A preliminary Investigation of the Daily Meridional Transfer of Atmospheric Water Vapour between the Equator and 40° N

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ABSTRACT

As a preliminary investigation into the daily variation of the meridional flux of atmospheric water vapour between the equator and 40° N, the transport has been computed from the analysed fields of specific humidity (q) and the v-component of the wind field. The analysis is based on the observations taken at a single synoptic time. The computations confirm earlier results that the bulk of the flux takes place between the 850 mb pressure surface and the surface of the earth. Between the 500 mb pressure surface and the upper limits of the Hadley and Ferrel cells the flux amounts to about 10 % of the total flux.

It is further shown that the Hadley cell circulation is the most important agent for transporting water vapour equatorward. North of 30° N the eddy flux is the dominant factor but appears to be of a much weaker intensity than the Hadley cell flux.

The computed values of the difference between the zonally averaged evaporation and precipitation ($\bar{E} - \bar{P}$), for five-degree latitude belts, obtained in this study, show greater intensity in the equatorial belt than those obtained empirically by other authors (e.g. Jacobs). Palmen obtained a similar result some time ago.

Harmonic analyses show that in the equatorial belt proper, no definite wave number can be ascribed to the eddy vapour transport. In the subtropics, wave numbers one and two appear to be the active agents, whilst beyond 30° N wave numbers four and six take over.

1. Introduction

It is well known that the release and transport of latent heat may over certain regions of the earth be of prime importance when it comes to explaining or forecasting the development and movement of atmospheric weather systems, or in explaining the general circulation features of the atmosphere. The mechanism by which this takes place is that latent heat release and transport changes the baroclinicity of the atmosphere which in turn affects the physical quantities such as momentum, vorticity and kinetic energy. In the study of tropical cyclones the occurrence of precipitation over a sufficiently large region is often cited as a trigger action mechanism for the off-set and development of a tropical storm. Due to lack of moisture studies, especially on a short time basis, it has not been possible to investigete the question as to what extent changes in momentum, vorticity, etc. are caused by latent heat release and transport.

The importance of the study of atmospheric water vapour has therefore been stressed, *inter alia*, by STARR (1951) and SUTCLIFFE (1956), and during the past decade several studies on either regional basis, e.g. WHITE (1951), BENTON and ESTOQUE (1954), HUTCHINGS (1957, 1961) or global basis, e.g. PEIXOTO (1958, 1959), STARR and PEIXOTO (1958 *a*, 1958 *b*), LUFKIN (1958), have been undertaken. For most of these studies upper-air data was still far from satisfactory and the accepted method of computing the vapour transfer was to evaluate time-averaged fluxes at the available radiosonde stations for a number of pressure surfaces and then either perform a spatial analysis with subsequent zonal averaging or to select aerological stations according to their geographic positions after which zonal averaging took place.

All water-vapour studies undertaken so far have been mainly concerned with seasonal or annual means, as the paucity of the observational network prohibited research on a daily basis. Since the inception of the International Geophysical Year the spatial distribution and instrumentation of the observational network has vastly improved over many regions of the world. The consequence of this is that, with the aid of the better understanding of the synoptic features of the atmosphere accrued during the last ten years, daily spatial fields of the meteorological elements can be constructed. DEFANT and VAN DE BOOGAARD (1963 *a*) showed the feasibility of obtaining the wind fields at a single time instant over an extended region from the equator to 40° N.

The purpose of the present undertaking will be to deal with the computation of the meridional flux of water vapour by means of the spatially analysed specific humidity and meridional wind component field at a single time instant and to show that the results are comparable with the hitherto obtained appropriate seasonal averages.

2. Data and analysis

In addition to the already available north–south v-component of the windfield for 00 GMT 12 Dec. 1957 as used by DEFANT and VAN DE BOOGAARD (1962*a*), the specific humidity field was analysed between 10° S and 50° N for the surface of the earth and the pressure surfaces 950, 900, 850, 700 and 500 mb. The specific humidity was computed from the mixing-ratio data which in turn were available from the in-detail plotted aerological soundings. The data were originally taken from the daily bulletins U.S. WEATHER BUREAU (1957), supplemented by the IGY microcards WORLD METEOROLOGICAL ORGANIZATION (1957) and aerological data published by the JAPANESE METEOROLOGICAL AGENCY (1957). The synoptic reports from land stations and ships were taken from the IGY cards. About 150 radiosondes and 300 pilot balloon observations were involved in the construction of the specific humidity and wind charts.

All data were plotted and analysed on Mercator projection synoptic maps, scale 1:25,000,000, standard parallel 22½°. As the centres of intensity of wind and specific humidity are closely related to frontal systems, extensive use was made of the already available contour analysis of the subtropical and polar fronts. The consultation of analysed synoptic maps, made available by various National Services was regarded as a great help. Despite the large amount of data available, over many regions the situation was still far from ideal and the question as to what extent a certain radiosonde observation was representative over a certain area was posed constantly. It was found expedient to plot a map of the cloud observations as given by the synoptic reports and in this way a fairly accurate delineation of moist and dry regions could be given. Over some regions, e.g. between Hawaii and the American continent, where upper air reports were completely absent, surface reports from ships were used in the best possible way to fill the gap.

3. Physical considerations

The equation for water balance of the atmosphere stems from the law of conservation of mass and prescribes that, in the absence of nuclear reactions, water substance cannot be created or destroyed. Consequently, the only way to change the water content of a mass particle at any point of the atmosphere is by addition or extraction of water.

As the storage effect of precipitable water in the atmosphere is small, even over short time intervals, the existence of regions of excess evaporation over precipitation necessitates transport of moisture by atmospheric wind systems. This transport can take place in any of the three phases, vapour, liquid or solid form. It is generally conceded that the largest transport occurs in the vapour phase, but over certain regions such as mountain plateaus or in the tropics, transport of moisture in the form of clouds may be appreciable. This study, like other investigations, will be concerned with the transport of water vapour only.

A complete derivation of the equation relating the zonally averaged meridional transport of water vapour to the difference between the zonally averaged evaporation and precipitation will not be given here. For this the reader is referred to PEIXOTO (1959) or LUFKIN (1959). Only a few of the basic steps are repeated here.

The equations are presented in the normal spherical co-ordinate system with λ longitude, φ latitude and z vertically upward. u, v and w are the scalar velocity components in the east, north, and upward direction, respectively.

Combining the equation for the conservation of water vapour with the equation of continuity, and performing a zonally averaging process and integration from the surface of the earth ($z = 0$) to the height $z = \bar{h}$ we obtain

$$\frac{\partial}{\partial t}\int_0^{\bar{h}} \bar{\varrho}\bar{q}\,dz + \frac{1}{a^2 \cos\varphi}$$
$$\times\left\{\frac{\partial\varphi}{\partial}\int_0^{\bar{h}} \bar{\varrho}\,\overline{qv}\,a\cos\varphi\,dz + \frac{\partial}{\partial z}\int_0^{\bar{h}} \bar{\varrho}\,\overline{qw}\,a^2\cos\varphi\,dz\right\}.$$
$$= \int_0^{\bar{h}} \bar{\varrho}\overline{\sum}\,dz. \quad (1)$$

Here ϱ is the density of air, q the specific humidity, a the radius of the earth and $\sum$ the sum of the "sources and sinks" of water vapour. The bar indicates averaging with respect to λ. In the derivation of the above formula it has been assumed that deviations from the mean density $\bar{\varrho}$ are small and therefore $\varrho = \bar{\varrho}$.

With respect to the third term it is found that q rapidly approaches zero with increasing height. Therefore the contribution by the upper limit to this term will be very small and may be neglected. Therefore,

$$\frac{\partial}{\partial z}\int_0^{\bar{h}} \bar{\varrho}\,\overline{qw}\,a^2\cos\varphi\,dz \approx (\bar{\varrho}\,\overline{qw})_{z=0}\,a^2\cos\varphi, \quad (2)$$

which is the contribution to the integral at the surface $z=0$. This contribution represents the flux of vapour through the lower boundary and as such may be included in the source and sink function $\overline{\sum}$ on the right-hand side of equation (1).

For practical computational purposes equation (1) may now be converted to pressure integration by the use of the hydrostatic approximation. This gives

$$\frac{\partial}{\partial t}\int_{\bar{p}_s}^{\bar{p}_h} \bar{q}\frac{dp}{g} + \frac{1}{a\cos\varphi}$$
$$\times\frac{\partial}{\partial\varphi}\int_{\bar{p}_h}^{\bar{p}_s} \overline{qv}\cos\varphi\frac{dp}{g} = \int_{\bar{p}_h}^{\bar{p}_s}\overline{\sum}\frac{dp}{g}. \quad (3)$$

For the pressure integration limits the approximation has been introduced that at $z=0$, $p=\bar{p}_s(\varphi)$ and at $z=\bar{h}$, $p=\bar{p}_h(\varphi)$, where the bar also represents zonal averaging.

Finally, assuming that the water vapour content of the atmosphere changes very little with respect to time, steady state conditions exist and consequently from (3):

$$\frac{1}{a\cos\varphi}\frac{\partial}{\partial\varphi}\int_{\bar{p}_h}^{\bar{p}_s}\overline{qv}\cos\varphi\frac{dp}{g} = \int_{\bar{p}_h}^{\bar{p}_s}\overline{\sum}\frac{dp}{g}. \quad (4)$$

This equation states that the latitudinal vergence of the pressure-integrated zonally averaged meridional flux of water vapour is equal to the pressure-integrated and zonally averaged sources and sinks of water vapour.

Of the sources and sinks of water vapour evaporation and precipitation are the largest. In comparison to the water moving through the evaporation–precipitation cycle, the amount of water entering and leaving a zone by diffusion and by transport in the form of liquid or solid may be neglected. (BENTON, BLACKBURN and SNEAD, 1950). Thus (4) may now be written:

$$\frac{1}{a\cos\varphi}\frac{\partial}{\partial\varphi}\int_{\bar{p}_h}^{\bar{p}_s}\overline{qv}\cos\varphi\frac{dp}{g} = \bar{E} - \bar{P}, \quad (5)$$

where $\bar{E} - \bar{P}$ represents the zonally averaged difference between evaporation and precipitation.

By evaluating the left-hand side of equation (5) it will thus be possible to get an estimate of the zonally averaged difference of evaporation and precipitation for belts of latitude. Estimates of this latter quantity have been obtained by JACOBS (1950) for the northern Pacific and Atlantic oceans by means of computing the evaporation and precipitation for seasonal periods. Due to inherent difficulties, e.g. computing rainfall over ocean areas and evapotranspiration over land areas, no estimates of $\bar{E} - \bar{P}$ are available for latitude belts as a whole. The aim of this undertaking will be to obtain $\bar{E} - \bar{P}$ by means of evaluating the atmospheric field of meridional vapour transport through the application of equation (5).

4. The meridional distribution of the zonally-averaged field of specific humidity, meridional wind field and water-vapour flux

(*a*) *The field of specific humidity.* From the analysed charts q, the specific humidity, was read-off at every five-degree latitude and longitude intersection and the zonal average computed from the formula:

$$q = \frac{1}{L}\int_0^L q\,dx \approx \frac{1}{N}\sum_{i=1}^{N} q_i. \quad (6)$$

Here L represents the "effective" length of the latitude, i.e. the part of the latitude actually occupied by the atmosphere and excludes the

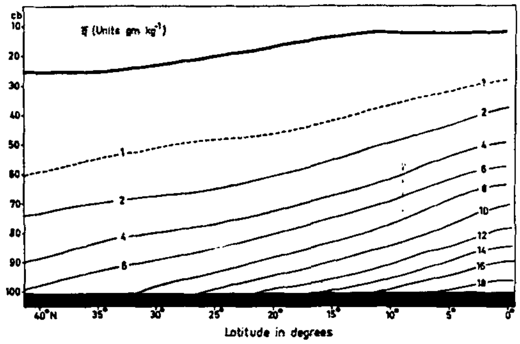

FIG. 1. The distribution in the meridional plane of $\bar{q}$, the zonally-averaged specific humidity.

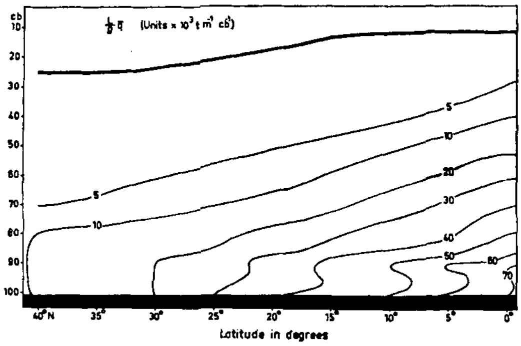

FIG. 2. Water content of zonal rings of one meter latitudinal extent and one centibar thickness as a function of latitude and pressure.

section occupied by the topography of the earth. The delineation of the earth's contours at each latitude was obtained from the excellent publication by BERKOFSKY and BERTONI (1960).

The distribution in the meridional plane of the zonally averaged specific humidity $\bar{q}$ is shown in Fig. 1 in units of gm/kg. The characteristic distribution of decreasing values of $\bar{q}$ with decreasing pressure and increasing latitude is the same as the 1950 winter season average obtained by PEIXOTO (1958). In the present case the values of q in the tropics are about one to two units higher than the winter average, but over the remainder of the region values are nearly the same.

The water content of zonal rings of one meter latitudinal extent, one cb thickness and in units of 10^3 tons/m/cb is presented in Fig. 2. Here again only the "effective" part of the volume of the zonal ring is taken into consideration. On account of this the somewhat smooth appearance of Fig. 1 disappears in the lower layers of the atmosphere and instead a tendency for constant water content with elevation is indicated between the surface and 900 mb. Otherwise, similar to Fig. 1, the water content of the air is highest in the lower troposphere of the equatorial regions decreasing with elevation and northward latitude.

Integration with respect to pressure gives the zonally averaged water content of the atmosphere as a function of latitude:

$$W = \int_{\bar{p}_h}^{\bar{p}_s} \bar{q}\,\frac{dp}{g}. \qquad (7)$$

The upper limit of integration $\bar{p}_h$ was determined by assuming that q is zero at the upper limits of the Hadley and Ferrel cells (indicated by the heavy lines in Figs. 1 and 2). The profiles of water content, being available to 500 mb, were thus completed by allowing the curves to go to zero smoothly between this pressure surface and the upper limit. Results of this computation are shown in Table 1. (Here and in all further computations planimetric methods were used to evaluate the integrals.)

The first row gives the water content for the zonal volumes in units of 10^6 tons/m for Dec. 12th, 1957. The second row gives the same figures but in units of gm/cm^2 which makes comparison with results obtained by PEIXOTO (1958) and BANNON and STEELE (1960) easier. Perusal of these figures shows that in the present case the values obtained in the tropics are higher than either the winter season or the January averages. Several reasons can be attributed to this. Both PEIXOTO and BANNON and STEELE integrated between the limits 1000–500 mb and used the trapezoidal rule. Compared with the method employed in the present study this may lead to an underestimate of about 15 % in the tropical regions. Another reason is that in December the intertropical convergence zone is still north of its average winter position and therefore compared with the seasonal averages the December values will naturally be somewhat higher than winter averages applicable around January.

(*b*) *The meridional wind field*. This field is discussed in DEFANT and VAN DE BOOGAARD (1963*a*) but for completeness' sake a short summary will be given. The zonally averaged v-component as a function of latitude and pressure is presented in Fig. 3 and shows clearly the

TABLE 1. *The meridional distribution of water vapour content.*

Latitude	40° N	35°	30°	25°	20°	15°	10°	5°	0°	Units
12 Dec. 1957	0.40	0.51	0.66	0.84	1.07	1.29	1.61	2.07	2.57	10^6 t/m
	1.3	1.6	1.9	2.3	2.8	3.3	4.1	5.2	6.4	gm/cm²
PEIXOTO, winter 1950	1.2		1.9		2.6		3.6		4.2	gm/cm²
BANNON and STEELE, Jan. 1951-55	1.0		1.6		2.6		3.7		4.2	gm/cm²

existence of an average direct meridional circulation between the equator and 30° N known as the Hadley cell regime. Between 30° N and 40° N an indirect circulation, the Ferrel cell regime, is indicated. Of particular interest is the south to north sloping upper limit of the Hadley cell coinciding with the upper limit of the zonally averaged wet-adiabatic lapse rate of the aerological soundings.

The centre of maximum southerlies (3.5 m/sec) is situated at about 11° N and an elevation of 900 mb. The centre of northerlies (4.5 m/sec) is at the same latitude but an elevation of 200 mb. The total mass circulation of the cell is 255 million tons.

It is worth mentioning that this single day cell agrees very well in structure and intensity with the winter seasonal average obtained by PALMÉN, RIEHL and VUORELA (1958). In their case a maximum mass transport of 260 million tons at 12° N was obtained.

(c) *The field of total vapour flux.* At every five degree latitude and longitude intersection the product qv (10^{-3} ton sec^{-1}/cb/sec^{-1}) was computed and latitudinal cross-sections constructed. Fig. 4, where for convenience's sake only every ten-degree latitude is given, shows the distribution of the vapour flux as a function of pressure. One striking feature, which presents itself, is the location of the intense centres of flux around the 900 mb and the weak transport above the 500 mb surface. This characteristic pattern is identical with that obtained by BENTON and ESTOQUE (1954) for the north American continent. In the more northern latitudes the intense regions of poleward (full lines) and equatorward (dotted lines) flux are coupled to the frontal and jetstream systems, which dominate the synoptic pattern in these latitudes. In some cases the strong winds associated with the jetstream can lead to substantial transport at higher elevations even though the specific humidity is low. An example of this is shown at 40° N, and the Greenwich meridian where flux values of 60 units are indicated around 500 mb. On relative basis this is, however, still low compared with transports of 120 to 140 units occurring at lower levels. South of 20° N the flux of water vapour is more strictly confined to the lowest layers of the atmosphere and are essentially due to the trade-wind regime and lines of low-level convergence and divergence. In that respect the mode of transport of water-vapour flux is completely different to that of momentum or kinetic energy transport where the centres of maximum intensity are around the 200 mb pressure surface. The flow pattern of vapour flux is therefore much more influenced by the topographic features of the earth. It may finally be remarked that high values of the product qv in the tropics are due to high specific humidity; in the middle latitudes these are essentially due to high wind speeds.

Fig. 5 shows the distribution in the meridional plane of the total water-vapour flux per centibar layer. The solid lines indicate poleward (positive) and the dotted lines equatorward (negative) flux. Units are 10^3 tons sec^{-1}/cb. A line of zero

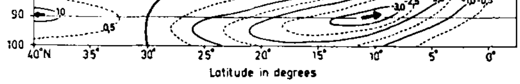

FIG. 3. The distribution in the meridional plane of the zonally averaged wind-component v (m·sec^{-1}).

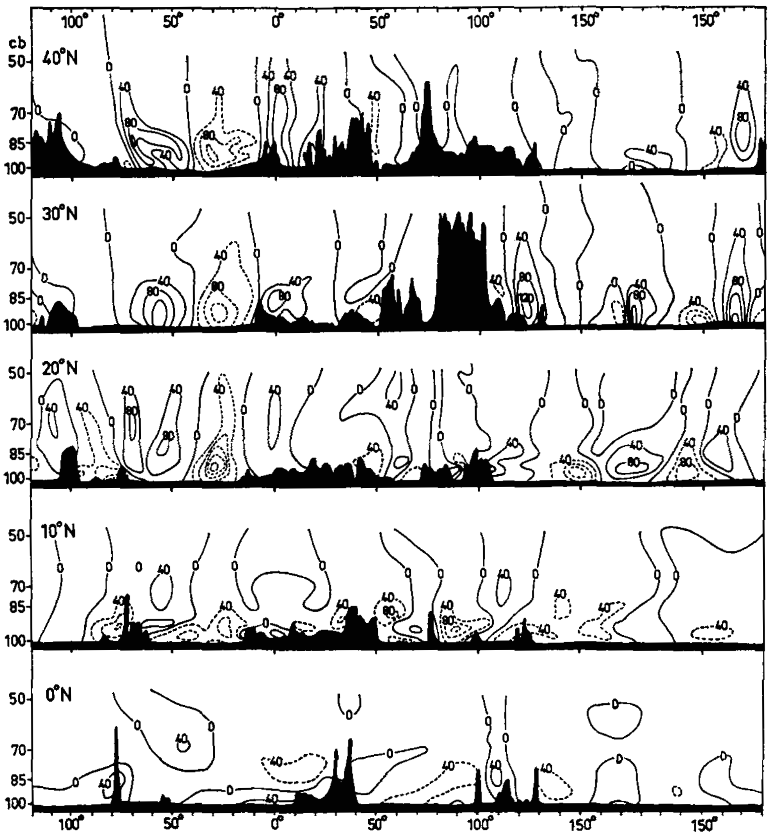

FIG. 4. The distribution with respect to longitude and pressure of the total water-vapour transport qv (10^{-3} ton sec^{-1}/cb/sec^{-2}).

transport sloping from about 26° N at the surface of the earth to about 300 mb at the equator, separates the regions of total poleward (full lines) and total equatorward transport of water vapour. The centre of maximum poleward flux of about 60 units is clearly indicated at about 30° N and 850 mb, whilst the centre of equatorward flux of intensity 125 units is situated at 11° N and 900 mb. The somewhat higher elevation of the positive centre of flux is entirely due to the upslope character of the subtropical and polar fronts, which forces moist air up to levels of stronger winds. As indicated in Fig. 5 the flux above the 600 mb pressure surface is small.

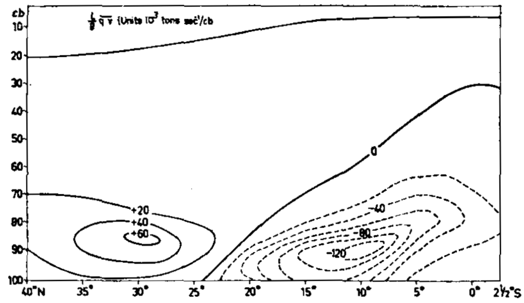

FIG. 5. The total water-vapour flux across latitude circles as a function of latitude and pressure.

(*d*) *The meridional circulation and eddy fields of vapour flux.* Indicating a zonally averaged quantity with a bar and the turbulent deviation with a prime, the total vapour transport field may be split into two component fields:

$$\overline{qv} = \bar{q}\bar{v} + \overline{q'v'}. \tag{8}$$

The first term on the right-hand side of the equation represents the contribution to the total of the vapour flux due to existing averaged meridional circulations, whilst the second term

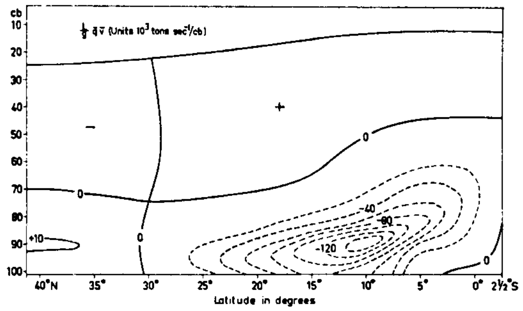

FIG. 6. Water-vapour transport due to the existence of the Hadley and Ferrel cell circulations.

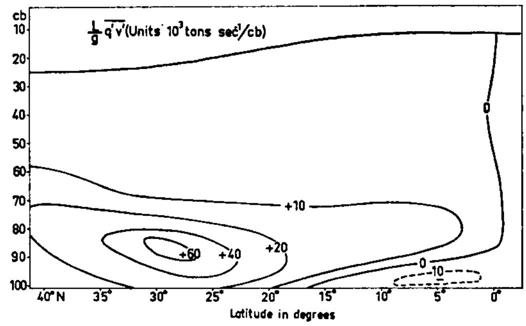

FIG. 7. Distribution in the meridional plane of water-vapour flux due to the existence of eddies.

expresses that part of the transport due to space turbulence.

The distribution in the meridional plane of the water-vapour flux performed by the Hadley cell is shown in Fig. 6. Between 25° N and the equator the distribution and capacity of the Hadley cell transport is almost identical to the total vapour flux, thus indicating that the bulk of water vapour is transported equatorward by the lower limb of the Hadley cell regime. It therefore confirms the ideas presented by PALMÉN (1955) where he stressed the importance of the Hadley cell in this respect. Due to low values of specific humidity, the function of the upper limb of the Hadley cell as a poleward agent of transport of water vapour is rendered negligible. Therefore any northward flux has to be taken care of by turbulence. The centre of Hadley cell transport, situated at 11° N and 900 mb coincides with that of the total flux but has an intensity of about 142×10^3 tons sec^{-1}/cb. The line joining the maximum transport at each latitude closely follows the line joining the maximum northerlies of Fig. 3. The capability of the lower limb of the Ferrel cell to transport water vapour poleward is indicated in Fig. 6 but comparing it with Fig. 5 it can clearly be seen that in these latitudes the turbulent flux dominates.

Equation (8) shows that the zonally averaged space-eddy water-vapour flux may be obtained by subtracting the Hadley-cell transport from the total vapour transport. The result of this computation is given in Fig. 7 in units of 10^3 tons sec^{-1}/cb. The bulk of the zonally averaged eddy flux is poleward and takes place below the 600 mb pressure surface. A small equatorward flux is indicated below 900 mb between the equator and 10° N. The maximum eddy transport of slightly more than 60×10^3 tons sec^{-1}/cb takes place around 30° N and between 800 and 900 mb.

Comparing Figs. 6 and 7 it is of interest to point out that apart from being opposite in direction the maximum intensity of the eddy flux is only about half the maximum Hadley cell flux and takes place at an somewhat higher elevation. Furthermore, it may be remarked that the eddy transport distribution may vary much from day to day, but that the daily variation of the Hadley cell flux is very small. It is, therefore, not expected that the characteristic distribution of Fig. 7 will differ much from day to day or for the season as an whole. The intensities obtained here are, however, much stronger than the seasonal values as will be pointed out in the next section.

(e) *The integrated field of water-vapour flux.* It is now of interest to study the three types of fluxes as pressure-integrated functions of latitude. The full line in Fig. 8*a* graphically shows the pressure-integrated profiles of the fluxes as a function of latitude. The total flux (full line) is negative south of 22½° N reaching a minimum of 18.5×10^5 tons/sec at 10° N. North of 22½° N the integrated total transport is positive reaching a maximum 12×10^5 tons/sec at 34° N. The Hadley cell transport indicated by the dotted line shows a minimum of 22×10^5 tons/sec also at latitude 10° N and a weak maximum of 2×10^5 tons/sec at latitude 35° N. The difference between the two provides the pressure-integrated eddy flux and is indicated by the dashed line in Fig. 8*a*. Apart from a slightly negative value in the vicinity of the equator the integrated eddy flux is positive everywhere. The eddy

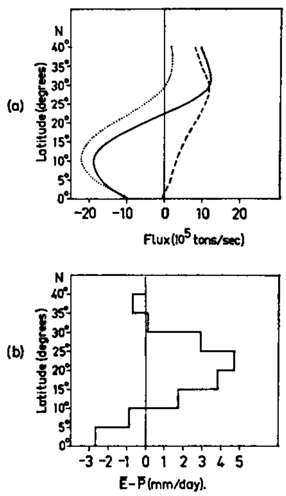

FIG. 8 *a*. Latitudinal profiles of the pressure-integrated total (full line), mean meridional (dotted line) and eddy (dot-dashed line) fluxes of water vapour.

FIG. 8 *b*. The zonally-averaged vergence of water-vapour in belts of five-degree latitude expressed as $\bar{E}-\bar{P}$ (mm/day).

transport is, however, insignificant between the equator and 20° N when compared with the integrated Hadley cell flux. North of 20° N the eddy flux becomes more dominant but still does not reach the magnitude of the Hadley flux.

Table 2 provides a means of comparing the single day integrated transports of total, eddy and Hadley fluxes with the values of the 1950 winter season (six months) obtained by PEIXOTO (1958). As only ten-degree values were computed in the latter case, Table 2 is restricted to these latitudes.

Qualitatively both computations show a similar distribution. In the tropics proper the Hadley contribution for the winter average is also dominant but less than the values obtained for single day. The winter eddy transport is also much smaller. It is positive (poleward) everywhere, which is not the case for the instantaneous computations where negative (equatorward) eddy flux occurs in the equatorial belt.

(*f*) *The zonally averaged vergence of water vapour.* Having the total flux of water vapour available as a function of latitude it is now possible, assuming steady state, to compute the convergence and divergence of water vapour in various latitude belts with the aid of eq. (5). Fig. 8*b* shows in units of mm/day for five-degree latitude belts the difference between the zonally averaged evaporation and zonally averaged precipitation as a function of latitude. In principle these results could be compared with actual zonally averaged evaporation and precipitation data, but due to difficulties of observing precipitation over the ocean and evapotranspiration over land, such figures are not available.

A qualitative verification may be made by observing that the distribution of $\bar{E}-\bar{P}$ falls into three distinctive belts, which in turn are closely connected with three separate weather regimes. In the belt 0°–10° N the precipitation exceeds the evaporation. As the prevailing weather regime is that of the equatorial trough zone with its abundant rainfall, for precipitation to exceed the evaporation is therefore naturally to be expected. The belt 10°–35° N is characterized by excess evaporation over rainfall, which can naturally be attributed to the fact that at this time of the year the dry anticyclonic weather conditions of the subtropical high pressure belt prevails over these latitudes. North of 35° N precipitation again dominates due to the active subtropical and polar front weather regime. In this respect the computational results can at least qualitatively be substantiated.

TABLE 2. *The integrated transports of total, eddy, and Hadley water-vapour fluxes in units of 10^5 tons/sec.*

Top row results obtained in present study, bottom row winter season 1950 obtained by PEIXOTO.

40° N			30°			20°			10°			0°		
Total	Eddy	Hadley	Total	Eddy	Hadley	Total	Eddy	Hadley	Total	Eddy	Hadley	Total	Eddy	Hadley
+ 10.0	+ 8.5	+ 1.5	+ 11.8	+ 11,8	0.0	− 5.0	+ 8.0	− 13.0	− 18.5	+ 3.5	− 22.0	− 9.5	− 0.5	− 9.0
+ 5.4	+ 3.8	+ 1.6	+ 3.8	+ 3.3	+ 0.5	− 3.7	+ 2.4	− 6.1	− 14.4	+ 2.0	− 16.4	− 9.3	+ 0.2	+ 9.5

TABLE 3. *Values of $\bar{E}-\bar{P}$ in mm/day.*

Latitude belt	0°–10° N	10°–20° N	20°–30° N	30°–40° N
Dec. 12th, 1957	−1,7	+2.8	+3.8	−0.3
PEIXOTO, Sept.-Mar. 1950	−1.0	+2.2	+1.6	+0.4
JACOBS, Sept.-Jan. (Atlantic and Pacific)	−0.7	+1.7	+2.4	+1.9

As direct quantitative verification is impossible it may first of all be of interest to compare the results of similar studies by other investigators. Table 3 gives such a comparison. The top row gives the values of $\bar{E}-\bar{P}$ obtained in the present investigation but this time in latitude belts of ten degrees. The second row is computed from the six-monthly winter averages of total flux of PEIXOTO's data and the last row the results of the investigations by JACOBS (1950) where three-monthly averages were computed. The figures in the table refer to the three months, November to January, and were interpolated from data given by Jacobs. Perusal of the figures in Table 3 clearly indicates that the single day quantities are higher than the seasonal values and will immediately elicit the questionability of comparing daily with seasonal averages. If the latter are strongly dependent on the presence of eddies, which themselves vary in intensity from day to day then such comparison is impossible, but if the seasonal quantity largely depends on such features as Hadley cell circulations, etc., which do not vary strongly from day to day, then such comparisons are well justified. Looking at Fig. 8*a* it will be noticed that in the belt 0°–10° N the total flux of water vapour is for more than 85 % due to the averaged meridional circulation features. These features, as discussed, before do not differ very much from the seasonal averages. The only differences that can be detected are that in the above-mentioned belt the meridional gradients of both the zonally averaged north–south wind and the zonally averaged specific humidity field are somewhat more intense than the seasonal values, which in turn will contribute to a more intense convergence of vapour flux. The computed $\bar{E}-\bar{P}$ values obtained for the belt 0°–10° N in the present investigation must be expected to be somewhat more negative than the seasonal value. Taking this into account and considering that (*a*) the data available in 1950 for this belt was poor, (*b*) the seasonal average is over a period of six months and (*c*) the neglect by PEIXOTO to take the intense vapour flux at around 900 mb into account, the computed single day $\bar{E}-\bar{P}$ value of −1.7 mm/day versus the six-monthly mean of −1.0 mm/day is plausible and acceptable.

A similar comparison is not possible in the case of the results obtained by Jacobs as his values only apply to the Pacific and Atlantic oceans and not to the latitude belts as a whole. It is, however, of interest to speculate somewhat on the implications if it is assumed that both values are valid for the month of December and that Jacobs' value is valid for the entire ocean area in the latitude belt under consideration. In such a case it is possible, by considering the land–ocean area ratio, to compute the zonally averaged excess evaporation over the zonally averaged precipitation for land alone. Performing this computation an $(\bar{E}-\bar{P})$ value of −14 cm/month (−4.7 mm/day) for the land area alone is obtained. If, for the moment, it is assumed that no evapotranspiration occurs ($\bar{E}=0$) the zonally averaged precipitation $\bar{P}$ for land in this belt will amount to 14 cm/month. From estimates obtained by MÖLLER (1951) the zonally averaged rainfall for land in the belt 0°–10° N for the month of December may be taken to be about 10 cm/month, which means that the theoretically computed $\bar{P}$ value is 40 % higher than the observed average. If allowance for evapotranspiration is made, e.g. $\bar{E}=10$ cm/month, the theoretically computed $\bar{P}$ value will exceed the observed $\bar{P}$ value by 140 %.

On the other hand, if it is assumed that the theoretically computed $(E-P)$ of −1.7 mm/day (−5 cm/month) for the total belt 0°–10° N is equally applicable to land and ocean, and $\bar{E}$ for the ocean alone is assigned the generally acceptable value of 10 cm/month, the zonally averaged precipitation $\bar{P}$ for the ocean alone will amount to 15 cm/month. This figure is about 10 % higher than the $\bar{P}$ value obtained by Möller for the ocean alone. Considering the uncertainty of estimating rainfall over the oceans such a figure is well within the margin of error.

PALMÉN (1955), considering the seasonal average of the Hadley cell mass circulation alone, obtained an $(\bar{E}-\bar{P})$ value for the belt 3° S–12° N of -3.0 mm/day (present investigation -3.2 mm/day for the belt $2\frac{1}{2}$° S–10° N). He pointed out that if Jacobs' results were correct, the net zonally averaged eddy vapour flux had to be large. The magnitude of this quantity was unknown to him at that time. However, both this investigation and that of Peixoto indicate that the eddy vapour flux in these latitudes is extremely small and may on the whole be neglected.

From the above discussions the following remarks may be made regarding the latitude belt 0°–10° N.

(1) Accepting the estimates of Jacobs and the estimates of precipitation and evaporation by various investigators, the mass circulation of the Hadley cell has to be reduced by more than half to allow for decreased equatorward flux of water vapour. This is contradictory to observational evidence.

(2) If in addition the mass circulation is of correct order of magnitude the net eddy vapour flux has to be increased by large amounts. This is also against observational evidence.

(3) In considering the sinks the thought may be entertained that a large net divergence through the lateral boundaries of liquid water in the form of cloud takes place. This will have to be equally true for daily as well as seasonal considerations. Over certain regions transport of moisture in the form of liquid water may certainly be important, but for a large region considered here this form of transport must be considered to be more of a corrective than a decisive nature.

(4) If the Hadley cell and eddy water-vapour flux are of correct magnitude the thought may be entertained that the $(\bar{E}-\bar{P})$ value obtained by Jacobs for the equatorial Pacific and Atlantic oceans is too low for the equatorial oceans as a whole. This means either that the Indian ocean will have to make up the deficit or that Pacific–Atlantic ocean value is too low by about a factor of two.

For the remaining latitude belts, poleward of 10° N, considerations similar to the one above cannot be undertaken because the eddy flux plays an equal or dominant role in these latitudes. However, comparing the eddy flux obtained here with Peixoto it can be seen that the single day's flux is much stronger than the seasonal average. Eddy activity is well above normal for this particular day and the resulting $(\bar{E}-\bar{P})$ may be taken as to represent the extreme values for the latitude and season in question.

(*g*) *The eddy flux of water vapour in the wave number domain.* Following VAN ISACKER and VAN MIEGHEM (1956) the eddy flux of water vapour along a latitude circle α, pressure surface p and for unit pressure thickness may be expressed in the form:

$$\tau_q = \frac{2\pi a \cos\varphi}{g}\left[\tfrac{1}{2}(A_n\alpha_n + B_n\beta_n)\right]. \qquad (9)$$

Here τ_q represents the contribution of the disturbance of wave number n on to the total eddy flux of water vapour, and where A_n, B_n and α_n, β_n are the Fourier coefficients of q and v resp. It is assumed that it is permissible to develop q and v, as functions of λ in Fourier series. Integrating (9) with respect to pressure gives:

$$T_q = \frac{2\pi a \cos\varphi}{g}\int_{p_n}^{p_s}\left[\tfrac{1}{2}(A_n\alpha_n + B_n\beta_n)\right]dp, \qquad (10)$$

where T_q represents the pressure-integrated eddy-vapour flux for wave number n. Having the Fourier coefficients available it is therefore an easy matter to compute the contribution of each wave number n to the total eddy transport.

Harmonic analyses up to wave number 15 were performed along each five degree latitude circle and for the surface and the 950, 900, 850, 700 and 500 mb pressure surface. In those regions where the earth's orography interfered with the pressure surface, q was taken as zero and v equal to the latitudinal average. Fig. 9*a* displays the contribution to the eddy flux by each wave number as a function of latitude. Units are 10^5 tons/sec.

Caution must, of course, be used in giving physical meaning to a mathematical device such as harmonic analysis to a single day's data but nevertheless some interesting features may be pointed out. In the belt 0°–10° N insignificant contributions are noticed, bearing out the fact that the main contribution in these latitudes is due to the mean meridional circulation. In the latitude belt 10°–25° N the significant positive

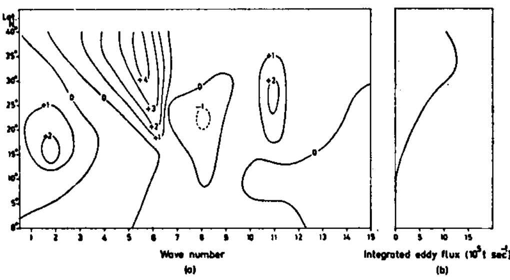

FIG. 9. Eddy flux of water vapour as a function latitude and wavenumber.

contribution is given by the long waves 1 and 2. As this is the belt of the subtropical anticyclone, the cells of which are strongly controlled by the land–sea distribution, it is very natural to surmise that the main contribution to the vapour flux should be on that scale. From 20° N, however, higher wave numbers begin to show their influence. Beginning with wave number 6 at 20° N the influence of wave numbers 5, 4, 3, etc., becomes dominant as one proceeds northwards. In the belt 25°–40° N waves 5 and 6, belonging to the Rossby waves, are on this day the important contributors to the poleward eddy flux of water vapour. In this respect it contradicts the results obtained by PEIXOTO and SALTZMAN (1958) where at latitude 30° N wave number 2 was the most important contributor at pressure surfaces 850, 700 and 500 mb. No explanation for this difference can be offered but it would be of interest to perform computations of the type presented in this section for a number of days in order to see how the spectrum varies. Fig. 9*b* shows the profile of the integrated eddy vapour flux as a function of latitude obtained by summing the wave numbers 1–15. Apart from the fact that maximum transport is situated somewhat more to the north, it compares well with the eddy transport obtained in Fig. 9*a*. This indicates that contributions by wave numbers higher than fifteen are small and negligible.

(*h*) *Conclusions.* In this investigation it has been shown that, for a single day and for the latitude belt 0°–40° N, distributions in the meridional plane, of correct order of magnitude, of total, mean and eddy water-vapour fluxes can be obtained by means of the space-analysed specific humidity and north–south wind component fields. The essential characteristics of the fluxes are:

(1) They occur mainly at low elevation.

(2) The total flux is directed equatorward south of 22½° N and reaches a maximum intensity at 10° N. North of 22½° N the total flux is poleward with a maximum intensity at 30° N. In absolute magnitude the total equatorward transport is about twice the poleward flux.

(3) The Hadley cell circulation is the principal agent for carrying water vapour equatorward.

(4) The eddy transport is generally poleward throughout the latitude belt. It is insignificant in low latitudes but dominates north of 25° N. The maximum eddy flux occurs at around 30° N but in absolute magnitude it amounts to only half the Hadley cell flux.

As the object of this study has been to show that for a single day acceptable results can be obtained by performing space analysis of specific humidity and the north–south wind component, the assumptions introduced have been similar to those presented by other investigators considering water-vapour transport on a seasonal basis. For single-day periods the assumption of steady state may be the most questionable one and therefore a similar investigation covering a few consecutive days is planned in order to obtain an idea of the error involved.

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REFERENCES

BANNON, J. K., and STEELE, L. P., 1960, Average water vapour content of the air. *Geophysical Memoirs* No. **102**. British Met. Office.

BENTON, G. S., BLACKBURN, R., and SNEAD, V., 1950, The role of the atmosphere in the hydrological cycle. *Trans. Amer. Geophys. Un.*, Vol. **31**, pp. 61–73.

BENTON, G. S., and ESTOQUE, M. A., 1954, Water vapour transfer over the North American continent. *Journ. Met.*, Vol. **11**, pp. 462–477.

BERKOFSKY, L., and BERTONI, E. A., 1960, Topographic charts at one degree intersections for the entire earth. Geophysical Research Directorate, Research Notes No. 42.

DEFANT, F., and VAN DE BOOGAARD, H. M. E., 1963 *a*, The Global Circulation Features of the Troposphere between the Equator and 40° N, based on a Single Day's Data. Part 1. The structure of the basic meteorological fields. *Tellus*, Vol. **15**, pp. 251–260.

HUTCHINGS, J. W., 1957, Water vapour flux and flux of divergence over southern England. *Quart. J.R. Met. Soc.*, Vol. **83**, pp. 30–48.

HUTCHINGS, J. W., 1961, Water vapour transfer over the Australian continent. *Journ. Met.*, Vol. **18**, pp. 615–634.

JAPANESE MET. AGENCY, 1957, *Aerological data of Japan.*

LUFKIN, D. H., 1959, Atmospheric water vapour divergence and the water balance at the earth's surface. Scientific Report No. **4**. General Circulation Project. Mass. Inst. of Technology.

MÖLLER, F., 1951, Vierteljahrskarten des Niederschlags für die ganze Erde. *Geogr. Mitt.*, Vol. **95**, pp. 1–7.

PALMÉN, E., 1955, On the mean meridional circulation in low latitudes of the northern hemisphere in winter and the associated meridional and vertical flux of angular momentum. *Soc. Scientiarum Fennica. Commentationes Physico-Mathematica*, **XVIII**, 8.

PEIXOTO, J. P., 1958, Hemisphere humidity conditions during the year 1950. Scientific Report No. **3**. General Circulation Project. Mass. Inst. of Technology.

PEIXOTO, J. P., 1959, A campa da divergencia do campa do fluxo do vapor do agua na atmosfera. *Revista da Faculdade de Ciencias de Lisboa 2a, Ser. B.*, Vol. **VII**, pp. 25–56.

PEIXOTO, J. P., and SALTZMAN, B., 1958, Harmonic analysis of the mean northern hemisphere water vapour distribution for the year 1950. *Geofisica pure e applica*, Vol. **41**, pp. 183–188.

STARR, V. P., 1951, Applications of energy principles to the general circulation. *Compendium of Meteorology*, pp. 568–574. American Met. Soc.

STARR, V. P., and PEIXOTO, J. P., 1958 *a*, On the global balance of water vapour and the hydrology of deserts. *Tellus*, Vol. **10**, pp. 188–94.

STARR, V. P., and WHITE, R. M., 1954, Balance requirements of the general circulation of the atmosphere. *Geophysical Research Papers*, **35**. Geophysics Research Directorate. Cambridge, Mass.

SUTCLIFFE, R., 1956, Water balance and the general circulation of the atmosphere. Presidential address Royal Met. Soc., April 25, 1956. *Quart. J.R. Met. Soc.*, Vol. **82**, pp. 385–394.

TUCKER, G. B., 1959, Mean meridional circulation in the atmosphere. *Quart. J.R. Met. Soc.*, Vol. **85**, pp. 209–224.

U.S. WEATHER BUREAU, 1957, *Daily bulletin northern hemisphere data tabulations.* December 12.

WHITE, R. M., 1951, The meridional eddy flux of energy. *Quart. J.R. Met. Soc.*, Vol. **77**, pp. 188–189.

WORLD MET. ORGANIZATION, 1957, *I.G.J. microcards.*