ON THE MERIDIONAL CIRCULATION AND RELEASE OF KINETIC ENERGY IN THE TROPICS

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ABSTRACT

The mean meridional circulation is computed for the northern hemisphere winter. The equatorward current in the low troposphere has a strength of 1–3 mps near 10 to 15 deg lat, and the return current, centered near 200 mb, has about the same speed. An inactive layer without much net poleward or equatorward component extends from 700 to 400 mb.

Next, the release of kinetic energy by the meridional circulation is calculated, and it is shown that this release equals the kinetic energy export from the tropics to the middle latitudes. Finally, the mean meridional circulation exports heat from the tropics poleward; at 15 deg lat this heat transport has the right order of magnitude to balance the net heat loss in higher latitudes.

1. Meridional circulation

The meridional and vertical flow of absolute angular momentum and the associated mean meridional circulation in the northern hemisphere trade wind belt have been treated in recent years by several authors (Bjerknes 1951, Mintz 1955, Mintz and Lang 1955, Palmén 1955, Pisharoty 1955, Van Mieghem 1956). Several aspects of the trade wind circulation could be connected satisfactorily, such as surface and free atmosphere stresses, and the meridional and vertical flow of absolute angular momentum associated with the mean meridional circulation and with the large-scale eddies. In Palmén's study the meridional circulation was determined directly from wind observations. The computations however were based on daily wind observations at only 19 stations around 13 deg N lat for January and February 1952; winds above 14,000 ft were not considered. It is therefore necessary to widen the investigation to comprise more wind data over a longer period and to include the upper troposphere.

In the present study the reports from 47 stations

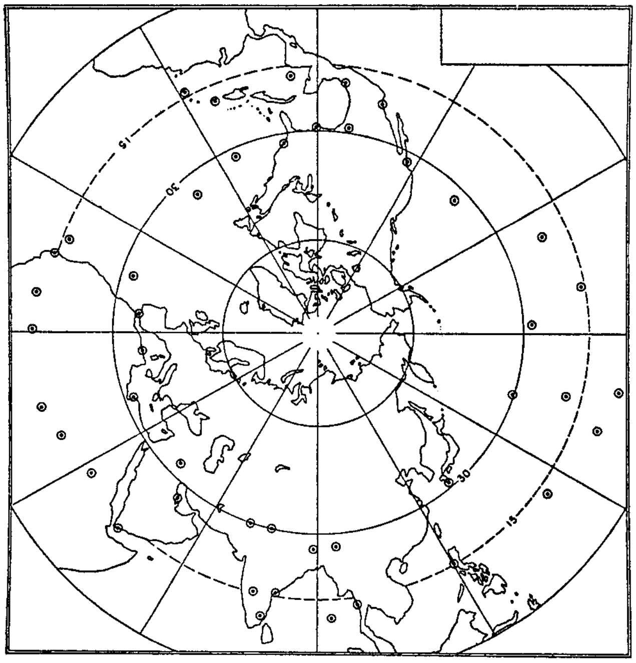

FIG. 1. Stations used for computation of mean meridional circulation.

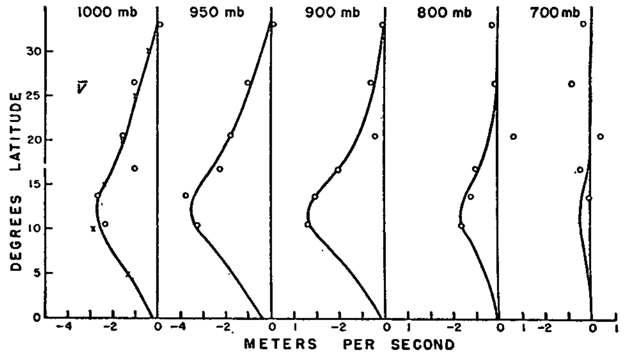

FIG. 2. Mean meridional circulation (m per sec) for different isobaric surfaces as a function of latitude.

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between 8 deg N and 36 deg N have been analyzed (fig. 1). Mean meridional and zonal wind components were calculated for December 1950, January, February and December 1951, January and February 1952. Some of these months were missing at a few stations. The mean components were determined at 1000 mb (surface), 950, 900, 800 and 700 mb. Above 700 mb the data became too sparse for satisfactory extension to higher levels. This deficiency was remedied later with observations from the winter of 1955–56.

Results from the earlier data will be discussed first. The stations were grouped in six belts, with mean latitudes at 10, 14, 17, 20, 26 and 33 deg N. Fig. 2 shows the mean values of the meridional component in each group for the different isobaric surfaces. Free-hand curves have been drawn to depict the most probable latitudinal distribution of the meridional component. At 1000 mb, surface values computed by Riehl and Yeh (1950) from ship observations have been entered with crosses. In drawing the curves least weight has been given to the groups centered at 17 and 20 deg, since there were fewer observations in these than in the other groups.

The mean southward component practically disappears at 700 mb, so that the total meridional mass exchange can be computed from fig. 2. From continuity the upper return current must transport the same mass as the lower equatorward flow. The distribution of the upper current along the vertical—a very important factor for computations of heat flux and kinetic energy production—could not be ascertained due to lack of sufficient high-level wind data. Some general deductions about the vertical profile of the meridional component however could be made from the equation of motion. This was done before the observations for the winter of 1955–56 became available.

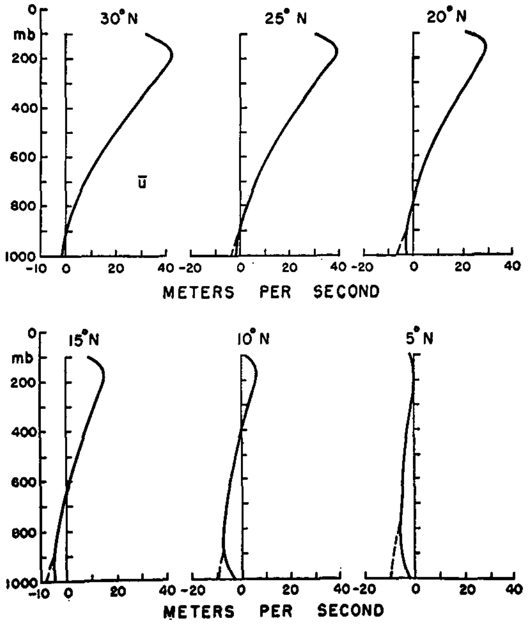

FIG. 3. Mean zonal wind component (m per sec) for different isobaric surfaces as a function of latitude. Dashed lines denote extrapolation of geostrophic wind to surface.

For a steady meridional circulation the mean[2] meridional wind component $\bar{v}$ is determined from the formula (Palmén, 1955)

$$\bar{v} = \frac{\bar{w}\dfrac{\partial \bar{u}}{\partial z} + \dfrac{1}{\cos^2 \varphi}\dfrac{\partial}{\partial y}\,(\overline{u'v'}\cos^2 \varphi) - \dfrac{1}{\rho}\dfrac{\partial \bar{\tau}_{xz}}{\partial z}}{2\Omega \sin \varphi + \dfrac{\bar{u}}{a}\tan \varphi - \dfrac{\partial \bar{u}}{\partial y}}. \qquad (1)$$

Here u, v, w denote the zonal, meridional and vertical wind components along the axes x, y and z, φ the latitude, ρ density, a the radius of the earth, Ω the angular velocity of the earth's rotation, and $\bar{\tau}_{xz}$ the mean component of the zonal shearing stress. The density is considered constant at each level.

The denominator of (1) expresses the dynamic stability of the mean motion; it can be calculated if the mean zonal wind distribution is known. For determination of the mean zonal flow, inhomogeneities of the data are not as critical as for the determination of the mean meridional flow. Hence this component was computed up to 100 mb for the six latitude belts, using also the values given by Mintz (1954) and Jenkinson (1954). The result, shown in fig. 3, is a compromise solution of the three sets of mean values.

Table 1 gives values for the different terms of the denominator at 500 mb and 200 mb using fig. 3; fig. 4 shows the result graphically. It is seen that the restraint on mean meridional displacements due to the mean dynamic stability is much less at 200 mb than at 500 mb. The mean dynamic stability decreases from the ground to the level of maximum zonal wind (200–150 mb); this favors mean meridional motions in the high troposphere.

The numerator of (1) depends on three terms. These represent the influence of the mean vertical velocity $\bar{w}$, the influence of the large-scale meridional eddy flux of westerly momentum, and the influence of the mean vertical shearing stress depending upon eddies of different scales. Close to the equator $\bar{w}$ is upward and $\partial \bar{u}/\partial z$ small; in the subtropics $\bar{w}$ is nega-

TABLE 1. Mean values of the terms in the denominator of (1) at 500 and 200 mb (10^{-5} sec^{-1}).

Latitude	200 mb				500 mb			
	$2\Omega \sin \varphi$	$\dfrac{\bar{u}}{a}\tan \varphi$	$-\dfrac{\partial \bar{u}}{\partial y}$	Σ	$2\Omega \sin \varphi$	$\dfrac{\bar{u}}{a}\tan \varphi$	$-\dfrac{\partial \bar{u}}{\partial y}$	Σ
30°N	7.29	0.38	0.25	7.92	7.29	0.18	0.15	7.62
25	6.16	0.28	−1.60	4.84	6.16	0.12	−1.10	5.18
20	4.99	0.16	−2.25	2.90	4.99	0.06	−1.25	3.80
15	3.77	0.06	−2.05	1.78	3.77	0.01	−1.10	2.68
10	2.53	0.02	−1.45	1.10	2.53	0.00	−0.70	1.83
5	1.27	0.00	−0.85	0.42	1.27	0.00	−0.30	0.97

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[2] In this paper bars denote averages over longitude and time, primes deviations from the average.

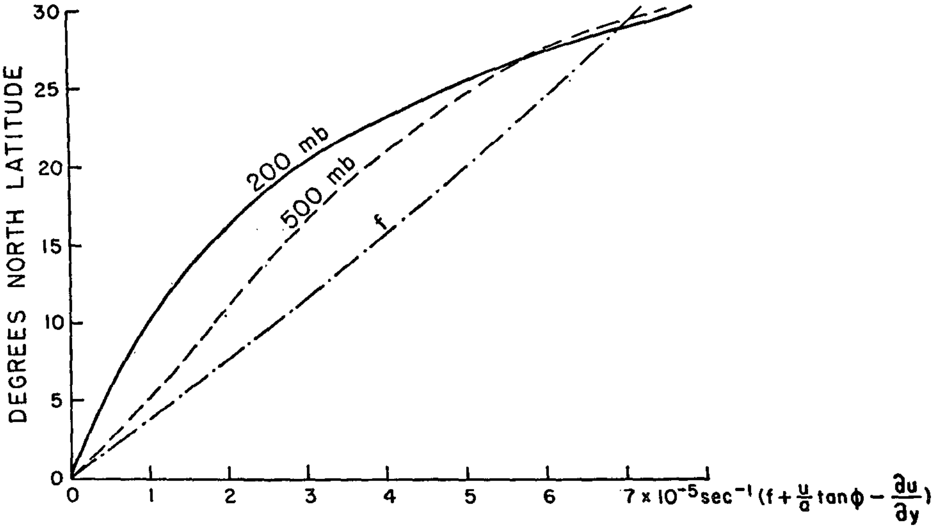

FIG. 4. Value of denominator in equation (1) at 500 mb and 200 mb as a function of latitude. Dash-dotted: the Coriolis parameter.

tive and $\partial\bar{u}/\partial z$ large. At intermediate latitudes $\bar{w}$ vanishes. Thus the first term in the numerator can be neglected between the equator and about 15 deg N; farther north it makes a negative contribution to $\bar{v}$ below the level of maximum wind, and vanishes near 200 mb. A maximal value of the term can be estimated as follows. Assume $\bar{w} = -0.3$ cm per sec near 20 deg N and $\partial\bar{u}/\partial z = 3 \times 10^{-3}$ sec^{-1}, both large values, then $\bar{v} = -18$ cm per sec if the Coriolis parameter at 20 deg lat is used for the denominator.

The second term in the numerator is positive over the whole trade wind cell according to Mintz (1955) and reaches a pronounced maximum at the level of strongest zonal wind. The meridional motion resulting from the first two terms in the numerator has been computed by Mintz and Lang (1955). The last term in the numerator expresses the divergence of the vertical eddy flux of momentum. Introducing a coefficient of eddy viscosity (μ), we can write

$$\frac{\partial}{\partial z}\bar{\tau}_{xz} = \frac{\partial}{\partial z}\left(\mu\frac{\partial\bar{u}}{\partial z}\right).$$

If we assume that μ is independent of height, the term depends only on the curvature of the mean zonal wind profile along the vertical. It therefore nearly vanishes in the middle troposphere where the zonal wind increases upward at a nearly constant rate. It reaches a pronounced maximum near the level of maximum west wind. There the curvature of the zonal wind profile is negative, and since the term enters in (1) with a negative sign, it acts in the same sense as the momentum divergence term. It follows from all these considerations that $\bar{v}$ should not be distributed in a semilinear manner through the troposphere above 700 mb, but that the return flow should be strongly concentrated in the high troposphere.

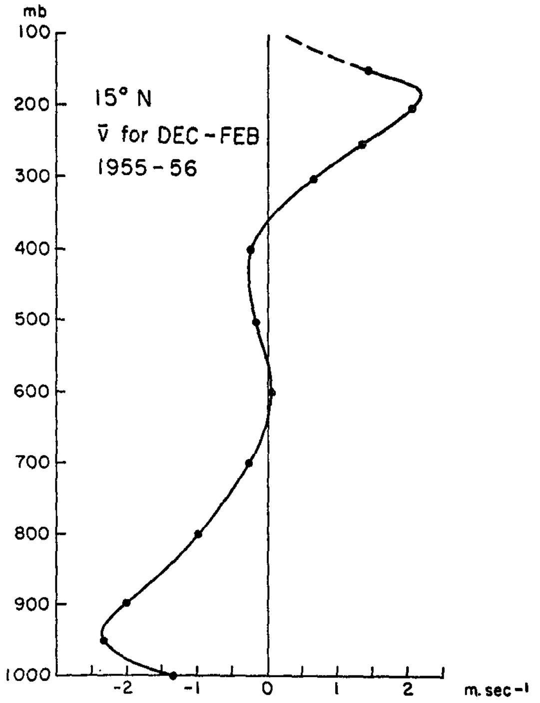

FIG. 5. Mean meridional wind component (m per sec) for 15 deg N lat, December 1955–February 1956, as a function of pressure. Data from 13 rawin stations near 15 deg N lat (*cf.* fig. 6).

In order to check the validity of this reasoning, the vertical distribution of $\bar{v}$ was computed for 13 stations around latitude 15 deg N for December 1955–February 1956. In these months the number of rawin stations was far greater than previously. Fig. 5 shows the result, which is in good agreement with the preceding reasoning. The mean meridional flow approximately vanishes between 700 mb and 350 mb; it reaches pronounced maxima in the high troposphere and near the ground. Exact mass balance however is not obtained, which must be attributed to the fact that large parts of the 15th parallel are still devoid of observations. Fig. 6 shows the distribution of the time average of the meridional flow around the latitude circle. In the lower troposphere the flow is directed equatorward almost everywhere, while a marked oscillation of north and south components occurs at high levels indicating a three-wave structure. Troughs are situated in the eastern Atlantic, eastern Africa and the Persian Gulf, and presumably in the central Pacific. A broad region with southerly winds must exist over the meteorologically uncharted area of the eastern Pacific, analyzed with dashed lines in fig. 6. Cross-sections drawn for other winter months have revealed a similar picture.

Using the result of the earlier reasoning we can now construct profiles of the probable vertical distribution of $\bar{v}$ at different latitudes. The result is presented in fig. 7 where the values of fig. 3 have been used up to 700 mb and where it has been assumed that the mass transport of the return current equals the southward transport below 700 mb. Table 2 gives the strength of this mass transport, and that of the vertical mass transport through the 700-mb surface required for balance. South of 20 deg lat the values are slightly larger than those previously computed by Palmén (1955); to the north they are somewhat smaller. The maximum flow again occurs near 12 deg N; according to the new data it amounts to 260 × 10^6 tons per sec, compared with the previous estimate of 230 × 10^6 tons per sec.

TABLE 2. Mean meridional mass transport (10^6 tons per sec) across different latitudes for December-February, and corresponding vertical mass transport across the 700-mb surface.

Latitude	30°N	25°	20°	15°	10°	5°	0°
Meridional circulation	18	56	121	228	245	154	35
Vertical transport		−38	−65	−107	−17	91	119

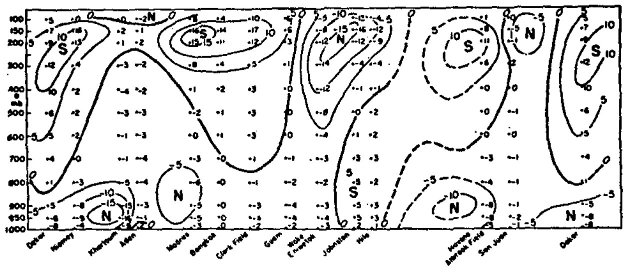

FIG. 6. Regional distribution of meridional wind component (kn) around 15 deg N lat, December 1955–February 1956. Dashed lines indicate interpolation in eastern Pacific.

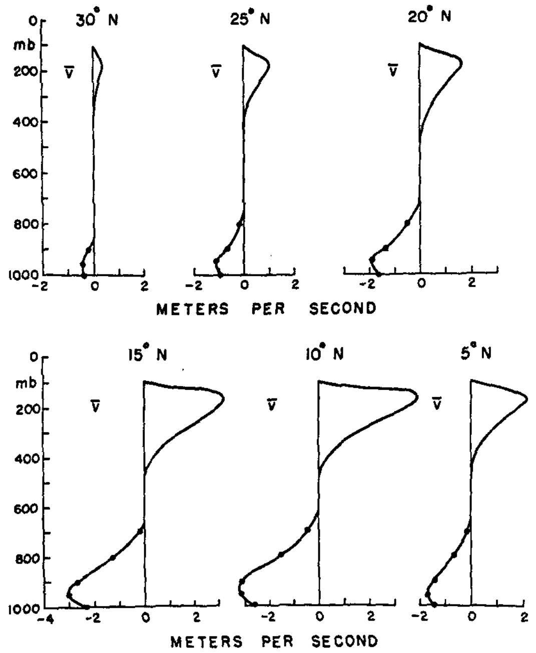

FIG. 7. Vertical profiles of the mean meridional circulation (m per sec) at different latitudes for December 1950, January, February and December 1951, January and February 1952, with upper portion interpolated according to the shape of the profile in fig. 5.

2. Release of kinetic energy

The work done by the meridional horizontal pressure gradient force in the mean meridional circulation per unit length of the meridian is given by

$$\delta W = -\,2\pi a \cos\varphi \int_{z_1}^{z_2} \bar{v}\,\frac{\partial \bar{p}}{\partial y}\,dz, \tag{2}$$

where p is the pressure and where the integration has been extended from the ground (z_1) to the top (z_2) of the circulation cell. Introducing the hydrostatic equation and the geostrophic wind relation,

$$\delta W = \frac{2\pi a f \cos\varphi}{g} \int_{p_2}^{p_1} \bar{v}\bar{u}_g dp, \tag{3}$$

where g is the acceleration of gravity, f the Coriolis parameter and $\bar{u}_g$ the mean zonal geostrophic wind at a given isobaric surface.

Equation (3) can be evaluated from figs. 2 and 3 assuming that the difference between $\bar{u}$ and $\bar{u}_g$ is sufficiently small to be neglected. The product $\bar{v}\bar{u}_g$ is graphically presented in fig. 8; Table 3 gives values of the work term for the layers 1000–700 mb and 700–100 mb, and for the whole troposphere. The production of kinetic energy reaches a maximum between 15 deg and 20 deg N, about 5 deg N of the latitude of strongest meridional mass transport. It is questionable whether the value at latitude 30 deg N is realistic.

TABLE 3. Production of kinetic energy by the mean meridional circulation per unit length of a meridian (10^4 kj m^{-1} sec^{-1}).

Latitude......	0°	5°	10°	15°	20°	25°	30°
1000–700 mb	0	1.5	5.2	4.3	2.1	0.4	0.0
700–100	0	0.0	3.4	11.0	13.4	12.0	5.5
1000–100	0	1.5	8.6	15.3	15.5	12.4	5.5

The mean meridional mass movement at that latitude is very weak, but on account of the strong meridional pressure gradient in the high troposphere an appreciable production of kinetic energy is computed.

The values of kinetic energy production for the return current and for the whole troposphere have been plotted in fig. 9. From these curves the kinetic energy production was computed in bands with width of 5 deg lat (table 4).

TABLE 4. Production of kinetic energy by the mean meridional circulation in different latitude belts (10^{10} kj per sec).

Isobaric layer	1000–700	700–100	1000–100
0–5°N	0.3	0.0	0.3
5–10°	1.8	0.8	2.6
10–15°	2.7	4.0	6.7
15–20°	1.7	7.1	8.8
20–25°	0.6	7.2	7.8
25–30°	0.1	5.0	5.1
0–30°N	7.2	24.1	31.3

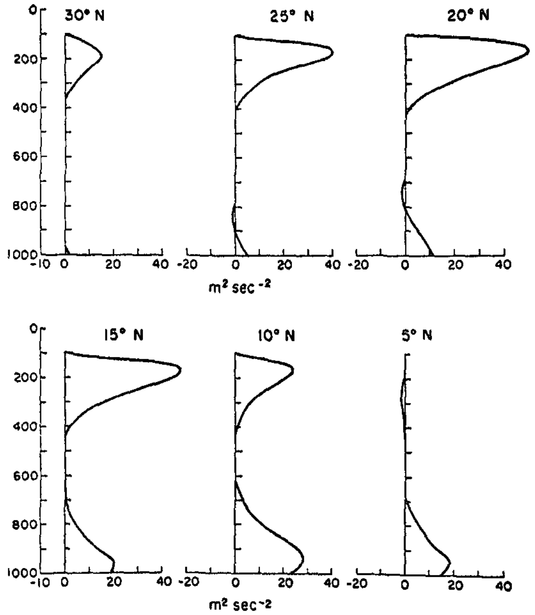

FIG. 8. The product $\bar{u}_g\bar{v}$ (m² per sec²) as a function of pressure for different latitudes. Computed from figs. 3 and 7.

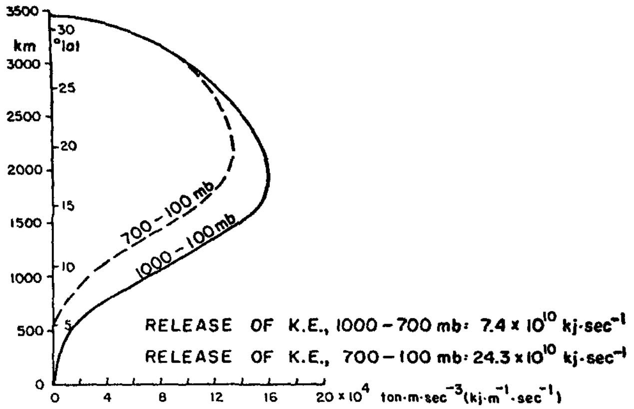

FIG. 9. Production of kinetic energy per unit length of a meridian by the mean meridional circulation for the layers 700–100 mb and 1000–100 mb. Calculated from fig. 8.

A part of the kinetic energy produced is exported across 30 deg lat; a part is converted into heat by friction. The frictional dissipation cannot be measured but an attempt will be made to estimate its magnitude.

The change of kinetic energy due to frictional dissipation (δF) in the meridional circulation cell per unit length of the meridian between the altitudes z_1 and z_2 is given by

$$\delta F = 2\pi a \cos \varphi \int_{z_1}^{z_2} \left(\bar{u} \frac{\partial \bar{\tau}_{xz}}{\partial z} + \bar{v} \frac{\partial \bar{\tau}_{yz}}{\partial z} \right) dz.$$

Here $\bar{\tau}_{xz}$ and $\bar{\tau}_{yz}$ denote the components of the shearing stress. After transformation and integration from the surface to the top of the atmosphere

$$\delta F = -\, 2\pi a \cos \varphi \bar{V}_0 \bar{\tau}_0 - 2\pi a \cos \varphi \int_0^\infty \left(\bar{\tau}_{xz} \frac{\partial \bar{u}}{\partial z} + \bar{\tau}_{yz} \frac{\partial \bar{v}}{\partial z} \right) dz, \quad (4)$$

where $\bar{V}_0$ denotes the average wind velocity at anemometer level. In (4) the first term represents the change of kinetic energy due to the mean surface stress $\bar{\tau}_0$, the second the change due to frictional dissipation within the atmosphere.

The mean surface stress and the mean surface wind speed can be computed from values previously given by Palmén (1955); resulting values of the first term in (4) are given in table 5, and shown graphically in

TABLE 5. Frictional dissipation of kinetic energy due to surface drag at given latitudes (10^3 kj m^{-1} sec^{-1}).

Latitude	0°	5°	10°	15°	20°	25°	30°
Dissipation	1.8	7.9	25.6	19.8	8.7	2.6	0.7

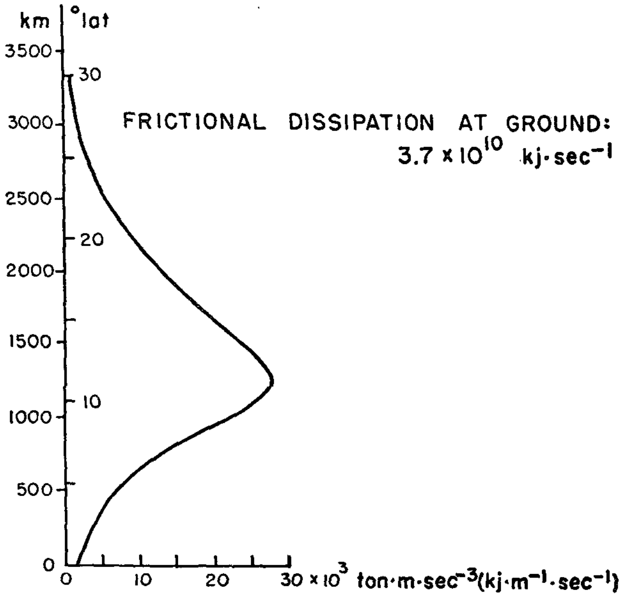

FIG. 10. Frictional dissipation of kinetic energy per unit length of a meridian due to surface stress.

fig. 10. According to this figure the total dissipation of kinetic energy between the equator and 30 deg N is 3.7×10^{10} kj per sec, a somewhat smaller value than the production of kinetic energy in the layer 1000–900 mb which amounts to 4.6×10^{10} kj per sec. The dissipation within the atmosphere however has not yet been considered. In the friction layer above anemometer level this dissipation cannot be very large since the air moving upward across the 900-mb surface in low latitudes has a somewhat higher kinetic energy than the air moving downward at higher latitudes (fig. 3). We may assume that the total frictional dissipation from the ground to 900 mb has the order of 4.5×10^{10} kj per sec, the same value as the generation in the layer. According to Brunt (1934) the dissipation in the free atmosphere is approximately as large as in the ground friction layer. If this estimate is correct, the total frictional dissipation between the equator and 30°N would be about 9×10^{10} kj per sec, and the net production of kinetic energy by the mean meridional circulation would amount to 22×10^{10} kj per sec approximately. It should be pointed out that in this calculation generation and dissipation of kinetic energy associated with space and time variations of the wind field have not been considered.

The net production just computed may be compared with the export of kinetic energy across latitude 30 deg N. If K denotes the kinetic energy per unit volume the meridional flux of kinetic energy across a latitude circle between the isobaric surfaces p_1 and p_2 is given by

$$\frac{2\pi a \cos \varphi}{g} \int_{p_2}^{p_1} \overline{Kv} dp$$

$$= \frac{2\pi a \cos \varphi}{g} \int_{p_2}^{p_1} (\bar{K}\bar{v} + \overline{K'v'}) dp. \quad (5)$$

The flux is again divided into two components, the flow due to the meridional circulation and the flow due to eddies. According to Mintz (1955) the eddy flux across 30 deg N evaluated between the ground and 200 mb was 16.6×10^{10} kj per sec for January and February 1949; Pisharoty (1955) computed 20.3×10^{10} kj per sec for the layer between the ground at 150 mb for the same period. Since the meridional circulation at 30 deg N is very weak, the first term on the right hand side of (8) only makes a small contribution of about 1.5×10^{10} kj per sec. Considering in addition the contribution from the layer 150–100 mb, the flow of kinetic energy northward across 30 deg lat can be estimated as 22×10^{10} kj per sec. If there is no appreciable net flow across the equator, the export of kinetic energy from the tropics equals the generation by the mean meridional circulation.

3. Heat transfer

In order to calculate the heat transfer (H) across 15 deg latitude, the net mass flow of 0.2 m per sec in fig. 5 was eliminated through adjustment of the meridional component at all levels by this amount. Then the mean height, temperature and moisture of all isobaric surfaces at 100-mb intervals for 15 deg lat was determined from all available stations for January 1956. The mean data are shown in table 6, together with the adjusted values of the mean meridional circulation.

We shall compute the tropospheric heat transfer due to the mean meridional circulation. This transfer is given by

$$H = \frac{2\pi a \cos \varphi}{g} \int_{p=100}^{p=1000} (A\bar{\Phi} + c_p\bar{T} + L\bar{q})\bar{v} dp, \quad (6)$$

where Φ is the geopotential, A the heat equivalent of mechanical work, c_p the specific heat at constant pressure, T temperature, L latent heat of condensation and $\bar{q}$ specific humidity. The vertical distribution of the heat content per gram of air is shown in table 6 and fig. 11. We observe a decrease of heat content from the surface to about 750 mb, then a slow increase

TABLE 6. Mean sounding and meridional circulation at 15 deg N, Winter 1955–56.

p(mb)	z(m)	$\bar{T}$(°A)	$\bar{q}$(g/kg)	$\bar{v}$(m/s)	$Ag\bar{z}$ (cal/gm)	$c_p\bar{T}$ (cal/gm)	$L\bar{q}$ (cal/gm)	Σ (cal/gm)
1000	40	297	12.5	−2.4	0	71.3	7.3	78.6
900	1000	291	7.5	−3.0	2.4	70.0	4.4	76.8
800	2010	285	4.5	−1.5	4.8	68.5	2.6	75.9
700	3135	281	2.5	−0.3	7.5	67.5	1.5	76.5
600	4500	273	1.7	0.5	10.8	65.6	1.0	77.4
500	5830	265	1.3	0.6	14.0	63.6	0.8	78.4
400	7510	253	0.8	0.6	18.0	60.8	0.5	79.3
300	9590	239	0.4	1.4	23.0	57.5	0.2	80.7
200	12,310	218		3.0	29.6	52.3	—	81.9
100	15,600	198		0	37.4	47.6	—	85.0

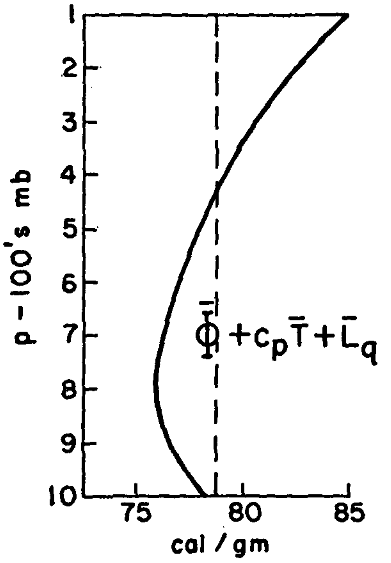

FIG. 11. Vertical distribution of total heat content of air (cal per gm) at 15 deg N lat for January 1956.

in the middle troposphere followed by a more rapid increase in the high troposphere.

This distribution is typical of the trades. At 100 mb the heat content is very high and can hardly be related to heating processes near the earth's surface. It is therefore clear that any tropospheric circulation cell transporting to higher latitudes heat given up by the earth's surface to the atmosphere must terminate at or slightly below 100 mb, i.e., in the vicinity of the mean altitude of the tropopause.

Eliminating the pressure-weighted vertical average of the heat content of the air which is 78.5 cal per gm and denoting deviations from this mean with an asterisk, the heat transfer is essentially given by $(A\bar{\Phi}^* + c_p\bar{T}^* + L\bar{q}^*)\bar{v}$; this function is portrayed graphically in fig. 12. The integrated amount is 1×10^{15} cal per sec or 4.2×10^{12} kj per sec which may be compared with previous estimates of the heat flow requirement across 15 deg lat. According to Baur and Philipps (1935) the flux in January is 2×10^{15} cal per sec. This figure cannot be considered as certain within a fairly large margin of error; other estimates have been much lower. The uncertainties of heat flux calculations from radiation and rainfall estimates are such that we can merely state that the computed transport by the meridional circulation has the right order of magnitude to account for the required poleward heat export across 15 deg N lat.

In the foregoing, the meridional eddy flux of sensible and latent heat has been neglected. According to Mintz (1955) the sensible heat transport by eddies was 0.03×10^{15} cal per sec for January–February 1949, a negligibly small value. To determine the eddy moisture flow the product Lqv was formed from daily values at 850, 700 and 500 mb for all stations used for computing figs. 5–6. When one sums this product over the stations at each level, the mean value $L\overline{qv}$ gives the total moisture transport. The eddy transport, denoted by primes, is obtained by setting $L\overline{q'v'} = L\overline{qv} - L\bar{q}\bar{v}$ where the primes indicate both space and time departures from the mean for the winter. Values of $L\overline{q'v'}$ were irregularly distributed with height. In units of 10^2 cal per gm × m per sec values were 1.7 at 850 mb, − 0.2 at 700 mb and 1.0 at 500 mb, where the positive sign indicates northward transport. If one assumes a mean value of 1.0 as representative for the whole layer 1000–500 mb—probably high—the northward eddy moisture flow is 0.2×10^{15} cal per sec, still nearly an order of magnitude less than the heat transport due to the mean meridional circulation. The total calculated heat transport, including the moisture flow, is 1.2×10^{15} cal per sec.

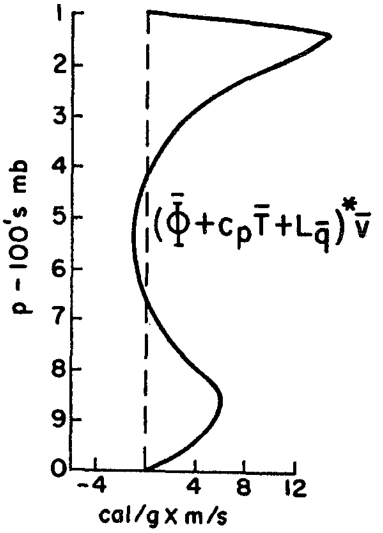

FIG. 12. Heat transport across 15 deg N lat computed from fig. 11 and fig. 5 (justified for no net mass flow).

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