{\{\bar{\sigma}\}} \frac{R}{P} \nabla^2 (\bar{\mathbf{V}} \cdot \nabla \bar{T}) - \\ &- \frac{f}{\{\bar{\sigma}\}} \frac{\partial}{\partial p} \left(\bar{\xi} \frac{\partial \omega}{\partial p} - \bar{\omega} \frac{\partial \bar{\xi}}{\partial p} \right) - \\ &- \frac{f}{\{\bar{\sigma}\}} \frac{\partial}{\partial p} \left(\frac{\partial \omega}{\partial y} \frac{\partial \bar{u}}{\partial p} - \frac{\partial \omega}{\partial x} \frac{\partial \bar{v}}{\partial p} \right) - \\ &- \frac{R}{c_p P \{\bar{\sigma}\}} \nabla^2 \bar{Q} \end{aligned} \quad (5)$$

where $\{\bar{\sigma}\}$ denotes average over the limited region under study.

The Himalayan mountains are considered in this study as a solid mass extending from surface to 500 mb level and shown as hatched area in Fig. 2(a). It is assumed that inside this region and at the periphery, the advections of temperature, vorticity etc are zero and that the vertical velocity is zero only inside the region and is non-zero at the periphery. The vertical velocity is assumed to be zero at 100 mb level, and at lateral boundaries at all levels.

The thermodynamic energy equation (Eqn. 6)

$$\begin{aligned} \frac{\partial}{\partial t} \left(-\frac{\partial \bar{z}}{\partial p} \right) + \bar{\mathbf{V}} \cdot \nabla \left(-\frac{\partial \bar{z}}{\partial p} \right) - \bar{\omega} \frac{\{\bar{\sigma}\}}{g} \\ = \frac{R}{gP} \frac{\bar{Q}}{c_p} \end{aligned} \quad (6)$$

and the ω -equation (Eqn.5) were combined to obtain the heating field in the following way. ω_a (adiabatic vertical velocity) was computed first from Eqn. (5) neglecting heating term. Using ω_a , the approximate value of heating, \bar{Q}/c_p was computed from Eqn. (6). Using this \bar{Q}/c_p in Eqn. (5), ω_d (diabatic vertical velocity) was computed and subsequently ω_d was used in Eqn. (6) to compute more accurate \bar{Q}/c_p . This procedure was repeated until two successive computations of \bar{Q}/c_p at the corresponding grid points differed by less than 0.1% of the average value. The convergence of this procedure was fast and within 20 cycles, the convergence had been achieved. Both

the diabatic vertical velocity, ω_d and the mean diabatic heating \bar{Q}/c_p obtained in the last cycle are shown in Figs. 2 and 3 respectively.

Eqns. (7), (8) and (9) give the expressions for the generation of zonal, standing eddy or stationary eddy and transient eddy available potential energy (Lorenz 1955).

$$\begin{aligned} G(A_z) &= \int_M \frac{1}{\{\bar{\sigma}\}} \left(\frac{R}{P} \right)^2 \frac{1}{c_p} [\bar{Q} - \{\bar{Q}\}] \times \\ &\times [\bar{T} - \{\bar{T}\}] dM \end{aligned} \quad (7)$$

$$G(A_s) = \int_M \frac{1}{\{\bar{\sigma}\}} \left(\frac{R}{P} \right)^2 \frac{1}{c_p} [\bar{Q}^* \bar{T}^*] dM \quad (8)$$

$$G(A_T) = \int_M \frac{1}{\{\bar{\sigma}\}} \left(\frac{R}{P} \right)^2 \frac{1}{c_p} \bar{Q}^* \bar{T}^* dM \quad (9)$$

$\{\bar{\sigma}\}$ is the static stability averaged over the area under study. As explained in detail in part II of this paper (Rajamani—see Ref.) the zonal available potential energy can be partitioned as the barotropic and the baroclinic contributions, the latter being proportional to the variance in temperature computed using space mean over limited region. Hence it is appropriate to consider $G(A_z)$ given in Eqn. (7) to represent the generation of that part which is the baroclinic contribution to the zonal available potential energy. Using Eqns. (7) and (8), the generations of zonal and standing eddy available potential energy have been computed. But the generation of transient eddy available potential energy (Eqn. 9) was not calculated as it would require the heating field for each day of the period. The generation terms were calculated for the small layer like 1000-850 mb etc before summing up to get the term for the entire troposphere from 1000 to 100 mb level (Table 1).

3. Discussion of results

3.1. Vertical velocity

Figs. 2(a-d) depict the vertical velocity field at 1000, 850, 700, 500 and 300 mb levels. The magnitude of vertical velocity is of the order of a few mm. s^{-1} . Such small values are to be expected, as the monthly mean synoptic patterns are very smooth compared to daily synoptic patterns.

Fig. 2(a) shows the ω -field for 1000 mb which is due to surface friction and topography. This figure portrays clearly the effect of the Western Ghats in the Peninsular India giving rise to a region of upward motion on the windward side, i.e., western side of the Ghats. The maximum value of $1.7 \times 10^{-4} \text{mb s}^{-1}$ is at 10 deg. N, 75 deg. E west of Nilgiri hills which is the highest part of the Western Ghats. The vertical velocity becomes zero on the leeward side approximately along 77.5 deg. E. Further east, there is a region of descending motion.

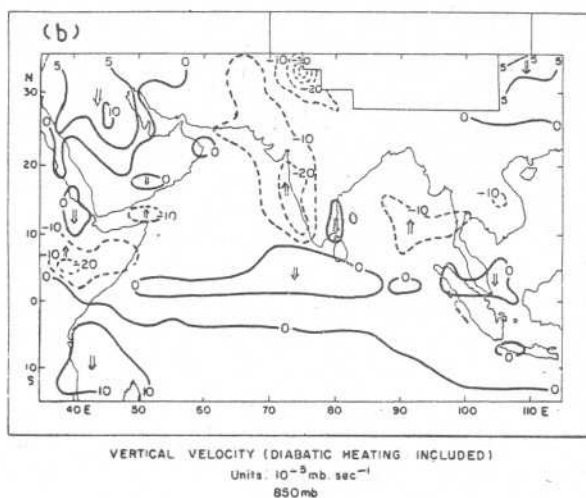


Fig. 2(b). Mean vertical velocity for July 1963 at 850 mb (10^{-5} mb/s)

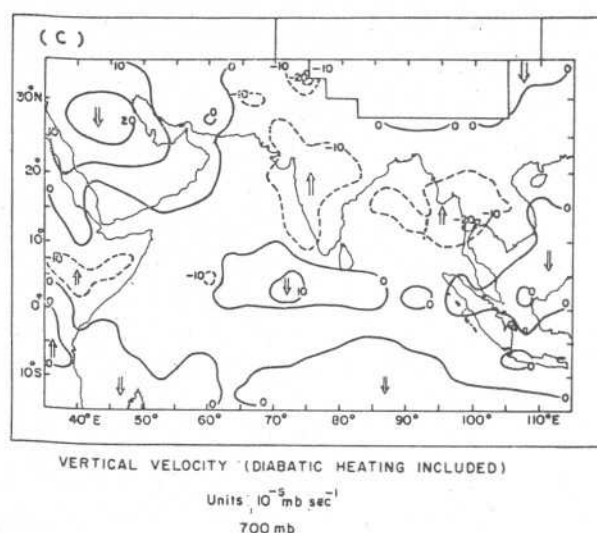


Fig. 2(c). Mean vertical velocity for July 1963 at 700 mb (10^{-5} mb/s)

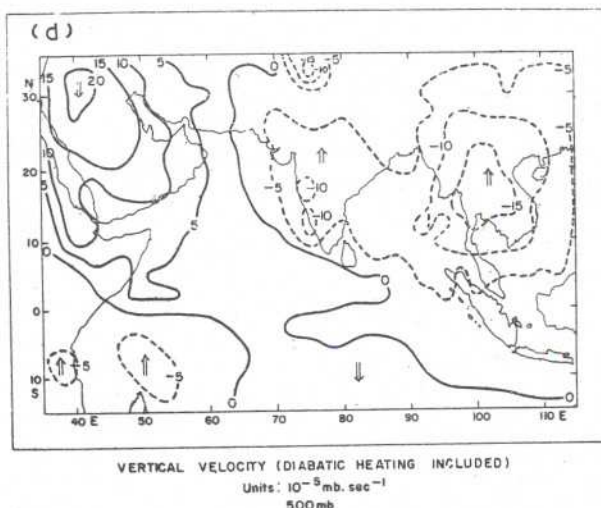


Fig. 2(d). Mean vertical velocity for July 1963 at 500 mb (10^{-5} mb/s)

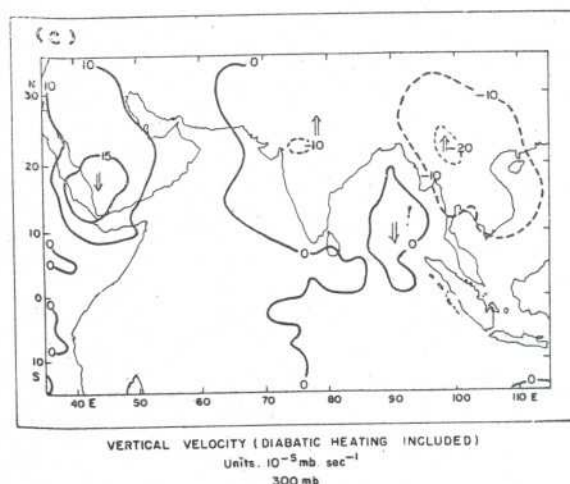


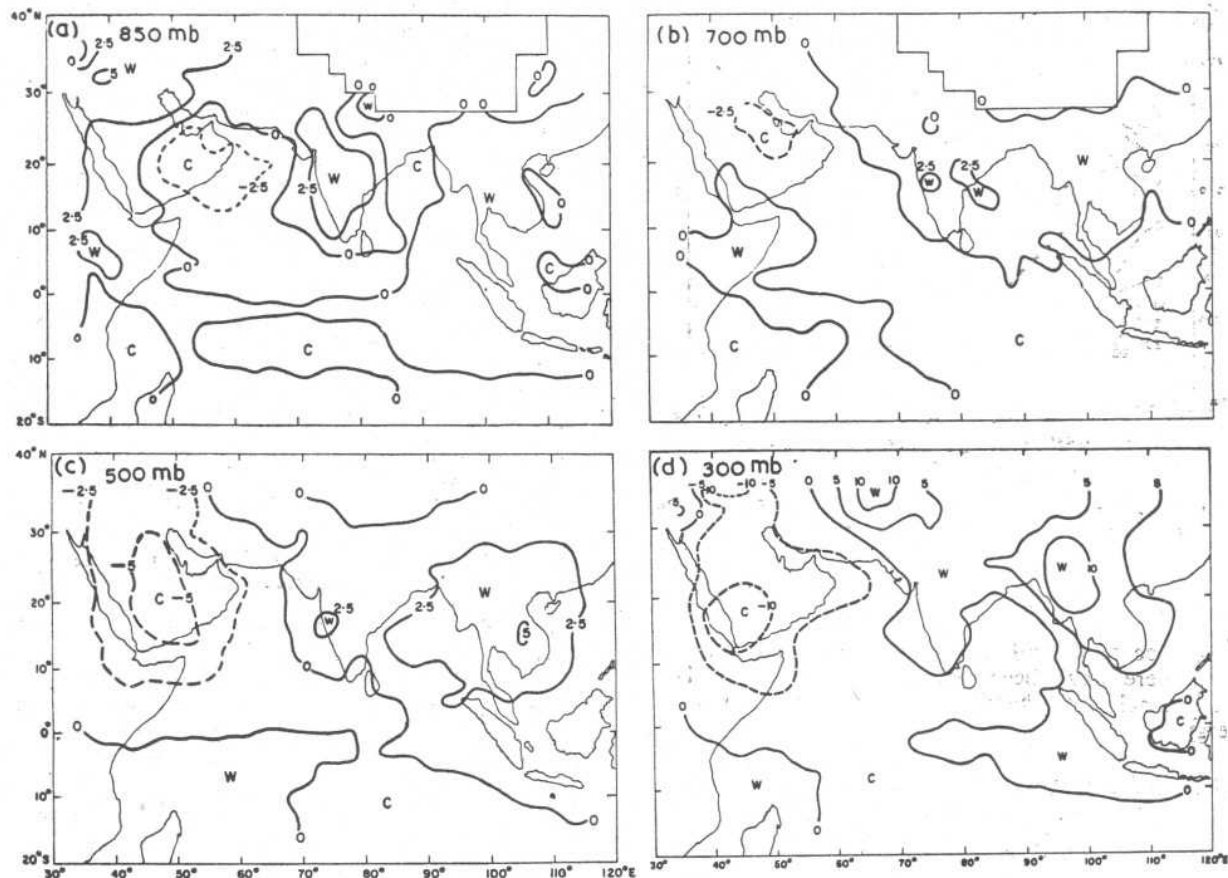
Fig. 2(e). Mean vertical velocity for July 1963 at 300 mb (10^{-5} mb/s)

Over the oceanic regions the vertical velocity is solely due to the surface friction. Over the Bay of Bengal, where the southwesterlies turn cyclonically, there is upward motion. In the region over the Arabian Sea bounded by 65°E and 85°E and 10°N and equator there is a downward motion where wind turns in clockwise direction. South of the equator, mostly there is upward motion, although small in magnitude.

At 850 and 700 mb levels (Figs. 2b and 2c) there is upward motion over India, Indo-China area and also over Somali coast and adjacent areas, and downward motion over Iran, Iraq, Arabia, etc and over most parts of the regions south of the equator. The vertical velocity field is weak south of the equator. At 500 and 300 mb levels (Figs. 2d and 2e), the region of downward motion extends further eastwards and southwards so that there is down-

ward motion west of 65°E (except for a small region from equator to 15°S and from 35°E to 65°E in case of 500 mb level) and ascending motion east of 65°E , i.e., over India and Indo-China area (except for a small region in the central Bay of Bengal at 300 mb level). In general the computed vertical velocity field shows upward motion over the rainy regions of south Asia, viz., Indo-China region, Indian region, etc. This justifies to some extent the simplification made in the ω -equation used in this study.

With the westerly winds in the lower troposphere, the upward motion over India and Indo-China area, the easterly winds in upper troposphere and downward motion over Arabia, Iran, Iraq etc the Walker circulation or the east-west circulation in x - p plane, is further corroborated.



Figs. 3 (a--d). Mean heating for July 1963 at different levels (10^{-1} deg. C/day) : (a) 850 mb, (b) 700 mb, (c) 500 mb, and (d) 300 mb

3.2. Heating fields

Heating field was computed for 850, 700, 500 and 300 mb levels; the heating is of the order of 0.5 deg. C per day. At 850 mb level (Fig. 3a) there is heating over India, Indo-China area and the area comprising Iran, Iraq and part of Arabia. Figs. 3 (c) and 3 (d) show that at higher levels also there is heating over India and Indo-China area. Over most parts of the oceanic areas there is cooling. The heating/cooling rates are large at higher levels (500 and 300 mb levels) and the heating field is similar to the vertical velocity field. This agrees with the finding of Reed and Recker (1971) who noticed that the strongest heating occurs at the levels 500-400 mb and that the vertical velocity is the most dominant of all the terms which contribute to diabatic heating.

It could be seen from these figures [Figs. 3 (a-d)] that in general over the regions of good rainfall, heating has taken place in spite of the radiative cooling that generally prevails. This suggests that the release of latent heat of condensation plays an important role.

Fig. 4 shows the meridional cross section of heating rates. There is heating north of 5 deg. N upto 35 deg. N at 850, 700 and 500 mb levels except for a small region from 25 deg. N to 30 deg. N at 700 mb and from 25 deg. N to 35 deg. N at 500 mb level. At

300 mb level, from the equator to 25 deg. N, there is cooling and north of 25 deg. N upto 35 deg. N there is heating at higher rates. Fig. 4 suggests that heating is more dominant than cooling of the atmosphere over the monsoon region. The mean heating over the column from 1000 mb to 100 mb of unit area of cross section is 0.2 deg. C per day (approx).

3.3. Generation of available potential energy

Table 1 shows that in the case of standing eddy, there is destruction of available potential energy in the layer 850-700 mb. However, in all other layers, there have been positive generation resulting in a net value of $3.7 \text{ J/m}^2/\text{s}$. In the case of zonal current, available potential energy of the order $1.3 \text{ J/m}^2/\text{s}$ is generated. Evidently, the generation of available potential energy is larger in the case of standing eddies than in the case of zonal current.

4. Conclusion

Although ω -equation used in this study has been simplified to facilitate the computations, the mean vertical velocity field obtained for July 1963 has been satisfactory with the regions of upward motion agreeing with rainy regions. Also this brings out the feature of a Walker circulation or east-west circulation over south Asia. Mean heating field obtained suggests the

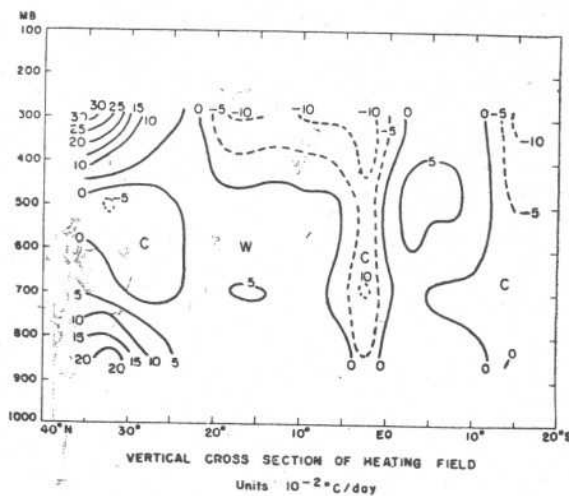


Fig. 4. Vertical cross-section of mean heating field for July 1963 (10^{-2} deg. C/day)

importance of the release of latent heat of condensation. There have been positive generations of both mean zonal and standing eddy available potential energy.

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Appendix

- A — Total Available Potential Energy (APE),
 A_Z — Zonal APE,
 A_S — Standing eddy APE,
 A_T — Transient eddy APE,
 K_Z — Zonal Kinetic Energy (KE),
 K_S — Standing eddy KE,
 K_T — Transient eddy KE,
 M — Mass of atmosphere over the area under study,
 σ — Static stability,
 R — Gas constant
 P — Pressure in mb
 T — Temperature
 u — x component of wind vector (eastward)
 v — y component of wind vector (northward)
 ω — Vertical wind component (dp/dt) in isobaric co-ordinates
 V — Wind vector
 \dot{Q} — Rate of heating
 g — Acceleration due to gravity
 c_p — Specific heat at constant pressure
 f — Coriolis parameter
 ζ — Relative vorticity
 η — Absolute vorticity
 k — Unit vector in the vertical

- τ — Wind stress vector
 ψ — Stream function
 C_D — Drag coefficient
 ρ_0 — Density of air at the surface
 h — Height of topography
 α — Specific volume
 $(\overline{})$ — Hemispherical mean
 $(\overline{})$ — Time mean (mean of July 1963)
 $[x]$ — Zonal mean of x

$$\frac{1}{\int_{35^{\circ}\text{E}}^{115^{\circ}\text{E}} d\lambda} \int_{35^{\circ}\text{E}}^{115^{\circ}\text{E}} x d\lambda$$

{x} — Mean over the area under study

$$\frac{1}{\int_{5^{\circ}\text{N}}^{35^{\circ}\text{N}} \int_{35^{\circ}\text{E}}^{115^{\circ}\text{E}} a^2 \cos \phi d\lambda d\phi} \int_{5^{\circ}\text{N}}^{35^{\circ}\text{N}} \int_{35^{\circ}\text{E}}^{115^{\circ}\text{E}} x a^2 \cos \phi d\lambda d\phi$$

- (') — Deviation from time mean
 (*) — Deviation from zonal mean
 ω_f — Vertical velocity due to frictional term
 ω_T — Vertical velocity due to topography.