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On the mass, heat and moisture balance in the field of monsoon depressions

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सार — मानसून के दो विशिष्ट महीनों यानी जुलाई एवं अगस्त में बंगाल की खाड़ी में बने 27 मानसून अवदाबों के आंकड़ों और उनके पश्चिमी से पश्चिमउत्तरपश्चिमी दिशा में सुपरिचित गमन के आधार पर संहति का माध्य पार्श्व अभिगमन, संवेद्य ऊष्मा, गुप्त ऊष्मा एवं नमी की गणना की गई है । अवदाब केन्द्र से दूरी के परिपेक्ष में अन्तर्वाह एवं बहिर्वाह परतों की गहराइयों का बंटन ज्ञात किया गया है । अधःशायी सतह के स्थानान्तरणों सहित कुल प्रभावी ऊर्जा के उर्ध्वाधर अभिवाहों का आकलन किया गया है । अवदाब क्षेत्र के शीर्ष से ऊर्जा में हुई हानि को ज्ञात किया गया है और पक्षाभ स्तर पर कृष्णिका विकिरण के प्रत्याशित मान से उसकी तुलना की गई है । सतह की ऊष्मा पूर्ति से उपजे संवहन तत्वों की सैद्धान्तिक संभाविता, जो केवल उत्प्लावन बलों के कारण उपजी है, ज्ञात की गई है । सतह के ऊष्मा स्रोतों एवं अभिगमों, उनकी क्षेत्रीय सीमा, अवरोहण दर और कम होती वायु राशि की वाष्पन मात्रा को ज्ञात कर उनका विवेचन किया गया है ।

इस अभिधारणा के साथ कि सारी नमी परिवर्तित होकर वर्षा बन गई है, अवदाब केन्द्र से अलग-अलग दूरियों पर अधिकतम वर्षा को ज्ञात किया गया है । परिसीमा स्तर पर त्वचीय घर्षण के कारण घर्षण ह्रास की गणना करके अवदाब के भीतर नेट प्रभावी ऊष्मा एवं प्राप्त गुप्त ऊष्मा के पदों में इसका निर्धारण किया गया है ।

ABSTRACT. Based on data of 27 monsoon depressions which formed in the Bay of Bengal during the two typical monsoon months, *i.e.*, July and August and moved in the customary westerly to westnorthwesterly direction, mean lateral transports of mass, sensible heat and latent heat and moisture have been calculated. Distribution of depths of inflow and outflow layers with distance from the depression centre has been obtained. Vertical fluxes of the total effective energy including the transfers from the underlying surface are estimated. Magnitude of energy loss from the top of the depression field have been obtained and compared with what may be expected from the black body radiation at the cirrus level. Theoretically possible extent of rise of the convective elements from the surface heat supply, rising exclusively out of buoyancy forces has been determined. Heat sources and sinks, their areal extent, the rate of descent and the amount of evaporation the subsiding airmass would evaporate from the surface have been obtained and discussed.

Maximum amounts of rainfall that could occur on the assumption that all converging moisture precipitates out, has been obtained for different distances from the depression centre. Frictional loss due to skin friction at the boundary layer has been computed and expressed in terms of the net effective heat as well as the latent heat gained within the depression.

1. Introduction

Various general circulation and numerical studies in the past have unmistakably brought out tropics as the heat source in maintaining the global energy balance in the atmosphere. In the tropical areas the summer monsoon trough over India and neighbourhood is the best known quasi-stationary perturbation of the general circulation. The mean pattern of the trough is sometimes distorted by eddies in its circulation known as monsoon depressions. These depressions, because of copious rainfall associated with them, exert profound impacts on the national economies. The latent heat released in the condensation process constitutes an important component in the thermal balance of monsoon depressions. The extensive thick cloud diminishes the insolation from reaching the surface, thus reducing the role of surface heating in this tropical system as a heat source. The southwest sector of the depression is a strong convergence zone and the resulting intense convection carries moisture deep into the upper troposphere and the rate of transport is also high. The outflow of the sensible heat is carried away by the upper easterlies. The budget of various parameters of the depression field is, therefore, worth investigation as this has not been much explored and satisfactory analysis not yet been made. In contrast, the energy balance in tropical cyclones (Palmen and Riehl 1957), the tropical monsoon trough zone (Berson 1961), equatorial trough zone (Riehl and Malkus 1958), Indian monsoon trough zone (Anjaneyulu 1969) have attracted considerable attention.

Palmen and Riehl (1957) studied average properties of tropical cyclones and found that the kinetic energy production within the cyclone can take place only if

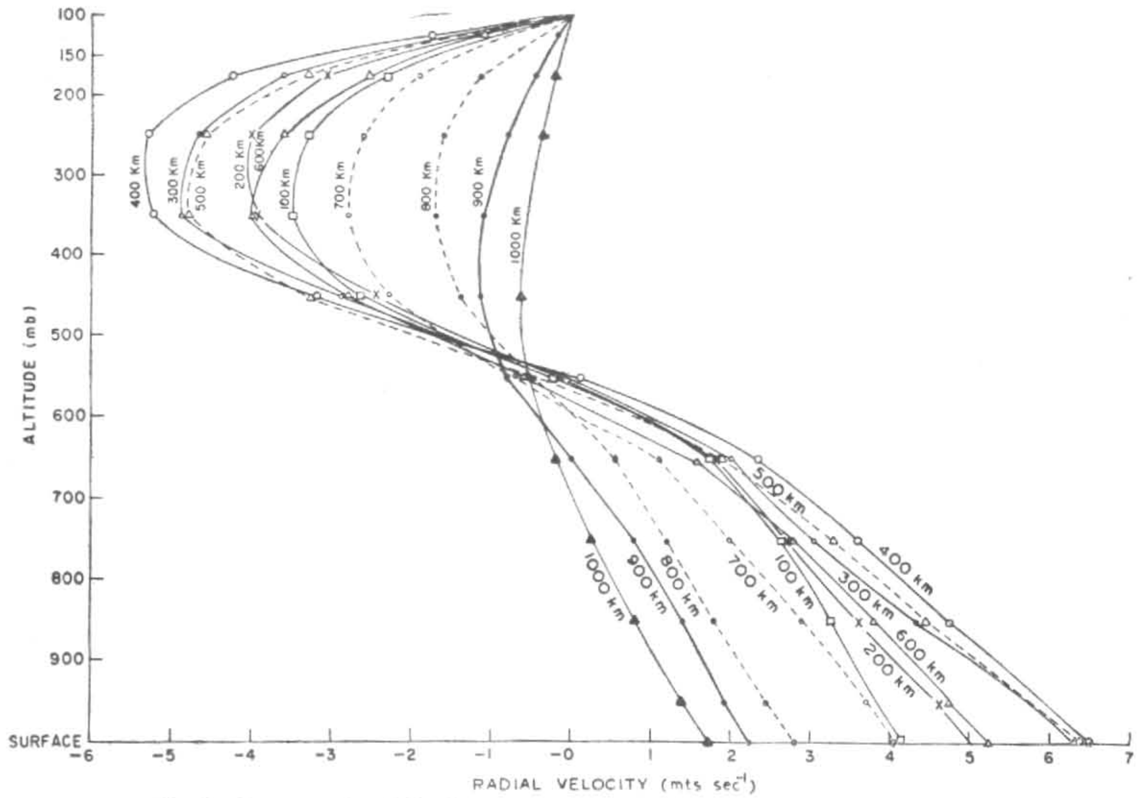


Fig. 1. Pattern of radial velocity against altitude (mb) for various radii circles

TABLE 1

The tangential velocity, V_θ radial velocity V_r (both in metres per sec), temperature T (in °A, 2 omitted) and humidity mixing ratio x (g/kg)

	100	200	300	400	500	600	700	800	900	1000 km
					850 mb					
V_θ	9.9	11.3	13.6	14.3	13.3	12.0	10.5	9.3	8.1	6.9
V_r	3.3	3.6	4.3	4.7	4.5	3.8	2.9	1.8	1.4	0.8
T	97.3	96.5	94.5	94.5	94.5	94.9	95.1	95.9	95.9	95.7
x	15.4	15.5	16.1	16.1	15.7	15.4	15.4	14.1	13.5	13.1
					500 mb					
V_θ	5.9	6.6	8.1	9.7	8.1	6.5	5.3	4.1	2.9	2.1
V_r	−1.5	−1.3	−1.5	−1.5	−1.9	−1.8	−1.5	−0.9	−0.9	−0.6
T	72.3	71.9	70.9	71.3	71.5	70.9	71.3	71.7	71.3	70.4
x	0.6	0.8	1.0	0.7	0.6	0.6	0.4	0.3	0.2	0.1
					300 mb					
V_θ	1.3	1.7	2.1	1.8	0.3	−0.8	−3.2	−2.7	−3.5	−3.5
V_r	−3.4	−3.9	−4.8	−5.3	−4.7	−3.9	−2.8	−1.7	−0.9	−0.4
T	46.1	46.1	46.5	46.5	46.6	46.3	46.9	46.5	46.9	46.9

Note : Positive V_θ means cyclonic and negative denotes anticyclonic. Positive values of V_r stands for inflow and negative for outflow. Columns denote distances from the centre of the mean depression.

it is of the warm core type. They showed that about 3% of the latent heat released is converted to kinetic energy and that a considerable export of potential *plus* internal energy occurs from tropical storms. Riehl and Malkus (1958) constructed energy budgets for 10 deg. latitude belts from the equatorial trough, polewards. They showed that a total of 1500-5000 active undilute cloud towers around the globe in the trough are adequate to balance its heat loss and provide for the export. They found that neither a gradual mass circulation nor simple convection — diffusion can transport heat upwards to balance the loss aloft. Berson (1961) examined the energy in the tropical monsoon trough from data of Malaysia, Indonesia and north Australia. He found mass circulation to be a determining factor of the horizontal transport of energy. Anjaneyulu (1969) estimated the mean heat and moisture over the Indian monsoon trough zone based on mean aerological data for July & August months. He noticed that the import of sensible heat is far less than the export and concluded that there exists a net surplus of sensible heat in the trough zone.

In the present study, budgets of mass, heat and moisture have been examined in the mean monsoon depression field and the flux of heat energy has been evaluated. This has been done for various radial distances of the field from the depression centre. The work involves volume integral of mass circulation, moisture advection and the energy equation and calculation of fluxes by line integral around the periphery of different circles. Velocities (tangential and radial), temperature and moisture field of the mean depression are given in Table 1.

2. Basic data

The data of 27 monsoon depressions which formed and moved in westerly or westnorthwesterly direction during July and August months between 1961 and 1974 have been composited to obtain the mean fields in Table 1. A distance of 1000 km from the surface position of the depression centre has been considered as the limit of the depression field. For the sake of analysis the depression field is divided into 10 annular rings each of 100 km width. The upper air temperature and humidity data are collected for 00 GMT soundings of 14 radiosonde stations within the fields of the various depressions. Wind data of about 35 pilot balloon stations and nearly 25 radio-wind stations have been utilised.

2.1. *List of notations used and their explanations*

A	Reciprocal of the mechanical equivalent of heat
c_p	Specific heat of air at constant pressure
F	Frictional force
K	Constant
L	Latent heat
Q_e	Surface latent heat source
Q_p	Condensation heat released
Q_s	Surface sensible heat source
R	Gas constant for air
R_a	Radiation deficit of the troposphere
R_e	Radiation excess received at the surface
T	Absolute temperature
$\mathbf{V}$	Wind velocity vector; V, the scaler velocity
W	Heat loss due to radiational cooling
dE	Element of total effective energy
dh	Element of heat added per unit mass
dp	Element of atmospheric pressure
dt	Element of time
$d\alpha$	Element of volume
g	Acceleration due to gravity
l	Periphery of the line integration; dl its element
q	Specific humidity, dq its element
V_r	Radial velocity component
V_θ	Tangential velocity component
∇	Del-operator
ρ	Density of air
σ	Stefan-Boltzman constant
w	Vertical velocity

3. Radial mass flow

A single parameter which is extremely important in evaluating the thermodynamic characteristics of any weather system is the distribution of radial velocity with distance and height. Various budget studies (mass, moisture, momentum, energy etc) also demand a fairly accurate determination of radial velocity. In the present case, radial velocities were computed, according to the methodology suggested by Jordon (1952) for each radial distance from the wind speed and wind direction of the composited data. Mean values have been obtained from the composite technique of George (1975). Radial transport due to eddy motion has not been considered. The mass flow can be obtained from the mass balance equation

$$\int_l \int_{\text{sfc}}^{100\text{ mb}} \frac{V_r}{g}\, dl \,.\, dp = 0$$

Curves in the outflow layer, for which the data were not large enough, were drawn under the following presumptions :

(*i*) radial outflow becomes zero at 100 mb,

(*ii*) the mass outflow across any radius exactly balances the mass inflow across the radius,

(*iii*) for any level the radial mass transport is proportional the radial velocity.

For different radii, vertical profiles of the radial flow are given in Fig. 1. Positive values are inwards and negative outwards. It may be pointed that, in order to satisfy mass continuity and for steady state, the integral of radial velocity (and hence the mass flow) from surface to 100 mb, must vanish at any radius.

TABLE 2

Horizontal flux of mass, sensible heat (c_pT+Agz) latent heat (Lq), and total effective heat ($c_pT+Agz+Lq$)

Layer (mb)	Mass (unit 10^{18} gm day^{-1})	Heat flux (units 10^{15} cal day^{-1}) c_pT+Agz	Lq	$c_pT+Agz+Lq$	Mass (unit 10^{18} gm day^{-1})	Heat flux (units 10^{15} cal day^{-1}) c_pT+Agz	Lq	$c_pT+Agz+Lq$
		(a) 100 km distance				**(b) 200 km distance**		
Sfc -900	0.22	16.17	2.53	18.70	0.51	37.38	5.97	43.35
900-800	0.18	13.45	1.49	14.94	0.40	29.80	3.36	33.16
800-700	0.15	11.35	0.84	12.19	0.29	21.95	1.62	23.57
700-600	0.09	6.94	0.30	7.24	0.19	14.63	0.63	15.26
600-500	—0.03	—2.35	—0.04	—2.39	—0.01	—0.78	—0.01	—0.79
500-400	—0.15	—11.94		—11.94	—0.27	—21.49		—21.49
400-300	—0.19	—15.43		—15.43	—0.44	—35.73		—35.73
300-200	—0.18	—14.96		—14.96	—0.44	—36.61		—36.61
200-150	—0.06	—5.11		—5.11	—0.16	—13.65		—13.65
150-100	—0.03	—2.61		—2.16	—0.07	—6.10		—6.10
Sfc -500		45.56	5.12	50.68		102.98	11.57	114.55
500-100		—50.05	0	—50.05		—113.58	0	—113.58
Sfc-100		—4.49	5.12	0.63		—10.60	11.57	0.97
		(c) 300 km distance				**(d) 400 km distance**		
Sfc-900	0.96	69.60	11.62	81.22	1.31	95.11	16.51	111.62
900-800	0.72	53.35	6.26	59.61	1.05	77.81	9.13	86.95
800-700	0.51	38.51	3.06	41.57	0.79	59.65	4.66	64.31
700-600	0.33	25.31	1.12	26.43	0.52	39.88	1.77	41.65
600-500	—0.04	—3.12	—0.06	—3.18	0.03	2.35	0.04	2.39
500-400	—0.47	—37.37		—37.37	—0.71	—56.52		—56.52
400-300	—0.81	—65.85		—65.85	—1.16	—94.31		—94.31
300-200	—0.78	—64.97		—64.97	—1.17	—97.46		—97.46
200-150	—0.30	—25.62		—25.62	—0.47	—40.14		—40.14
150-100	—0.12	—10.46		—10.46	—0.19	—16.59		—16.59
Sfc -500		183.65	22.00	205.65		274.80	32.11	306.91
500-100		—204.27	0	—204.27		—305.02	0	—305.02
Sfc -100		—20.62	22.00	1.38		—30.22	32.11	1.89
		(e) 500 km distance				**(f) 600 km distance**		
Sfc-900	1.61	116.89	19.32	136.21	1.57	114.61	17.84	132.35
900-800	1.23	91.14	10.45	101.59	1.26	93.49	10.46	103.95
800-700	0.91	68.71	5.01	73.71	0.93	70.03	5.30	75.33
700-600	0.53	40.76	1.70	42.46	0.53	40.60	1.75	42.32
600-500	—0.13	—10.17	—0.14	—10.31	—0.20	—15.60	—0.24	—15.84
500-400	—0.90	—71.64		—71.64	—0.96	—76.42		—76.42
400-300	—1.34	—108.81		—108.81	—1.33	—108.00		—108.00
300-200	—1.27	—106.05		—106.05	—1.21	—100.79		—100.79
200-150	—0.46	—39.33		—39.33	—0.42	—35.83		—35.83
150-100	—0.18	—15.71		—15.71	—0.17	—14.84		—14.84
Sfc-500		307.33	36.34	343.67		303.13	35.11	338.24
500-100		—341.54	0	—341.54		—335.88	0	—335.88
Sfc- 100		—34.21	36.34	2.13		—32.75	35.11	2.36
		(g) 700 km distance				**(h) 800 km distance**		
Sfc -900	1.44	105.12	15.84	120.96	1.09	79.79	11.77	91.56
800-900	1.12	83.10	9.30	92.40	0.79	58.85	6.00	64.85
800-700	0.78	58.89	4.29	63.18	0.54	40.88	2.70	43.58
700 600	0.42	32.30	1.13	33.43	0.24	18.50	0.58	19.08
600-500	—0.25	—19.55	—0.20	—19.75	—0.23	—18.03	—0.14	—18.17
500-400	—0.89	—70.84		—70.84	—0.62	—49.35		—49.35
400-300	—1.09	—88.62		—88.62	—0.75	—60.90		—60.90
300-200	—1.03	—86.10		—86.01	—0.71	—59.21		—59.21
200-150	—0.36	—30.78		—30.78	—0.25	—21.37		—21.37
150-100	—0.14	—12.22		—12.22	—0.10	—8.74		—8.74
Sfc-500		259.86	30.36	290.22		179.99	20.91	200.90
500-100		—288.47	0	—288.47		—199.57	0	—199.57
Sfc -100		—28.61	30.36	1.75		—19.58	20.91	1.33
		(i) 900 km distance				**(j) 1000 km distance**		
Sfc -900	0.97	70.91	10.38	81.29	0.78	57.17	7.96	65.13
900-800	0.70	52.15	5.11	57.26	0.44	32.74	3.12	35.86
800-700	0.40	30.28	1.88	32.16	0.14	10.58	0.59	11.17
700-600	0	0	0	0	—0.11	—8.46	—0.21	—8.67
600-500	—0.40	—31.32	—0.16	—31.48	—0.30	—23.43	—0.09	—23.52
500-400	—0.57	—45.31		—45.31	—0.36	—28.58		—28.58
400-300	—0.55	—44.71		—44.71	—0.31	—25.23		—25.23
300-200	—0.40	—33.40		—33.40	—0.21	—17.53		—17.53
200-150	—0.11	—9.41		—9.41	—0.05	—4.27		—4.27
150-100	—0.04	—3.50		—3.50	—0.02	—1.75		—1.75
Sfc-500		122.02	17.21	139.23		68.60	11.37	79.97
500-100		—136.33	0	—136.33		—77.36	0	—77.36
Sfc -100		—14.31	17.21	2.90		—8.76	11.37	2.61

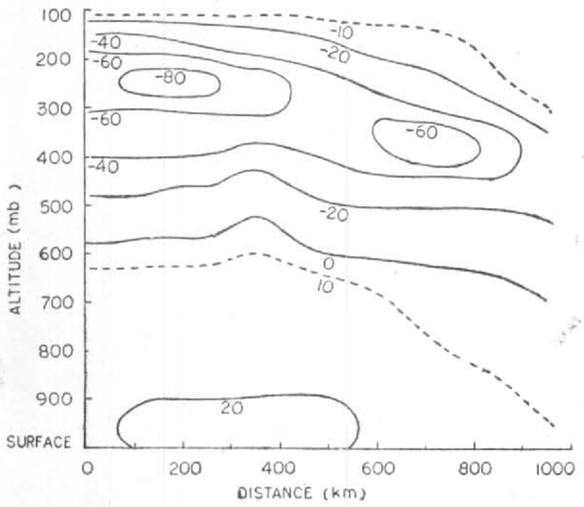


Fig. 2. In-curvature field (in degree) (Positive values denote inflow and negative values outflow)

The mass flow profiles reveal some interesting features. The inflow layer is well defined and for all practical purposes restricted to a depth between the surface and about 550 mb. The inflow level rises and spreads out reaching a peak height of about 550 mb at 300 km and then descends gradually beyond 300 km. In the outer field of the depression after 800 km, the inflow is confined to below 600 mb and at 1000 km, it is 700 mb.

The changes in the magnitude of mass inflow across adjacent radii are found to be gradual and smooth. The inflow increases with decreased elevation through most of the friction layer — this observation was in fact an assumption made by several other workers like Miller (1962, 1964, 1966), Riehl and Malkus (1958), Rosenthal (1961) etc in their study of Atlantic hurricanes.

The maximum outflow in the outflow layer was invariably noticed at 300 mb upto 800 km, descending to 400 mb at 900 km and 500 mb at 1000 km from the centre. Another feature is the sudden and sharp decrease in the outflow at elevations above 200 mb, noticeable close to the depression centre and beyond 800 km.

If the inflow occurs in the lower troposphere say upto about 500 mb and in the mid-troposphere there is an inactive layer, *i.e.,* a layer of neither outflow or inflow, this indicates absence of mid-troposphere cooling, and has been termed ventilation, by Simpson and Riehl (1958). Presence of inflow upto 500 mb and active ventilation in the mid-troposphere is perhaps a necessary condition for the development and maintenance of smaller and less intense tropical weather systems like depressions. Riehl and Malkus (1961) also observed an inflow upto 500 mb and active ventilation in the mid-troposphere in hurricane *Daisy* which according to them was a weak cyclone compared to other Atlantic hurricanes.

The angle of inflow and outflow at various levels have their own importance in the study of inflow-ventilation-outflow mechanism. In the present study radial and tangential velocities have been utilised to compute the in-curvature or the indraft angles and their spatial distribution depicted in Fig. 2. The in-curvature has relatively uniform maximum near the centre of depression and minimum at the outer boundary. From this one may be tempted to assume that much of the assymetry in the storm is due to its motion. The indraft angles attains their highest values in the lowest troposphere decreasing vertically upwards upto about 550 mb. In the lower troposphere the mean indraft angles are found remarkable constant upto 500 km of the centre beyond which they decrease abruptly. The values close to the surface level agrees well with those found by Hughes (1952) for Atlantic hurricanes though the angles are found somewhat smaller. In the outflow layer the outflow angles increases gradually upwards. Their largest values of about 80 deg. are observed between 100 & 300 km at an altitude of about 250 mb. Another zone of somewhat higher angle of outflow is located at relatively lower altitudes of 300-400 km, towards the outer envelope of the field.

4. Budget of the effective energy

4.1. The basic elements which govern the heat and moisture fluxes in any circulation system are the temperature, humidity and the wind velocity. In the present study magnitude of these parameters have been included for three representative levels, *viz.,* 850 mb, representing lower troposphere, 500 mb the middle troposphere and 300 mb for the upper troposphere in Table 1.

At any level not much variations are observed in temperature and humidity which can be assumed to be nearly constant. However, variations in tangential and radial velocities are quite substantial and it is these variations which contribute to the observed changes in the structure of the fluxes in the depression field.

4.2. *The heat balance equations*

The first law of thermodynamics can be written as

$$dE = c_p\, dT - \frac{RT}{\rho}\, dp + L\, dq$$

$$i.e.,\ \rho.\ dE = \rho\, (c_p dT + L dq) - dp \qquad (1)$$

Rate of change is given by :

$$\rho\, \frac{dE}{dt} = \rho \left(\frac{c_p\, dT}{dt} + L\, \frac{dq}{dt} \right) - \frac{dp}{dt}$$

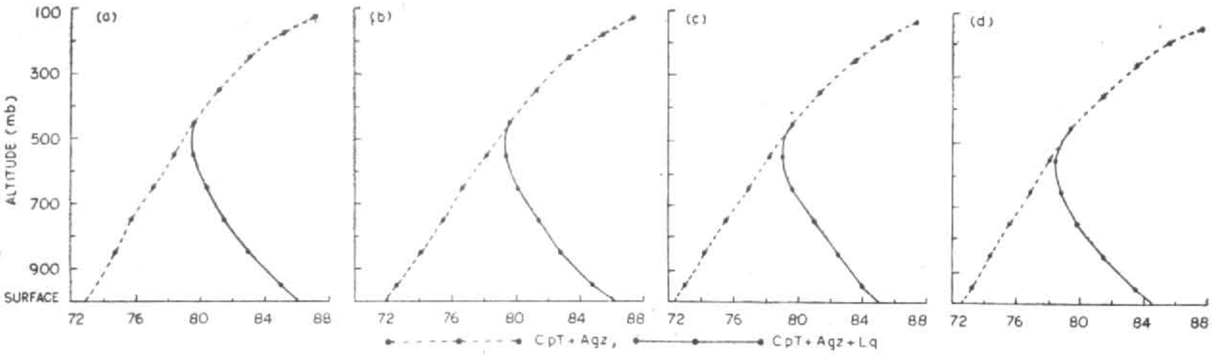
$$= \rho\, \frac{d}{dt} \left(Lq + c_p T \right) - \frac{dp}{dt} \qquad (2)$$

This equation is valid if

(*i*) variations of density gradients with time are small

(*ii*) eddy transfer of energy is negligible.

Mooley (1951) computed density from surface to upper troposphere for July and August months for



Figs. 3 (a-d). Vertical profile of the sensible heat ($c_pT + Agz$) and total effective energy ($c_pT + Agz + Lq$) across 100, 400, 700 and 1000 km respectively radius (Unit: cal/gm)

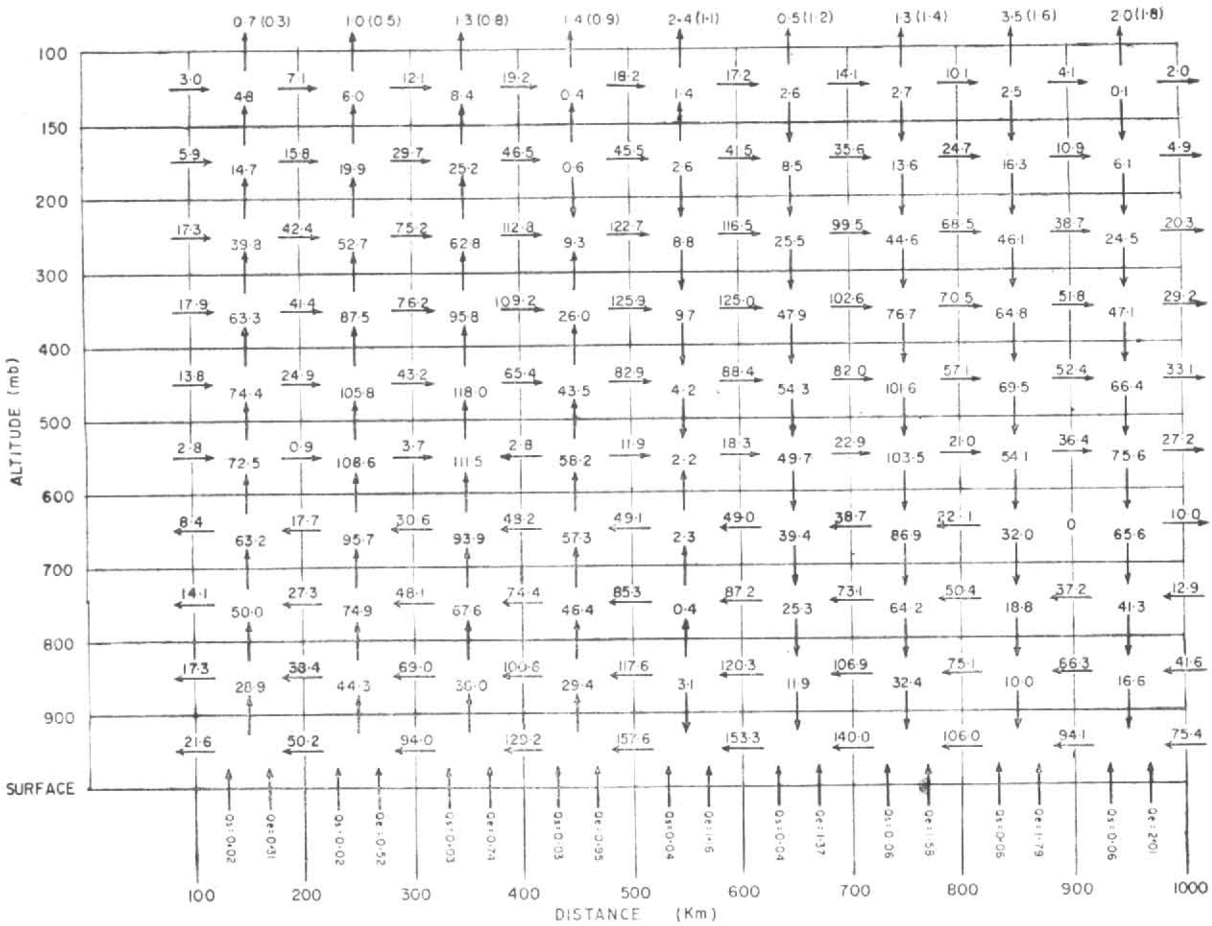


Fig. 4. Budget of total effective energy and the surface vertical fluxes of sensible and latent heat (Unit : 10^{13} cal/sec). Figures on the top in the parenthesis represents black body radiation

selected stations in India. From the table he prepared and seen that variations in the density are not large. The first assumption may, therefore, be accepted as valid. Anjaneyulu (1969) has shown that the eddy flux of energy in the monsoon trough zone is about 1% of the total energy transport. Hence magnitude of the discrepancy is neglecting the eddy motion would not affect the results.

From the hydrostatic assumption, Eqn. (2) can be written as

$$\rho \frac{dE}{dt} = \rho \frac{d}{dt}\left(c_pT + Lq \right) + Ag\rho \frac{dz}{dt}$$

$$= \frac{d}{dt} \rho \left(c_p T + Agz + Lq\right) \qquad (3)$$

From the equation of continuity

$$\rho\ \frac{dE}{dt} = \frac{\partial}{\partial t}\ \rho\ (c_p\ T + Agz + Lq) + \\ + \rho\mathbf{V}.\nabla(c_pT+Agz+Lq) \qquad (4)$$

Neglecting local changes, Eqn. (4) becomes

$$\rho\ \frac{dE}{dt} = \mathbf{V}.\nabla(\rho\ c_pT+Agz+Lq) \qquad (5)$$

This energy equation can be integrated over depression field applying Gauss's divergence equation

$$\int \rho\ \frac{dE}{dt}\ da = \int_l \int_{\text{sfc}}^{100\text{ mb}} (c_pT + Agz + Lq)\ V_r.\frac{1}{g}.dl.dp$$

$$\int_l\int_{\text{sfc}}^{100\text{ mb}} (c_pT + Agz)\ \frac{V_r}{g}\ .dl.dp + \int_l\int_{\text{sfc}}^{100\text{ mb}} Lq.\frac{V_r}{g}.dl.dp \qquad (6)$$

Eqn. (6) may be divided into

$$\int_l\int_{\text{sfc}}^{100\text{ mb}} (c_pT+Agz)\ \frac{V_r}{g}\ dl.dp = Q_s + Q_p - R_a \qquad (7)$$

and

$$\int_l\int_{\text{sfc}}^{100\text{ mb}} Lq\ \frac{V_r}{g}\ dl\ .\ dp\ =\ Q_e + Q_p \qquad (8)$$

4.3. *The heat budget*

The divergence of sensible and latent heat have been determined across different radius circles from Eqns. (7) and (8) and presented in Table 2. Since in a monsoon depression the kinetic energy is a small fraction of the total energy, it has been negleted. For any radius circle large import of sensible heat energy is found to take place in the lower troposphere upto 500 mb. The export of sensible heat in the upper troposphere is slightly more than the import thus leaving a net export of sensible heat. This net surplus increases from 4.49×10^{18} cal/day at 100 km from the centre to a maximum of 34.21×10^{18} cal/day at 500 km. It gradually decreases and attains a value 8.75×10^{18} cal/day at the outer fringe of the depression field. Though the maximum import occurs in the level close to the surface, the export upto 800 km have their largest values between 400 & 200 mb. Beyond this distance the level of maximum export descends to 600 - 500 mb layer.

Ananthakrishnan *et al.* (1965) determined the amount of precipitable water vapour in the atmosphere over India. Computation of the precipitable water has been made from the moisture in the layer from surface to 500 mb for July-August for the area through which depressions moved. The amount of moisture above 500 mb, contributes to about 7% of the total precipitable water in the entire troposphere. In constructing the vertical profiles of the energy, as well as in computing the fluxes in the succeeding paragraphs, the moisture above 500 mb have, therefore, been neglected.

Except for a small outward flow in the 600-500 mb layer there is inflow of latent heat in all the levels, thus giving rise to net surplus of latent heat import in the depression field. The import follows the same pattern as that of the sensible heat with regard to the lateral distribution, and has a maximum value of 36.34×10^{18} cal/day at 500 km and a minimum value of 5.12×10^{18} cal/day, 100 km of the depression centre. The values for the radius of 800 km, are in close agreement with the fluxes obtained by Anjaneyulu (*loc cit*) for the monsoon ellipse which had approximately the same area.

Examination of values at various distances reveal that there is a net inward surplus of total effective energy. The horizontal transport of the total heat in the inflow layer initially increases with decreasing radius and attains peak value around 600 km. Subsequently it decreases as it approaches the depression centre. The mean profiles of the distribution of the sensible and the total effective energy are also worked out for each of the radii. A few cases of these profiles have been depicted in Fig. 3. Sensible heat being sum of enthalpy and the potential energy, it increases with height. On the other hand, the decrease in water vapour content with elevation is reflected in the vertical distribution of the latent heat energy. The net result is that the total effecive energy becomes bimodal in character; a maximum close to the surface and in upper troposphere and a minimum in the 600-500 mb layer.

4.4. *Vertical energy exchange*

The transfer of sensible and latent heat from the ocean/land surface has been computed as follows :

The transfer of sensible heat (Q_s) and latent heat (Q_e) from the surface to the atmosphere is given by :

$$R_e = Q_s + Q_e \qquad (9)$$

Values of R_e have been computed indirectly from the incoming shortwave radiation. For this purpose total global radiation during July and August at Calcutta, Visakhapatnam, Nagpur and Bombay are used. Assuming an albedo of 0.2 (which appears fair enough for these moist months since the surface is normally covered by the green foliage, (*cf,* Budyko 1958), the energy available for the transfer from the ocean/land surface have been determined. Q_e was determined on assuming net surface evaporation rate as 5 mm per day in the area through which depression traverse (Rao *et al.* 1971). Q_s is then determined as a residual of the above equation. The vertical exchange of the total effective energy including the flux of sensible and latent heat from the underlying surface, is shown in Fig. 4. No allowance has been made for the loss of radiation from the top of the atmosphere. Similarly the net effect of radiation cooling and heating due to release of latent heat in the troposphere have been ignored. This is because utilising the radiation data for 15 deg., 20 deg. and 25 deg. N as given by Rao (1963), it is seen that radiation contributes hardly 0.1 to 0.2% to the

TABLE 3

Percentage of latent heat energy to the total effective energy

Layer (mb)	Radial distance (km) 100	200	300	400	500	600	700	800	900	1000
Sfc-900	13.5	13.8	14.3	14.4	14.2	13.3	13.1	12.9	12.8	12.2
900-800	10.0	10.1	10.5	10.5	10.3	10.1	10.1	9.3	8.9	8.7
800-700	6.9	6.9	7.4	7.2	6.8	7.0	6.8	6.2	5.8	5.3
700-600	4.1	4.1	4.2	4.2	4.0	4.1	3.4	3.0	2.9	2.4
600-500	1.5	1.6	1.9	1.5	1.4	1.5	1.0	0.8	0.5	0.4

transfers. Thus the radiative processes are one order of magnitude less important than dynamic process in the thermal balance. As may be seen from Fig. 4, that the net inflow in the lower troposphere is transported upwards into the upper troposphere through vertical convective motion for export up to about 500 km. Between 500 and 600 km, the pattern is rather complex with both upward and downward motion occurring in the vertical columns. However, downward transport dominates the field beyond 600 km. At the top of the diagram net excess for each of the annulus has been indicated by upward arrow. This is the energy available for loss through radiative process.

Using the equation

$$W = \sigma T^4 \tag{10}$$

the energy loss due to radiation are also computed, by assuming black body radiation at a temperature equivalent to that at the top of heavy cirrus cloud deck. For this purpose heights and temperatures of tops of *ci* clouds as given by Deshpande (1965) have been availed. He found that the mean height of cirrus tops ranged between 40,000 & 43,000 feet (about 12 to 13 km) and have a mean temperature of about —56 deg. C. The value thus computed are given at the top of Fig. 4 alongwith the outgoing radiation obtained from the heat budget. A comparison between these two sets revealed a remarkable closeness for a good number of annulii. The heat loss from the top of the depression is slightly large in magnitude than can be accounted for by radiative cooling for a few annular zones, though the order remains unaltered. Fluxes of latent heat from the surface has also been indicated in the bottom row in Fig. 4. In computing these fluxes, the surface evaporation of 5 mm/day was assumed as in the earlier paragraphs. The flux decreased from 2.0 unit in the outermost ring to 0.1 unit near the depression centre.

4.4.1. *Vertical extent of convection*

The heat energy balance can be utilised to determine, theoretically, the vertical extent to which an air parcel can penetrate purely from its surface heat content. Differentiating the first law of thermodynamics

$$dh = c_p T + Agdz$$

with respect to time, it becomes

$$\frac{dh}{dt} = c_p \frac{dT}{dt} + Ag \frac{dz}{dt} \tag{11}$$

Neglecting the heat sink, *i.e.*, radiation and omitting the transfer of heat from the surface, if the heat gained is purely due to condensation of water vapour,

$$dh = -Ldq$$

Thus for a convective element rising unaided of any external heat agency we get

$$-L\frac{dq}{dt} = \frac{d}{dt}(c_p T + Agz) \tag{12}$$

which on integration becomes

$$c_p T + Agz + Lq = \text{constant} \tag{13}$$

It may be seen from Fig. 3 that for the total heat curve the values in the lowest surface level and the outflow layers in the upper troposphere are more or less similar. For the atmospheric heat engine, from the second law of thermodynamics, it follows that heat gained at the surface can penetrate to a maximum height where the heat content equals the maximum value of $c_p T + Lq$ at the surface. In the present case, the maximum value observed at the surface is 85 cal which corresponds to a temperature of 305°A and specific humidity of 20.0 gm/kg. The 85 cal/gm line in the upper troposphere has been shown in Fig. 5 as a thick line and is located at about 200 mb (12 km). This marks the limit upto which the cumuli can reach purely from buoyancy. Above the level of minimum total effective energy, *i.e.*, the top of normal moist layer, the buoyancy process enables cumulus to penetrate the upper troposphere. Penetration of *cb* beyond 40,000 feet (12 km) in monsoon season in a large number of cases is confirmed by observation (Deshpande 1964) and thus is in complete agreement with the value of about 12 km obtained in the present study.

4.5. *Energy transformation and heat source and sink*

The thermal field of the global earth-atmosphere system is such that lower tropospheric levels near the equator acts as a heat source while high levels in the polar regions are sinks. The balance is disturbed briefly when the weather system form and move. In this situation proper appraisal of heat source and sinks in the atmosphere, has to be made. The potential wet bulb temperature in tropical areas, as is well known, decreases with height. Consequently, a warm zone would be formed if the air from the sub-cloud layer alone penetrates in the core region. It has been showed by Riehl (1954) that this warming could lead to a fall of surface pressure to about 1000 mb. But, within the active rainfall area, the middle and upper tropospheric

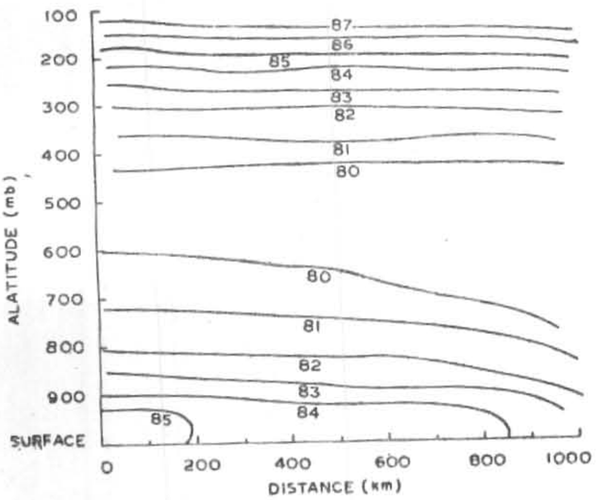


Fig. 5. Vertical cross-section of total effective energy ($c_pT + gz + Lq$) [Unit : cal/gm]

TABLE 4

Frictional dissipation of energy

Radius (km)	Energy loss due to friction (10^{25} ergs day^{-1})	Percentage to latent heat energy loss	Percentage of total effective energy flux
100	1.5	0.07	0.57
200	9.2	0.19	2.22
300	25.3	0.27	4.38
400	47.3	0.35	5.98
500	59.6	0.39	6.69
600	85.5	0.58	8.66
700	77.9	0.61	10.64
800	50.1	0.57	9.01
900	39.5	0.53	3.26
1000	32.8	0.72	2.99

temperatures are much larger than could possibly be accounted by pure moist adiabatic ascents. To explain such observed thermal structure, a local heat source within the circulation field has to be assumed.

For the summer monsoon system as such, from kinematic considerations Das (1962) has concluded that northeast region of India acts as a source while subsidence occurs over the northwestern parts of the sub-continent. From the vertical fluxes shown in Fig. 4, there appears a heat source at 500 km around the depression centre. Besides, within 500 km of the depression centre, the specific humidity increases but the temperature has not been found to vary much. This suggests that not much flux of heat takes place from the surface to the atmosphere. In nearly all radial zones, the value of sensible heat exceeded by about 14 cal/gm in the upper troposphere than at the surface. Assuming heat loss due to radiation as 1 deg. C/day, a fairly reasonable value (Rao 1963), the heat loss becomes 0.24 cal/day. At this rate the upper tropospheric outflow would reach the surface again after nearly 58 days. On this basis the vertical velocity (downward) works out as 245 m/day.

If one has to treat the depression as a closed system we should have a cold source of sufficient magnitude in the depression surroundings so as to neutralise the heat gained in the core at the same time. As the inflow which in the present case is found upto about 600 mb must equal the downward mass transfer,

$$\int_A w\rho \, dA = -2\pi R' \int_{\text{sfc}}^{600\text{ mb}} \frac{v_r}{g} \, dp \qquad (14)$$

where R' is the radius of inflow and A, the area of descent.

As seen from the Fig. 4, 500 km can be assumed as the limit of the area of active ascending motion since over this area the vertical transport of heat is everywhere directed upwards. Averaging the values of V_r within this area, from the above equation the radius of descent comes out as nearly 2000 km, which is the wave length of depression wave. This leads to presume that cold sink of sufficient extent does exist in the depression and that the assumption of closed circulation system may be valid. This is also supported by the moisture computations. From Table 2, it is seen that average precipitation in an area of 500 km radius of the depression comes out as 8 cm. If the descent occurs within an area of radius of 2000 km, the mean evaporation must be 0.5 cm/day. This value is quite realistic and is also supported by work of Rao *et al.* (1971).

It, therefore, appears that the monsoon depression can be regarded as a closed system with source region within 500 km of the centre and sink upto 2000 km. The source and sink observed may not be entirely fortuitous since confirmatory evidence is available from thermal and moisture flux processes.

5. Moisture balance and rainfall

The divergence of moisture can be obtained from the expression:

$$\int_l \int_{\text{sfc}}^{500\text{ mb}} \frac{1}{g} \, q \, V_r \, . \, dl \, . \, dp \qquad (15)$$

As in the case of heat budget, the water vapour balance have been worked out for different radial circles. The horizontal moisture flux, in units of 10^9 gm/sec is depicted in Fig. 6. Maximum flux occurs in the zone of active rainfall activity. Anjaneyulu (1969) computed the water vapour flux in the monsoon ellipse mentioned earlier and obtained net moisture accumulation within the area as 28×10^{15} gm/day. This compares favourably with the figure 20.9×10^{15} gm/day observed in the present case for area covered by 800 km radius.

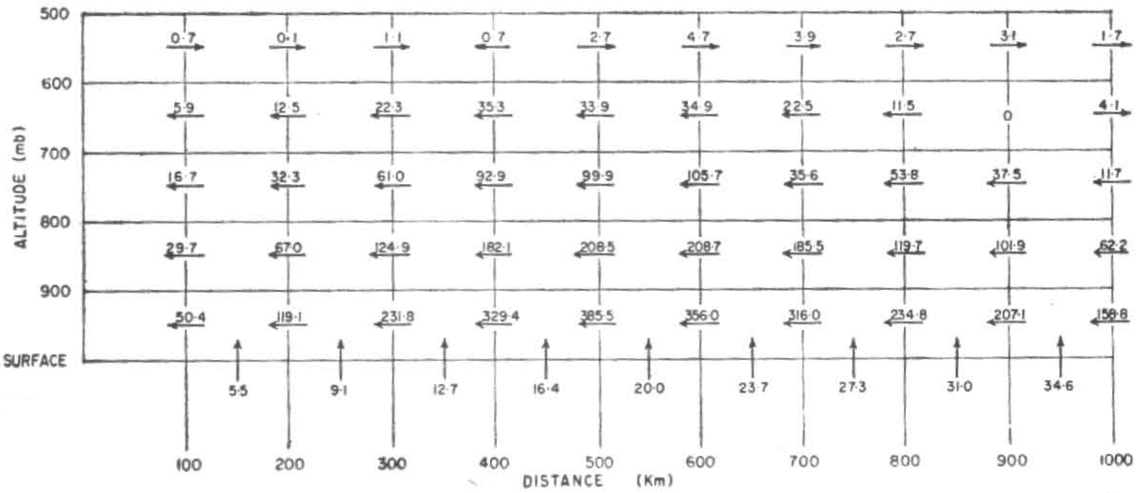


Fig. 6. Divergence of moisture and computed evaporation flux from the surface (Unit : 10^9 gm/sec)

Assuming that all the moisture converging within an area falls out as liquid precipitation, it is possible to estimate the maximum probable rainfall rates corresponding to these convergences. These expected amounts in cm/day for different radii circles are shown below :

Radial distance (km)	100	200	300	400	500
Rainfall (cm)	28.1	15.9	13.4	11.0	8.0
Radial distance (km)	600	700	800	900	1000
Rainfall (cm)	5.4	3.4	1.8	1.2	0.6

It is seen that the rainfall decreases rather gradually as the radius decreases till about 100 km when the increase becomes abruptly large. Comparing again the mean rainfall of the monsoon ellipse obtained by Anjaneyulu (*loc cit*), it is seen that his value of 1.4 cm agrees with the average rainfall of 1.8 cm in 800 km radial circle.

An interesting item is the comparison of the amount of moisture *in situ,* supplied by the underlying surface, with that advected horizontally within the field. Assuming a net surface evaporation rate of 5 mm/day, as is done in earlier sections, the transport of water vapour has been compared below, with the evaporation in different circles.

Ratio of surface moisture supplied to that advected horizontally

Radial distance (km)	100	200	300	400	500
Value of the ratio (expressed as percentage)	1.8	3.1	3.7	4.6	6.3
Radial distance (km)	600	700	800	900	1000
Value of the ratio (expressed as percentage)	9.3	14.7	27.9	42.9	80.2

For the 1000 km radius circle, the surface evaporation is nearly three-fourth of that advected within the area. If the data beyond 1000 km had been considered, the ratio would have been more impressive in the sense that the entire moisture build up would have come from the evaporation. The ratio for 900 and 800 km circles are also quite large being 43 and 28% respectively. The ratio of evaporation to horizontal moisture advection later falls sharply and is 15% for 700 km radius, and 6% for 500 km radius while at 100 km, the surface supplied a mere 2% of that transported with the area. The figures in the inner circles, may not appear to be large but their net effect in moisture inflow and thereby contributing to the enhanced rainfall observed may not be overlooked.

The proportion of the latent heat to that of the total effective heat energy has also been worked out. This has been done for different vertical depth and for various radial circles and given in Table 3. In a layer between surface & 900 mb, about 12 to 14% of the total heat is in the form of latent heat. The latent heat represents about 9 to 10% of the total heat energy in the layer from 900 to 800 mb while between 800 & 700 mb, it is between 5 & 7% latent heat hardly contributes to the total heat above 600 mb.

6. Frictional loss at the boundary layer

Energy loss within a storm field is a rather difficult parameter to evaluate due to lack of data and also because of the number of factors involved. In the present study, dissipation of energy at the boundary layer due to frictional force has been computed as per the relationship given by Taylor (1916). Accordingly the skin friction between the wind and the underlying surface on an unit area can be expressed as:

$$F = K \rho V^2 \tag{16}$$

Multiplying the above expression by V and integrating with respect to area and time, the total loss of energy at the boundary layer can be obtained. For this purpose V is the resultant wind at 0.3 km above mean sea level. In this computation, energy loss due to internal friction has been neglected.

Magnitude of the energy loss for different radial circles has been shown in Table 4. Energy loss due to friction as a percentage of the latent heat energy and the flux of the total effective energy are also given.

The energy loss due to skin friction increases upto about 600 km and then falls. The values of the energy loss are very small and never exceed 1% of that gained due to latent heat. It is only in the outer regions of the depression field are the values more than 0.5%. For a portion of tropical storm within 4 deg. latitude of the centre, Hughes (1952) found that this value amounted to 2% of the energy gained through condensation for that portion. In the present study the values are very small, thus suggesting that the energy loss due to friction is negligible to that gained from precipitation.

In sharp contrast, the loss due to friction forms substantial proportion of the total effective energy in different radii circles of the depression (*vide* column 4 of Table 4). Very close to the centre and at the outer periphery of the depression, while this is less than 5%, between 600 & 800 km, nearly 10% of the total effective energy may be dissipated due to friction.

7. Conclusions

An attempt has been made in this paper to describe some of the salient features of the transports of mass, heat and moisture and their transformation occurring within the field of a mean monsoon depression. The following conclusions are drawn :

(*i*) Inflow and outflow layers are both well defined. The former, near the centre of depression is restricted upto 500 mb, but descends down as the distance increases. Maximum outflow is observed between 400 & 300 mb. Presence of active "ventilation" is also observed at about 500 mb.

(*ii*) Net surplus of latent heat and the total effective heat energy exists at all distances.

(*iii*) Marked upward heat transfer takes place upto 500 km. Beyond 600 km, except for the vertical exchange with the underling surface and at the top of the atmosphere, where some upward flux continues, downward transport dominates the field.

(*iv*) The maximum theoretically possible vertical extent of convection from buoyancy point of view is about 12 km.

(*v*) Analysis of areas of ascent and descent and thermal and moisture aspects suggest depression to be a close system with heat source 500 km around the centre and sink region extending to some 2000 km.

(*vi*) Maximum rainfall, on the assumption that all converging airmass precipitates out, is about 28 cm. Theoretically possible rainfall agrees, in general with that actually observed in association with depression.

(*vii*) Latent heat constitute about 10% of the total heat energy in the lower troposphere.

(*viii*) Large amount of energy is lost between 500 and 600 km, due to friction. This value does not exceed 1% of the heat gained due to condensation.

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