Circulation and Energy Balance in a Tropical Monsoon

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Abstract

Mean monthly meridional and zonal circulations in the region of the Indonesia-Australia summer monsoon and associated transfers of heat energy are investigated using upper air soundings, rainfall and other data from two seasons. Daily sounding data are used to determine also the eddy flux of latent heat. A large increase of meridional circulation strength and energy transfer from November to December on the northern side of the monsoonal trough suggests that establishment of the regime is associated with a comparatively abrupt southward shift and intensification of the Northern Hadley cell in the relevant longitude sector.

When the monsoon has become established latent heat relase in the region exceeds net export of sensible heat from it by a comparatively small amount. These quantities dominate the remaining ones in the heat balance, viz. net radiative cooling in the troposphere and sensible heat transfer from the ocean. Items in the heat budget are compared with those obtaining during the pre-monsoonal regime in the area and on the winter side of the equatorial trough zone at large, the latter according to RIEHL and MALKUS (1959).

Gain of sensible heat is also assessed separately for the lower and upper troposphere in terms of gain or loss of internal energy, the work done by horizontal pressure gradient forces and by vertical pressure gradient forces (gain of potential energy). The vertical upward flux of energy across mid-tropospheric level is deduced and the importance of buoyancy in bringing about this flux is indicated.

1. Introduction

According to recent compilations of the earth-atmosphere radiation budget (cf. J. LONDON, 1957) poleward transport of heat across 10° N latitude in late winter amounts to some 60 per cent of the maximum poleward heat flux at 30° N which is $10.5 \cdot 10^{14}$ cal sec^{-1}. These figures illustrate the important role of the equatorial trough zone (ETZ) as a major source or distributor of heat energy for balancing losses in the remainder, more than three-quarters, of the winter hemisphere.

The outflow of sensible heat from the troposphere overlying the ETZ occurs chiefly in the diverging upper limb of the mean meridional (toroidal) mass circulation associated with the Hadley cells in the two hemispheres. This export of heat occurs as flow of enthalpy and of potential energy the sum of which is nearly equal to 'potential heat' (cf. BALL, 1956) and will be referred to as such for ease of style. Since radiation losses in the troposphere are considerable, the heat budget in the ETZ requires an accordingly large rate of latent heat release in precipitation. It has long been known and demonstrated on detailed observations (RIEHL ET AL., 1950, RIEHL and MALKUS, 1957) that the adjacent tradewind regimes act as formidable accumulator of latent heat. In a later paper (1959) the latter authors assessed latent heat transfer by mean meridional mass flow and eddy flux across the winter boundaries of the ETZ—i.e. at 10 degrees latitude from the trough—to 6.6 in the above units whereas transfer of sensible heat in the lower half of the troposphere amounted to less than half this value.

Net import of latent heat into the ETZ implies within an excess of precipitation over evaporation from the underlying ocean surface. The average rate of upward flux of total heat energy across the 500 mb level within the zone must evidently be very high as it has to account for the accumulations below that level of latent and sensible heat due to lateral import, evaporation and sensible heat flux from the underlying surface. Allowing for a reduction due to net radiative flux divergence in the lower troposphere the required rate of upward flux is about 17 units.

The authors demonstrated that such rate of upward flux cannot be nearly accomplished by mean mass ascent and/or vertical diffusion. They proposed a mechanism by which powerful updrafts in the protected cores of some 1,500 to 3,000 large thunderstorm cells would carry heat energy to great heights at a rate exceeding the required average rate of transfer, the compensating somewhat slower and broader downdrafts enabling dryer air to make contact with the surface energy source. The hypothesis has as yet to be confirmed from special observations and measurements.

There is the parallel problem of large-scale inhomogeneity in longitude, of meridional flow, temperature, pressure height and humidity mixing ratio. This would give rise to standing-eddy flux of energy. The cited authors have shown that ocean—continent differences in heat energy are in themselves rather small, but no attempt has been made to evaluate the differences in meridional circulation strength that occur between longitude sectors. Evidently standing-eddy flux of latent heat cannot be disregarded in the moisture balance on a global scale (STARR and PEIXOTO, 1958).

It would seem appropriate to treat monsoonal circulation regimes—the most prominent among seasonal standing-eddies—as integral parts of the tropical circulation. The summer monsoons especially deserve attention in view of high average rainfall, the year-to-year variations of which have such a strong impact on the national economies of the countries concerned. Thus studies have been made of transfers of heat and zonal momentum associated with the Indian monsoons (RAO, 1959); and of transfer and seasonal balance of zonal angular momentum in the region dominated by the Indonesia-Australia summer monsoon (BERSON and TROUP, 1961). The objective of the present paper is a quantitative assessment of the circulations and associated transfers of heat energy as well as heat balance in the latter region. Changes from pre-monsoon to established monsoon regime and differences with respect to the winter ETZ will also be considered.

Notation

$\mathbf{V}$ horizontal velocity vector
u, v, w west-east, south-north and vertical velocity components
p_0 sea level pressure; p_1 constant pressure close to tropopause level ($\approx$ 100 mb)
S area of horizontal cross section, $d\sigma$ element of S
B boundary of S; dl line element of B
E $= c_pT + Agz + Lq$, total heat energy
E_d $= c_pT + Agz$, 'potential heat' approximately
δR_a vertically integrated net radiative flux divergence in the troposphere
R_0 net radiation on ocean surface
H, G upward flux of sensible heat from ocean surface and in the ocean respectively
ε evaporation; P precipitation
$\overline{A}$ monthly mean value of A; A' departure of daily from monthly mean
$\overline{A}^*$ departure of $\overline{A}$ from pressure mean,

$$\frac{1}{\delta p}\int^{\delta p} \overline{A}\, dp$$

Other symbols above and in the following have their usual meaning unless explained.

2. The heat balance equation

The change of heat energy in unit time averaged over a month, in a volume determined by S and $p_0 - p_1$ assuming hydrostatic equilibrium, is given by

$$\frac{1}{g}\iint_{p_1}^{S\,p_0} \frac{\partial \overline{E}}{\partial t}\, dp\, d\sigma =$$

$$= -\frac{1}{g}\int\limits^{S}\left[\frac{1}{p_0 - p_1}\frac{\partial p_0}{\partial t}\int\limits_{p_1}^{p_0}\overline{E}\,dp\right]d\sigma - \qquad (1)$$
$$B$$

$$-\frac{1}{g}\iint\limits_{p_1}^{S\,p_0}\mathrm{div}_2\,\overline{\mathbf{V}E}\,dp\,d\sigma + \int\limits^{S}(H + L\varepsilon - \delta R_a)\,d\sigma$$
$$C \qquad\qquad\qquad D$$

where the transfer of heat energy from ocean to atmosphere $H + L\varepsilon$ must satisfy the energy balance equation at the surface

$$R_0 = H - G + L\,\varepsilon \qquad (2)$$

Equation (1) in which bars have been omitted for convenience over all variables except E and V, is correct under the assumptions that

(i) Eddy transfer of energy across the isobaric surface p_1 is negligibly small in comparison with transfers at the ocean surface.

(ii) Horizontal density gradients in the troposphere are sufficiently small to omit a term containing $\overline{\mathbf{V}\,\nabla_2\,\varrho}$ which otherwise would occur in eq. (1).

Owing to the absence of any but weak baroclinity in the tropical atmosphere overlying an ocean region assumption (ii) is very much justified. Assumption (i) cannot be a priori justified but it should not lead to discrepancies of a magnitude affecting the results and their interpretation.

The term B represents a change of energy in the volume by a change of mass contained in it. It is proportional to the area-integrated change of sea-level pressure since no mass can be transferred across the constant pressure surface p_1.

The term C is the negative horizontal flux divergence in the troposphere and represents therefore lateral import to, or export of heat energy out of the volume according to whether it is positive or negative.

In the expression for flux divergence, the contribution by mean meridional motions may be separated into two parts, viz.

$$\int\limits^{\delta p}\overline{v}\,\overline{E}\,dp = \frac{1}{\delta p}\int\limits^{\delta p}\overline{v}\,dp\int\overline{E}\,dp + \int\limits^{\delta p}\overline{v}^{\star}\,\overline{E}^{\star}\,dp \equiv A_0\,(\overline{v}) + A_t\,(\overline{v}) \qquad (3)$$

and similarly for zonal transport. Here A_0 is the transport due to a net mean mass flux and A_t is a toroidal transport (PRIESTLEY, 1949), which is determined by the covariances of $\overline{E}$ and $\overline{v}$ or $\overline{E}$ and $\overline{u}$ respectively. The asterisk on either energy or velocity component may be omitted in the integrand. Introducing eq. (3) into (1) it can be shown that the term B cancels out with the term containing $A_0\,(v)$ and $A_0\,(u)$. Eq. (1) thus simplifies to

$$\frac{1}{g}\iint\limits^{S\,\delta p}\frac{\partial E}{\partial t}\,dp\,d\sigma =$$
$$= -\frac{1}{g}\iint\limits^{S\,\delta p}\mathrm{div}_2\left(\overline{\mathbf{V}}^{\star}\,\overline{E} + \overline{\mathbf{V}'\,E'}\right)dp\,d\sigma + \int(H + L\varepsilon - \delta R_a)\,d\sigma \qquad (4)$$

where $\overline{\mathbf{V}'\,E'}$ is the residual synoptic-scale eddy term. Its evaluation strictly requires to deduce daily means from several soundings each day. The transfer term would then not include mezo-scale eddy flux such as is produced by a seabreeze circulation (see section 4).

By Gauss' theorem the first integral on the right of eq. (4) is equal to the outflow of energy through the vertical 'wall' bounding the volume. The quantities $\overline{V_n^{\star}}\,\overline{E}$, $\overline{V_n'\,E'}$ in which V_n is the outward normal wind component, can be determined from zonal and meridional transports and the orientation of the sides of the selected quadrangle. The implications of interpolation between sounding stations will be discussed in section 4.

3. Energy and wind profiles

The toroidal fluxes of total and potential heat are uniquely determined by the vertical profiles of $\overline{u}$, $\overline{v}$ and $\overline{E}$, or $\overline{E}_d$ respectively. Typical profiles are shown in Figure 1 for Singapore and Darwin. Very similar profiles of $\overline{E}$ and $\overline{E}_d$ result for longitude-averaged seasonal means at 15° N (PALMÉN, RIEHL and VUORELA, 1958) and on the winter boundaries of the ETZ (RIEHL and MALKUS, 1959). Heat energy decreases upwards above a shallow neutral layer and increases above the 700 mb level, first moderately, then more sharply in the uppermost troposphere. (The sharpest increase just below the 100 mb level is a

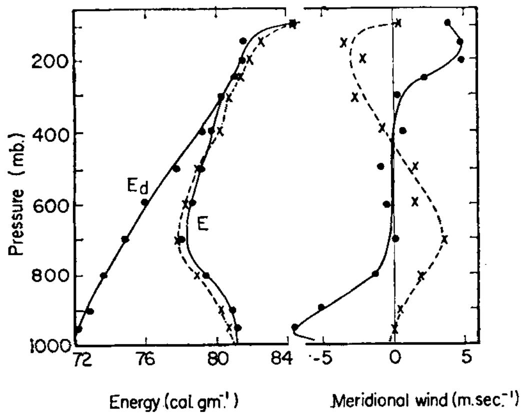


Figure 1. Distribution of total heat energy (E), potential heat (E_d) and meridional wind component over Singapore and Darwin, January 1957.

further complication which may be disregarded here.) The increase in the upper troposphere of E_d reflects traverses of markedly subadiabatically stratified layers by the sounding balloons in a great majority of all soundings. This apparently persistent stable layer is present in the tropics over both oceans and land (cf. cited authors; RAO, 1960). It is seen that total heat energy at 150 mb exceeds the value at the surface where the actual source is located. However, energy must be transported upwards to great heights because the excess amount in the equatorial zone is carried away, poleward as potential heat, effectively by meridional mass flow in the uppermost troposphere as will be seen from the profile of meridional velocity. The consequences of the presence, in the average profile, of a stable layer in the upper troposphere for an effective mechanism of energy transfer to great heights has been mentioned in the Introduction. The compilation of energy balance in the monsoonal regime (section 8) will give an opportunity to assess the position in such a regime.

A glance at the curve for $\overline{E}_d$ (at Singapore) shows that apart from the mentioned anomaly just below the tropopause this can be taken as a straight line, to a first approximation. Toroidal transport of potential heat can therefore be determined as a function of vertical wind profile. In the analysis of local (station) transfers of potential heat it is of advantage to determine the regression

$$A_t = A_{to} + b\,S \qquad (5)$$

where S is the strength of the circulation, $S_v = \overline{v}_{\max} - \overline{v}_{\min}$ being defined as meridional circulation strength and $S_u = \overline{u}_{\max} - \overline{u}_{\min}$ as *relative* zonal circulation strength. The coefficient b in the regression equation depends according to the foregoing consideration on the mean slope of the E_d-curve in a diagram such as Fig. 1. In eq. (5) the subscripts max., min. denote the *primary* maximum and minimum in the profiles of the wind components ($S_v > 0$ in the Northern, $S_v < 0$ in the Southern hemisphere Hadley-type circulation). In the subtropics S_v is presumably closely related to the meridional circulation strength in the manner defined by PRIESTLEY and TROUP (1954). S_u is in the nature of a vertical shear of zonal wind. Regressions according to eq. (5) will be evaluated in section 5.

4. The region and data considered

Selection of an oceanic area dominated by the circulation associated with the Indonesia-Australia northwest monsoon, yet at the same time suitable for the calculation of energy flux divergence, is dictated by the location of the sounding stations. During parts of December and as a rule during the whole of January and February, the regime of the westerly monsoon extends from Malaya across the Timor, Arafura and Banda Seas to the coastal belt of northern Australia, and westward in the Indian Ocean to about 90° E longitude (Fig. 2).

At or near the northern fringe is Singapore, near the southern fringe Cocos Island on the edge of the Southeast tradewinds, Darwin, and Port Hedland on the Australian coast, the latter station in the orbit of the circulation around the shallow heat low over the western parts of the continent. Detailed calculations of energy transports and flux divergence were made from the data of these four stations. Meridional transports and circulation parameters were also evaluated for Koror in the Pacific Ocean to the east of Mindanao (Philippines), in about the longitude of Darwin; and for six additional sounding stations in the Australian region north of 33° S (Fig. 2 a).

The periods used for all these stations, except Koror, are November to February

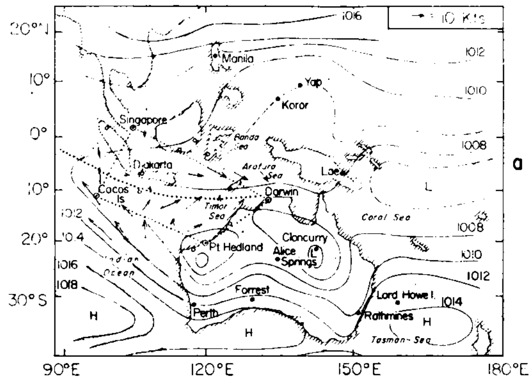
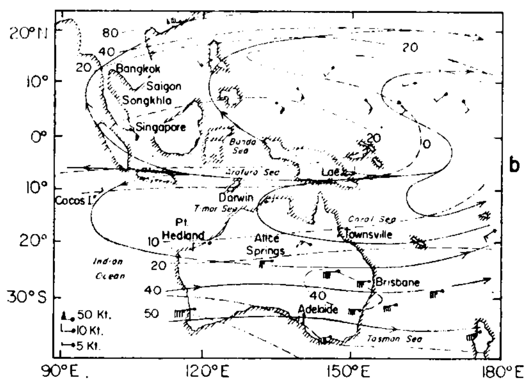


Figure 2. (a) Normal M.S.L. pressure field for January. Position of intertropical convergence in monsoon regime indicated by heavy dotted line. Arrows are vector means of surface wind in five degree latitude-longitude squares. Stations used: ● rawinsonde, ○ rainfall.



(b) Streamlines and isotachs of average 200 mb flow in February (After TROUP, 1961 a).

1956—57 and 1957—58, the latter season during IGY. The data used in the cross section showing meridional circulation during the December to February season (Fig. 3) vary from station to station, but none cover less than four seasons.

The area selected for the assessment of heat balance is indicated on Fig. 2. In this region with the exception of the extreme northern portion, November is a pre-monsoonal month. During December marked often abrupt changes occur in the circulation and in rainfall. During January the monsoonal regime is usually established in the region but is nevertheless irregular and fluctuating (HANNAY, 1945; TROUP 1961 a).

Considerations of moisture balance require an evaluation of average precipitation in the area. This is based on fifteen stations within or close to the boundaries of the area, as indicated on Fig. 2 a. Radiation data are taken from various sources in the literature but mainly from LONDON (1957). However, for an estimation or radiative flux divergence in the troposphere according to a formula by MÖLLER (1960) cloud amount and dewpoint depression at 300 mb are required. For the former recent climatological atlases have been consulted, the latter has been evaluated from the basic sounding data.

As to the representativeness of monthly means evaluated from soundings carried out once a day mention was made in section 2 of the desirability to exclude seabreeze effects. In the event coastal stations figure prominently, but the bias for seabreeze components in the lowest 3,000 ft is not thought to be serious except during the first season (noon soundings) at Port Hedland and possibly Darwin. In the final assessment of heat balance during February (Table 4) both seasons have been averaged so that a systematic error would be reduced to negligible size.

Such error could arise also from linear interpolation in connection with the evaluation of energy flux divergence from transports at the four mentioned stations. However, as seen in Fig. 3, the meridional component in the upper troposphere changes across the relevant region from positive values in the north to negative

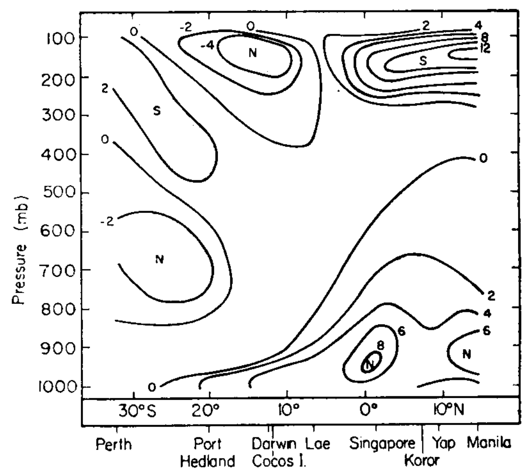


Figure 3. Meridional (toroidal) wind component in vertical section from Australian Heat Low across monsoonal trough, December to February average. Speed in knots.

in the south, presumably in a quasi-linear manner. Similarly, in the lowest layer—separated from the upper limb of the circulation by an 'inactive layer' (cf. PALMÉN, RIEHL and VUORELA, 1958)—the northerly component decreases southwards from the equator to about 20° S.

The *zonal* flow is strongest in the upper troposphere. Here the easterly 'jet stream' at about 50,000 ft which generally in the region is located somewhat north of the equator, shifts to a few degrees south of it with the start of the monsoon season. Average streamlines and isotachs at the 40,000 ft level for February according to TROUP (1961 a) are shown in Fig. 2 b. The maximum easterly component at this level remains well over the northernmost portion of the area in question. The presence of stronger westerlies close to the ITC between, say, Darwin and Singapore (Fig. 2) may however involve errors in the layer surface to 700 mb.

For completion of monthly series IGY cards were used. In both seasons, but especially in the pre-IGY season, soundings were not made on single days or during sequences of days. The computation of eddy fluxes requires in such cases careful interpolation. In the event interpolation was obviated where possibly by treating slightly shorter periods. Processing of daily wind components was facilitated by the use of CSIRAC Electronic computer.

5. Transports

(*a*) *Meridional energy transports*

Toroidal and eddy transports of total heat energy at stations between 10° N and 35° S are set out in Table 1. To compare these with the rate of energy flow across latitude circles as required by radiation unbalance on a global scale (last column), February has here been selected; for heat storage in the atmosphere and ocean may then be regarded as negligibly small. Values north of the equator were deduced from LONDON's (1957) radiation budget, values south of the equator by judicious interpolation between these and the distribution calculated poleward of 45° S by GABITES (1960).

Table 1 should be taken as giving merely a broad survey of meridional heat transfer in the general wider region marked by continental cooling in the north and continental heating in the south. The northward heat flux at stations near the northern edge of the monsoonal trough is consistently stronger than required by global radiation unbalance. At Cocos Island and Darwin, near the southern end of the trough, fluxes are southward of consider-

Table 1. Meridional heat transport during February, northward counted as positive. Unit 10^5 cal cm^{-1} sec^{-1}. Toroidal transport refers to total heat energy, eddy flux to latent heat only (except in footnoted value).

Station	Lat. Long. (Degrees)		Toroidal		Eddy		Total Av. 57—58	Radiation Unbalance
			1957	1958	1957	1958		
Tropics								
Koror	7.4 N	134.5 E		4.9				1.4
Singapore	1.4	100.9	1.0	2.7	0.4	0.2	2.1	1.0
Cocos Island	12.1 S	103.9	—1.5	—1.4	—0.4	—0.6	—1.9	—0.1
Darwin	12.4	130.8	—1.4	—1.6	—0.2	—0.8	—2.0	
Port Hedland	20.3	118.6	—0.6	—1.1	0.4	—0.1	—0.7	—0.4
Cloncurry	20.6	128.1	—4.1	—4.1				
Subtropics								
Perth	31.9	115.9	1.9	3.7				
Forrest	30.9	128.1	—2.0	0.3				
Lord Howe Isl.	31.5	159.1	—0.4	—1.1			—2.2[1]	—0.8
Rathmines	33.1	151.7	—0.7	—1.7				

[1] Eddy flux of enthalpy and moisture for Auckland (N.Z.) for period Jan.—Feb., 1950—52, unpublished data (Priestley & Troup, 1954).

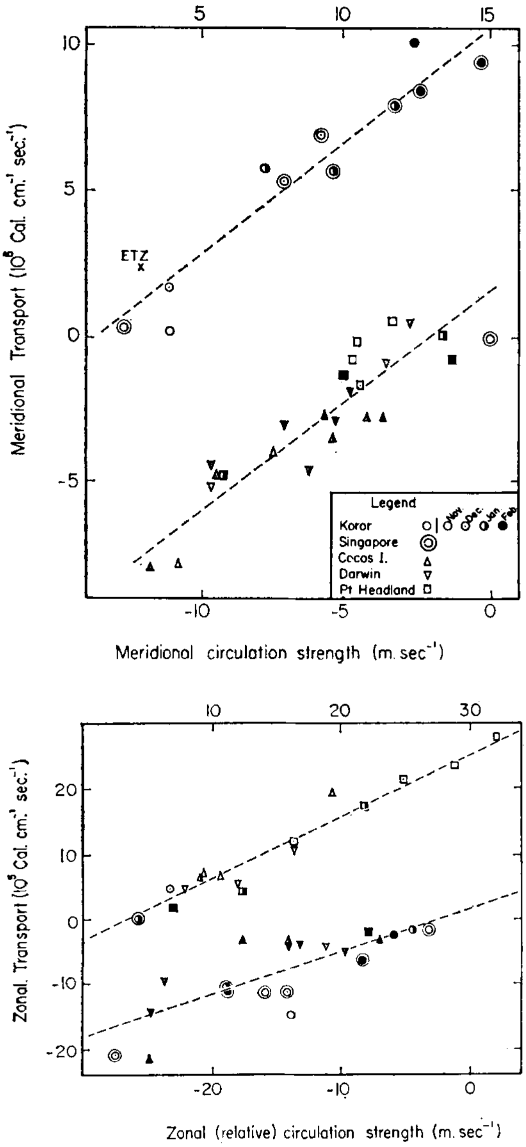


Figure 4. Toroidal flux of potential heat as function of circulation strength (a) Meridional, (b) Zonal. Sloping lines indicate regressions according to eqs. 5 a and 5 b.

able magnitude whereas the global radiation budget at this time of year does not require meridional transfer of appreciable magnitude across the relevant latitude circles. Further south at Port Hedland, poleward transport differs but little from the radiation value. There is thus a suggestion of increased meridional heat flux divergence in the monsoonal longitude sector as compared with the belt.

Cloncurry in Queensland, Rathmines near Newcastle and Lord Howe Island in the Tasman Sea show poleward transports in excess of that required by average radiation values. If this were interpreted as an additional effect of continental heating, it would be consistent with the figures for Perth on the west coast which indicate a marked equatorward energy transport.

(b) *Circulation strength and potential heat transport*

In Fig. 4 a and b potential heat transports are shown as function of the corresponding circulation strengths. As the meridional circulation reverses from Northern to Southern Hadley sense across the axis of the monsoonal trough (cf. Fig. 3), the net transports reverse from northward to southward. The regression equation (5) becomes

$$A_t(\bar{v}, \bar{E}_d) = \begin{matrix} -1.0 + 0.77\, S_v & S_v > 0 \\ 1.6 + 0.76\, S_v & S_v < 0 \end{matrix} \qquad (5a)$$

where the constants and S_v are in the units of Fig. 4. Correlation coefficients $r\,(A_t, S_v)$ have the values 0.95 and 0.91 in the positive and negative range respectively.

The ETZ position in Fig. 4 a refers to the values of S_v and A_t derived from the profile of meridional flow and from potential heat flux on the winter side at a distance ten degrees latitude from the position of the trough during the December—February and June—August seasons. The flux value is therefore the average of northward flux coupled with N-Hadley circulation at 5° N and southward flux coupled with S-Hadley circulation at 2° N. The Singapore (1.4° N) and Koror (7.4° N) plots for November, i.e. before the establishment of winter monsoon, show agreement in order of magnitude with average hemispheric-scale toroidal circulation strength and heat flux. However, from November to December there takes place at Singapore a sharp increase of circulation and flux which by February have four times the magnitude of the ETZ winter values.

The stations south of 10° S show predominantly S-Hadley circulation of moderate strength in agreement with the long-term average during the December—February period

(Fig. 3). However, in November and December rather small negative values obtain at Darwin and Port Hedland. These reflect low-level northerly components associated with sea-breezes, an effect of sampling bias which was discussed in section 4. This effect too is responsible for an inconsistency of low-level components as between Figures 2 a and 3.

The regression equation for zonal transport is:

$$A_t\,(u, E_d) = \begin{matrix} -3.2 + 0.94\,S_u & S_u > 0 \\ 1.6 + 0.66\,S_u & S_u < 0 \end{matrix} \qquad (5\text{b})$$

with coefficients 0.96 and 0.77. Since the high-level maximum easterlies are located just south of the equator, both northern and southern station plots are found in the positive and negative range of S_u. Absence of noticeable trend from November through February at Singapore reflects the influence of the semi-annual wave in the equatorial upper easterlies (cf. CLARKSON, 1955; TROUP, 1961 a). Noteworthy is the unexpectedly small scatter about the regression line in the range of moderate and large positive S_u occurring mainly in November and December well south of the equator, where deep easterlies decrease with height or shallow easterlies are overlain by westerlies.

(c) Sensible and latent heat transport

It is informative to plot meridional transport of potential heat (henceforth more suitably referred to as sensible heat) against latent heat transport in a scatter diagram (Fig. 5). Throughout the period November to February at all stations except Port Hedland, latent heat is transported into the monsoonal trough from north and south while sensible heat flux diverges, the export of the latter exceeding invariably import of the former kind of energy. At Singapore meridional energy exchange increases abruptly from pre-monsoon to monsoon regime, while the values for Koror indicate a similar but more gradual change. The data for Cocos Island and Darwin indicate reverse changes although they are less abrupt and lack consistency. At Port Hedland little change is apparent in the meridional energy exchange which remains low throughout.

But for the abruptness and magnitude noted at Singapore, these changes would be consistent with a gradual seasonal southward shift of the Hadley circulations in either hemisphere.

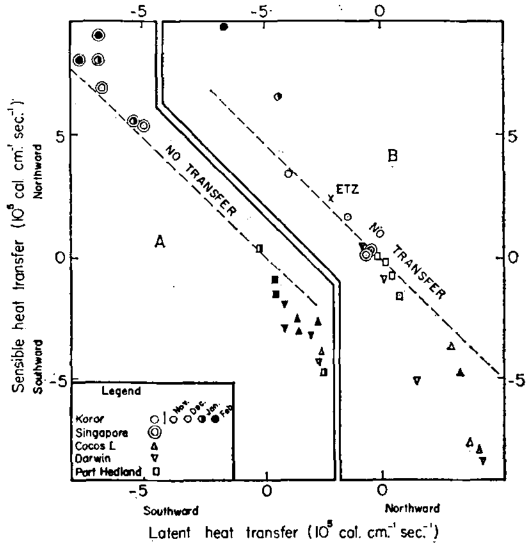


Figure 5. Meridional (toroidal) transport of sensible heat versus transport of latent heat. Sloping lines indicate zero net transport by meridional mass circulation. A. Monsoonal regime. B. Pre-monsoonal or transitional (Koror in winter Trades).

A similar analysis of zonal transports is not feasible on account of the limited sampling. It has been eastablished in recent studies (PALMER, 1958; BERSON and TROUP, 1961) that frictional control on time averaged large-

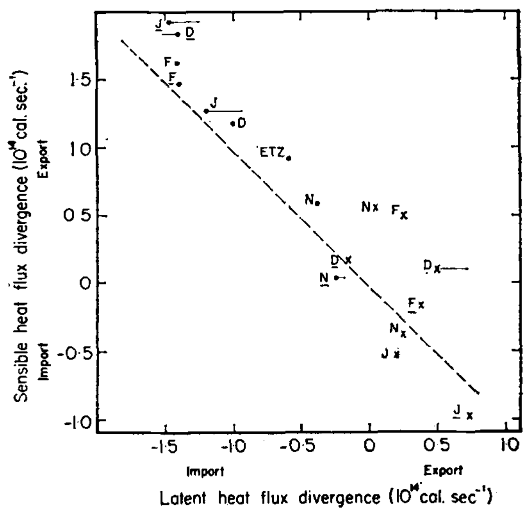


Figure 6. Components of integrated energy flux divergence in monsoonal regime. ● meridional, × zonal flux divergence. Letters denote months, plain 1956—57 season, underlined 1957—58 season. (For horizontal lines see text.)

scale zonal flow—i.e. above a shallow planetary friction layer—is very weak even in the tropics to within a few degrees latitude from the equator. 'Uncontrolled' shifts in longitude of zonal patterns are indeed not uncommon. A large and widespread anomaly in high-tropospheric zonal flow in the tropics occurred e.g. in January 1958 and has been analysed in some detail by TROUP (1961 b). It is reflected in Fig. 6 as an exceptionally large magnitude of zonal divergence of both latent and sensible heat flux over the portion of the monsoonal trough that now will be considered in some more detail.

6. Flux divergence of sensible and latent heat

The representation in Fig. 6 is similar to that in Fig. 5. Export or import of *total* heat by either components of flux divergence can be inferred from the appropriate displacement of the plots from the sloping line. Consistent export by *meridional* flux divergence is readily verified. On the average over the two sampling seasons the excess of sensible heat export over latent heat import has the same order of magnitude as the equivalent net export from the winter end of the ETZ[1]. This agreement proves, if nothing else, the adequacy of the station network for the method of calculating flux divergence.

Noteworthy is the small overlap, in this representation, between meridional and zonal flux divergence: latent heat import into the monsoonal trough coupled with sensible heat export out of it for meridional divergence and predominantly vice versa for zonal divergence. With the establishment of monsoon and an accompanying marked increase in the meridional transfer of sensible and latent heat energies, the overlap vanishes altogether. Furthermore, during the monsoon period, average zonal flux divergence is considerably less in magnitude than meridional flux divergence. Since toroidal transports do not represent residual terms between large quantities (see Fig. 1) and therefore are subject to errors no larger than compound product errors of $\bar{v}$ into $\bar{T}$, or $\bar{v}$ into $\bar{q}$, etc., one must accept as real this lack of compensation between meridional and zonal flux divergence with respect to the two forms of energy.

However, net divergence of *total* heat flux emerges as a relatively small residual term between meridional and zonal flux divergences, similarly as is generally the case with mass flux. Although this makes it difficult to assess accurately the heat budget, it will be of little consequence for determining the broad organisation of lateral and vertical energy exchange.

7. Eddy flux

According to calculations by PALMÉN et al. (1958) poleward eddy flux of latent heat across 15° N in winter is $2 \cdot 10^{14}$ cal sec^{-1}, i.e. by an order of magnitude smaller than the transfer of total energy by mean meridional circulations across that latitude. Riehl and Malkus computed the moisture eddy flux as a residual term in the ETZ heat balance. They found that it is directed outward (poleward) on the winter side of the ETZ and represents slightly more than the net export of total heat by mean meridional circulations.

The monsoonal regime of Indonesia and Australia is marked by not infrequent disturbances such as tropical cyclones, 'rain depressions' and surge-like movements of the ITC. A direct calculation of the moisture eddy flux seemed therefore desirable. Some results are set out in Table 2. The vector fluxes during monsoon period indicate moisture export from the region. The *resultant* fluxes at Singapore, Darwin and Port Hedland are consistent with horizontal moisture gradients in accordance with the large-scale distribution of land and sea, but the southward flux at Cocos Island appears unduly large.

As seen from the last row in Table 2 there is an increase in the magnitude of eddy transports from November to December which is mainly a transitional period. Furthermore, during monsoon regime eddy fluxes are considerably stronger on the southern end of the monsoonal trough than on its northern side. This would reflect the positions and intensities of the synoptic-scale disturbances.

The effect of including moisture eddy flux in the calculation of flux divergence is shown in Fig. 6 where horizontal lines indicate ap-

[1] Rate of outflow from a sector of the relevant 10° latitude belt equal in size to the trial calculation area (Fig. 2), so that effectively heating rates per unit area are compared.

Table 2. Eddy-flux of latent heat during 1956/57 and 1957/58 seasons, northward and eastward counted positive. Unit 10^4 cal cm^{-1} sec^{-1}. ('Monsoon' refers here to January and February averages.)

Station	Meridional			Zonal			Vector Monsoon	
	Nov.	Dec.	Monsoon	Nov.	Dec.	Monsoon		
Singapore......	−1.1	−1.5	2.0	−1.1	−3.7	−2.2	3	310°
Cocos Island....	−2.5	−5.8	−9.0	2.3	−1.1	3.7	10	160°
Darwin.........	−0.2	−19.2	−3.7	1.6	8.6	7.9	9	120°
Port Hedland...	−2.5	−3.1	−3.8	−2.1	1.3	−4.5	6	230°
Average of Modulus......	1.6	7.4	4.6	1.8	3.7	4.6	—	—

propriate additions or subtractions in excess of $0.03 \cdot 10^{14}$ cal sec^{-1}. The effect is seemingly of some consequence in December and January and is to decrease meridional flux convergence (import) or to increase zonal flux divergence (export) as calculated from mass circulations alone.

Eddy flux of enthalpy has been found to be completely negligible at 15° N during winter according to calculations by PALMÉN et al. (1958). For stations near the equator RAO (1960) showed that it is not always negligible compared with moisture eddy flux. The magnitude of the eddy flux of enthalpy and its divergence have not been investigated here; however to ignore this eddy flux and its divergence is consistent with assumption (ii) in section 2.

8. Sensible heat balance

(a) General considerations

Evaluation of the heat storage term in eq. (4) showed it to be of smaller order of magnitude than integrated flux divergence. In view of this and the minor role of eddies in the transfer of heat energy, the case may be treated as a quasi-stationary heat-driven mass circulation. As such it is sustained by surface-sensible heat source and by release of heat of condensation (precipitation heat, LP). The energy used in the work done by vertical pressure gradient forces against gravity as the air ascends in the monsoonal trough zone ("gain of potential energy"), is partly off-set by loss of internal energy due to expansion and net radiative cooling. It is, however, not a priori clear in which manner transfers and balance of energy are arranged as between the lower troposphere, i.e. the monsoon layer proper, and the overlying part of the troposphere; and in which layer mainly work is done by horizontal pressure gradient forces at the boundaries by the surrounding atmosphere on the system (the integral of $V_n\, p\, \varrho^{-1} \equiv R\, V_n T$).

The work by eddy stresses at the lateral boundaries and by viscous stresses at the surface—the latter tacitly omitted in eq. (4)—represent additional losses of energy. The former effect is thought to be very small. As to the latter a check calculation applying the average value of surface zonal stress in the region (cf. BERSON and TROUP, 1961) and surface zonal wind components (Fig. 2) showed that the frictional stress term is less than 1 per cent of horizontal energy flux divergence in the troposphere. In the following both these works term will be omitted.

In order to transform eq. (4) into a balance equation for sensible heat only, we eliminate evaporation heat and latent heat flux convergence with the aid of the moisture balance equation for the troposphere

$$L \int^{S} (P - \varepsilon)\, d\sigma = -\frac{L}{g} \oint \iint_{p_1}^{p_0} \overline{V_n\, q}\, dp\, dl \quad (7)$$

Omitting eddy flux of enthalpy we thus obtain

$$L \int^{S} P\, d\sigma - \frac{1}{g} \oint \int_{p_1}^{p_0} \overline{V_n^{\star} E_d}\, dp\, dl = \int^{S} Q_s\, d\sigma \quad (8)$$

where $-Q_s = H + \delta R_a$ is the sensible heat source for the troposphere (actually a sink), i.e. exclusive of conversion of latent into sensible heat.

(b) Formulation for two layers

The 'monsoon layer' (I) may be represented by the layer $p_0 - 550$ mb, the 550 mb level being but little higher than the equatorward ends of the moist tradewind layer. As evident from Fig. 1 the moisture content is not quite negligible above 550 mb, even on the average over a month. In disregarding release of latent heat in the upper layer (II) we shall not so much introduce an inaccuracy as rather consider the upper boundary of the monsoonal layer as not strictly specified. Release of latent heat of fusion in the layer II is negligible and the process of fusion is of no consequence for the balance except if it is systematically bound up with melting in the layer I. This is not typical for the tropics (cf. RIEHL and MALKUS, 1959). The balance equations for sensible heat for the layers I and II may be written in the form

$$\int^{S} (LP - Q_s)\, d\sigma = \frac{1}{g} \oint \int_{p'}^{p_0} \overline{V}_n^{\star}\, \overline{E}_d\, dp\, dl + \int^{S} (\overline{\varrho w\, E})_{p'}\, d\sigma \qquad (9\text{ a})$$

and

$$- \int^{S} \sigma R_{aII}\, d\sigma = \frac{1}{g} \oint \int_{p_1}^{p'} \overline{V}_n^{\star}\, \overline{E}_d\, dp\, dl - \int^{S} (\overline{\varrho w\, E})_{p'}\, d\sigma \qquad (9\text{ b})$$

where p' is the pressure at the top of the monsoon layer.

Consistently with the definition of p', in the expression of the vertical (upward) transfer of energy E may be approximated by $c_v\, T + A\, g\, z$.

The average upward transport of potential energy $A\, g\, z$ at the level p' represents a portion of the total work by vertical pressure gradient forces in the monsoon layer, since

$$g \int_0^{z'} \overline{\varrho w}\, dz = g\, (\overline{\varrho\, w\, z})_{z'} - g \int_0^{z'} \overline{\bar{z}\, \varrho\, \frac{\partial w}{\partial z}}\, dz\,, \quad (10)$$

by parts. On account of the continuity equation the integral on the right, after integrating once more over the area S, can be and commonly is, interpreted as outflow of potential energy through the vertical surface bounding S. Integrating from surface to z_1, since $\overline{(\varrho w)}_1 = 0$ by assumption, the first term on the right of eq. (10) vanishes and the work by vertical pressure gradient forces is equal to the outflow of potential energy.

With respect to each of the layers, I and II, the first term on the right of eq. (10) is finite and must occur with opposite sign in equations (9 a) and (9 b). This is true also for the upward transport of internal energy through the level p'. In Table 4 the first term on the right in eq. (10) figures as "vertical flux divergence of potential energy", the second term as "horizontal flux divergence of potential energy".

(c) *Analysis*

Results of the computations on equations (7), (8) and (9) are summarized in Tables 3 and 4.

As will be seen from Table 3 monthly rainfall amounts vary considerably from one month to another, even within the monsoon period. The variation of precipitation heat is however broadly similar to the variation of net sensible heat export from the monsoonal trough zone. With the exception of December 1957, the difference between them is positive,

Table 3. Tropospheric Heat budget. Gains and Losses in units 10^{-2} ly min^{-1}. For symbols see text.

Period	Area Rainf. mm	Prec. Heat	Sens. Heat Export	Establishment of Monsoon (cf. Troup 1961 a, b)
Dec. 56	150	20.7	17.1	Monsoon onset early in month
Feb. 57	235	34.7	28.7	Fully established regime
Dec. 57	200	27.2	27.5	Onset in N. Australia late December
Jan. 58	145	19.4	12.8	Anomalous high-level zonal winds
Feb. 58	195	28.6	23.3	Fully established regime
Average Monsoon		26.0	21.8	
		Q_s 4.2		δR_a 12.0 H 7.8
Winter ETZ		21.6	13.2	
		8.4		13.5 5.1
Pre-Monsoon (Nov. 57)		8.3	—4.2	
		12.5		

Table 4. Two-layer sensible heat budget for Indonesia-Australia summer monsoon (February). Unit 10^{-2} ly $\min^{-1}$.

	Flux Divergence			Heat Source	
	Horizontal	Vertical	Total		
LAYER I					
Internal Energy....	3.4	25.5	28.9	$-\delta R_a$	−6.4
Hor. pressure Forces....	1.4		1.4	H	5.2
Potential Energy....	−16.9	17.4	0.5	LP	31.6
	−12.1	42.9	30.8		30.4
LAYER II					
Internal Energy....	−44.0	−25.5	−69.5	δR_a	−7.1
Hor. pressure Forces....	−17.8	.	−17.8		
Potential Energy....	99.1	−17.4	81.7	LP	0
	37.1	−42.9	− 5.6		−7.1
TROPOSPHERE					
Net loss and gain......			25.2		23.3

and its average value is slightly more than 15 per cent of precipitation heat. The position appears to be significantly different from winter ETZ as there sensible heat export amounts to only 60 per cent of precipitation heat. The data in Table 3 for a pre-monsoonal regime indicate that with markedly less rainfall in the region horizontal heat transfer can even reverse, since in this particular month (November 1957) import obtains. This is not confirmed by November 1956 for which available rainfall data were however indadequate for an assessment of P.

If as suggested in Table 3 during the monsoon regime by and large, Q_s is positive and small, if not negligible, in magnitude compared with precipitation heat, then according to eq. (8) net radiative cooling would somewhat exceed surface heat transfer. According to London's tables the summer value of δR_a in the belt 0—20° N is 13.5 units. It is based on average cloudiness and vertical moisture distribution in the belt. Applying Möller's formula (see section 4) to the monsoonal data, a value of 12 obtains. With this estimate H can be evaluated from eq. (8) and turns out to be 8 units.

Evaporation heat has here been determined for individual months from P and resultant moisture flux convergence according to eq. (7), the average value being 14 units and differing little from Riehl & Malkus' evaluation by means of an aerodynamic bulk formula. Accepting H and $L\varepsilon$ as representative, the Bowen ratio would be 0.55 and so considerably higher than the ETZ value (0.4) obtained from net radiation and evaporation according to eq. (2) assuming $G = 0$.[1] If for consistency London's value for R_0 (19 units) is taken here too, then from eq. (2) one finds $G = -3$. In view of the uncertainty of the precise magnitude of the other terms in eq. (2) no significance can be attached to this result except that it indicates a correct order of magnitude of ocean heat flux G (MALKUS, 1960). Estimates of R_0, ε and H from various other sources such as ALBRECHT'S (1940) measurements in the Java Sea and Budyko's Atlas of Heat Balance would not help to clarify the position; for a ten per cent error in P would change H by a large fractional amount.

Detailed heat budget for monsoon and overlying layer.—The above considerations give some assurance that the following analysis will be in order, with tolerance of the above discrepancies. Table 4 summarizes the two-layer budgets for February selected as only suitable sampling period on the basis of Table 3. The right-hand side of Table 4 serves again as a check on the values for net flux divergence, within the limits of the stated tolerance. H has here been obtained from the calculated evaporation amounts with the assumption that the Bowen ratio is 0.4. The discrepancy between the left and right sides of the Table is 2 units, i.e. for the whole troposphere.

As might have been anticipated the lower layer gains sensible heat at a high rate while the upper layer loses at a much lower rate, the net gain being 25 units. To understand how these gains and losses are brought about, one must

[1] For oceans in general a range of Bowen ratio 0.1 to 0.2 is held to be valid but the higher value in the waters of the equatorial trough zone is supposed to be consistent with the downdraft hypothesis mentioned in Introduction.

consider more closely the work by vertical pressure gradient forces (gain of potential energy). As will be seen in layer I the relevant term is negligibly small. This would indicate that the effect of buoyant forces is there large even when averaged over the whole area under consideration; for we may write

$$\int \varrho w \, d\sigma = \frac{1}{S} \int w \, d\sigma \cdot \int \varrho \, d\sigma + \int \varrho^{\star\star} w \, d\sigma \tag{11}$$

where $\varrho^{\star\star}$ is the departure of density from the area mean value. Where $\varrho^{\star\star}$ is negative buoyancy is positive and so the second term on the right must be large negative in the monsoon layer to reduce the term on the left side to a small residual. The discriminating effect of buoyancy in effecting vertical heat transfer introduced by PRIESTLEY and SWINBANK (1947) has been proposed also by Riehl and Malkus in the mentioned updraft hypothesis. As corollary the downdraft mechanism is qualitatively verified by the consideration of large negative buoyancy forces required to explain the very high gain of potential energy in the upper layer (II), viz. 82 units.

A relatively large negative value of the *horizontal* pressure gradient work term in the layer II implies that work is done by the environment on the system. This is consistent with a meridional equatorward drift (amounting to about 1 m sec^{-1}) the presence of which was found necessary to maintain zonal angular momentum balance in the monsoon-dominated portion of the ETZ in the eastern hemisphere (BERSON and TROUP, 1961); for the horizontal pressure gradient south of the high-level maximum equatorial easterlies is directed almost due northward and is much stronger than gradients in the lower troposphere.

Comparison with non-monsoonal regime.—A glance at Table 5 will show that in regard to the equatorial regimes organization of transfers and sensible heat gain is very much alike: the lower branch of the circulation is self-maintaining and in addition provides more than the energy necessary to sustain the circulation in the overlying troposphere where the net loss of sensible heat is incurred. The remaining excess energy is transferred in the upper limb to regimes outside the equatorial trough zone, i.e. in the case of the monsoon regime, to north- and southward with rather indefinite net zonal transfer (cf. Fig. 6). Excess energy is there made available to be consumed in reducing radiative cooling or by release of potential energy during descent.

Table 5. Sensible heat gain in equatorial and tropical regimes. Unit 10^{-1} ly min^{-1}. For last row see text.

	Southern Monsoon		Winter ETZ		Northern Trades
Layer (mb)	1,000 — 550	550 — 100	1,000 — 500	500 — 125	1,000 — 700
Internal Energy	2.9	— 7.0	1.5	— 3.0	0.3
Hor.pressure Forces	0.1	— 1.8	0.0	— 0.6	0.1
Potential Energy	0.1	8.2	0.4	3.0	— 0.2
	3.1	— 0.6	1.9	— 0.6	0.2
Upward flow of Energy	4.3		2.3		
$\overline{w}^\star$	10.6		6.5 km/day		

As to the magnitude of the net heat gain, in the monsoonal regime this is almost twice as large as in the winter ETZ. The corresponding factor in turnover of energy is however in excess of two.

The last two rows serve to indicate the magnitude of the buoyancy effect discussed earlier. For this can also be gauged by the enormous value for the rate average vertical mass flow by which the energy transfer across mid-tropospheric level could be accomplished ($\overline{w}^\star = \int \overline{\varrho\, w\, E}\, d\sigma / \int \bar{\varrho}\, \bar{E}\, d\sigma$).

The comparison with heat gain in the tradewind regime is illuminating. According to the analysis by RIEHL and MALKUS (1957) the North Pacific tradewind layer during spring gains sensible heat at the rate 0.2 in the units of Table 5, or only about 5 and 10 per cent of the gain in the lower troposphere of the equatorial regimes. In the Trades potential energy is lost as the air descends through the top of the tradewind inversion. As mentioned earlier, the role of the tradewind system in the general circulation is that it acts as an accumulator of latent heat which is made available to the equatorial trough zone, but the rate of supply to the monsoon regime is greatly stepped up on account of the enhanced mass circulation.

Conclusion

It has been shown that the mean mass circulations associated with the monsoon are the determining factor in the horizontal transfer of heat energy. The effect of synoptic-scale eddies is to counteract the mean-flux convergence of latent heat there being evidence that this is mainly due to ocean-land moisture gradients. The effect is much smaller than on the winter side of the equatorial trough zone at large where it represents a substantial item in the heat budget, according to the study by Riehl and Malkus. This difference may not be a real one since in the latter case eddy transfer of moisture has been deduced as a residual term in the heat balance.

The check on quasi-balance between the rates of net sensible heat export and latent heat conversion by an estimate of radiative sources and sinks was of necessity a crude one but served nevertheless a useful purpose. Thus there is some assurance on the magnitude of vertical heat transfer at mid-tropospheric level and the necessity of invoking buoyancy effects for its mechanism. The problem of vertical heat exchange in an equatorial regime has here been formulated in a somewhat different way than earlier by those authors and shown up on mainly independent data. It remains a problem to be solved both theoretically and by means of observational techniques as yet not elaborated.

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