On the mechanism of the vertical heat flux and generation of kinetic energy in the atmosphere

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ABSTRACT

Satisfactory results have been achieved in establishing balance between the latitudinal distribution of heat sources and sinks in the atmosphere and the meridional energy flux computed from observations. Mostly, however, studies of the atmospheric heat balance have been limited to latitude belts of the whole atmosphere or at least the whole troposphere, whereas few attempts have been made to establish heat balance for layers of different height. In the present paper a tentative model of the vertical heat flux in the extratropical region of the Northern Hemisphere is presented. It is shown that the vertical heat flux, necessary for heat balance in the vertical, essentially is a result of small-scale and large-scale turbulence. In the latter type of processes the polarfront cyclones play a prominent role. It is also shown that the vertical heat flux due to large-scale eddy processes is closely correlated to the generation of kinetic energy.

1. Poleward flux of energy

The purpose of this study is not to establish a mean annual heat budget of the Northern Hemisphere, but to discuss the physical and dynamic processes involved in the poleward and vertical heat fluxes necessary for atmospheric energy balance. Since these fluxes are especially strong during the winter (December–February) the discussion will be limited to this season.

The latitudinal energy balance of the atmosphere is determined by the effect of radiation, the flux of heat from the earth's surface, the heat of condensation in the atmosphere and the divergence of the poleward atmospheric energy flux. The sum of the radiative imbalance (generally negative), the heat flux from the surface of the earth and the heat of condensation determines an atmospheric heating function which generally is positive in low latitudes and negative in high latitudes. In latitudes with a positive heating function hence the divergence of the poleward energy flux must be positive, and in latitudes with a negative heating function it must be negative. If R_a denotes the effect of radiation, Q_s the turbulent heat flux from the earth's surface and LP the heat of condensation (L = heat of vaporization, P = observed rate of precipitation), all expressed in calories per minute and cm² of the earth's surface, or by ly/min, the energy balance for a belt of latitude with the area A is given by

$$R_a + Q_s + LP + \frac{W_{\varphi n} - W_{\varphi n+1}}{A} = 0. \qquad (1)$$

Here $W_{\varphi n}$ and $W_{\varphi n+1}$ represent the poleward energy flux through the latitudinal boundaries at φ_n and φ_{n+1}.

In the following the mean values of R_a, after LONDON (1957), will be used. LP is computed from MÖLLER's data (1951) of the mean precipitation for the same season, but somewhat modified for the tropics according to MEINARDUS (1934). The last term in Eq. (1) is taken from a recent paper by HOLOPAINEN (1965). The most difficult term to estimate satisfactorily from observations is Q_s. The value of this was therefore determined as a residual from Eq. (1).

The meridional energy flux at a fixed latitude φ is given by

$$W_\varphi = \frac{2\pi a \cos\varphi}{g}\left[\int_0^{p_0} (c_p \bar{T} + \bar{\phi} + \bar{k})\bar{v}\,dp + \int_0^{p_0} (c_p \overline{T'v'} + \overline{\phi'v'} + \overline{k'v'})\,dp\right]. \qquad (2)$$

Here a is the earth's radius, g the gravity constant, c_p the specific heat of air at constant pressure, ϕ the geopotential, k the kinetic energy per unit mass and v the meridional wind component. The bars indicate mean values along a whole parallel circle for the time period considered and the primes local deviations from the mean values for all synoptic times during the same period. The first integral in Eq. (2) then represents the poleward flux of energy due to mean meridional circulations, whereas the second integral represents the poleward eddy flux of energy.

In evaluating the energy flux Holopainen considered only the atmosphere below the 100-mb surface. He disregarded the kinetic energy, which generally is a small quantity in this connection and substituted the meridional wind component by the corresponding geostrophic component in the integral for the eddy flux. With these assumptions the expression in Eq. (2) is reduced to

$$W_\varphi = \frac{2\pi a \cos\varphi}{g}\left[\int_{100\,\mathrm{mb}}^{1000\,\mathrm{mb}} (c_p \bar{T} + \bar{\phi})\bar{v}\,dp + \int_{100\,\mathrm{mb}}^{1000\,\mathrm{mb}} c_p \overline{T' v_g'}\,dp\right], \tag{3}$$

where the first integral represents the "circulation flux" and the second integral the "eddy flux".

Fig. 1 shows, according to Holopainen, the latitudinal distribution of the total poleward energy flux in winter composed by the "circulation flux" and the "eddy flux". The circulation flux was computed from the mean mass circulation in the tropical "Hadley cell" after PALMÉN & VUORELA (1963) and the reversed mean mass circulation in the extratropical "Ferrel cell" after MINTZ & LANG (1955). As eddy flux Holopainen used MINTZ' (1955) values for the winter season.

If Holopainen's values for W_φ are used in Eq. (1) the unknown quantity Q_s can be evaluated for selected zones of latitude. The achieved Q_s values, however, are not quite reliable because of possible errors in the other terms. Concerning these errors and the possibility to determine the poleward energy flux directly from the atmospheric heating function the reader is referred to Holopainen's original paper

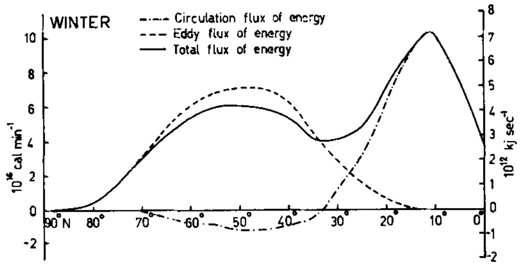

FIG. 1. Poleward flux of energy in winter after Holopainen (units: 10^{16} cal/min or 10^{12} kj/sec).

and to another recent paper by MANABE, SMAGORINSKY & STRICKLER (1965).

Fig. 1 shows two peaks in the total poleward energy flux, one very pronounced somewhat north of 10° N and one weaker around 50° N. The former is a result of the meridional circulation in the tropics (Hadley cell), whereas the latter is dominated by the eddy flux, somewhat modified by the opposite energy flux due to the reversed meridional circulation (Ferrel cell). An extrapolation indicates that the poleward energy flux vanishes around latitude 5° S, which corresponds to the mean position of the equatorial trough in winter. Around 32–33° N the entire energy flux is of the eddy character.

2. Model of energy balance in low and high latitudes

It is suitable to study the heat budget separately in the region north of latitude 32° N and south of it. The northern region may be characterized as the region of the Ferrel cell in winter; here the energy flux largely is dominated by eddy processes. The southern region between 32° N and 5° S represents the region of the Hadley cell which here dominates the poleward energy flux.

A tentative energy budget for these two regions is presented in Fig. 2. In this the atmosphere between the North Pole and 5° S is divided into four blocks (I–IV), separated by the quasi-horizontal 500-mb surface and a vertical wall at 32° N. In each block the mean radiation deficit R_a (after London), the liberation of latent heat LP and the divergence of the vertical heat flux must be compensated by the poleward eddy flux of heat through the vertical wall at 32° N, since no poleward flux of energy

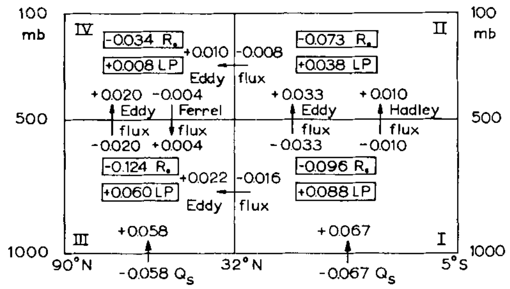

FIG. 2. Tentative energy budget of the region 5° S–90° N in winter and vertical energy fluxes through the 500-mb surface (units: ly/min).

occurs at 5° S and at the North Pole. In the figure the vertical heat flux at the earth's surface is evaluated as a residual term from Eq. (1); the vertical flux is assumed to approach zero at 100 mb. The heat of condensation is apportioned to the layers below and above 500 mb according to the vertical distribution of specific humidity. All values in Fig. 2 are given in ly/min. The values of the eddy heat flux through 32° N on the two sides differ because the total heat transport through this latitude was converted to ly/min by considering the horizontal areas of the regions.

The values of Q_s in Fig. 2 seem too large. The average ratio Q_s/LP is 0.85 in the northern part and 0.53 in the southern part. The above ratio is approximately equal to the ratio between the flux of sensible and latent heat from the ground and should, according to BUDYKO (1963), have an annual mean value of 0.45 in the northern part and 0.20 in the southern part.[1] Although Budyko's annual values of the above ratio possibly are too small for the winter season, when Q_s is large in oceanic regions (PETTERSSEN *et al.*, 1962), the most probable explanation of the discrepancy is that the rate of precipitation has been underestimated. But also a relatively slight reduction of the radiational cooling R_a would considerably reduce the values of Q_s.

The vertical arrows in Fig. 2, with the figures inserted, show the vertical heat fluxes at the 500-mb level. The separation between the eddy fluxes and the fluxes due to the mean meridional mass circulation was achieved in the following way.

The vertical heat flux across an horizontal area A is expressed by

$$\int_A c_p \varrho w T dA = A c_p [\varrho w][T] + A c_p [(\varrho w)'' T''], \quad (4)$$

where ϱ is the density of air and w the vertical wind component. Here the brackets denote an areal mean value of the quantities for the time period considered and the two primes the deviations from the mean values. In the latitudinal zones considered $[\varrho w]$ disappears, because the latitude wall at 32° N separates the closed Hadley and Ferrel circulations from each others. By assuming hydrostatic balance and replacing w by $\omega = dp/dt$ the mean vertical heat flux per unit area, $[W_z]$, may be expressed by

$$[W_z] = -\frac{c_p [w'' T'']}{g}. \quad (5)$$

It should be observed that the above expression represents the vertical heat flux across a horizontal level. If this is replaced by a corresponding isobaric surface a term $-[\omega'' \phi'']/g$ representing the flux of potential energy has to be added. The temperature fluctuations in an isobaric level are generally somewhat smaller than the corresponding variations in a horizontal level. However, the vertical flux computed for a horizontal surface from Eq. (5) would not be too erroneous if the fluctuations ω'' and T'' are computed in an isobaric surface corresponding to the mean pressure at the level.

From the meridional distribution of the mass circulation in the Hadley and Ferrel cells and the corresponding distribution of the mean temperature the vertical heat flux resulting from these mean circulations can be evaluated. The difference between the total flux computed from Eq. (5) and the circulation flux then gives the eddy flux. Both types of fluxes are marked in Fig. 2. In the southern part the Hadley circulation carries energy upwards across the 500-mb surface, whereas the Ferrel circulation in the northern part carries energy downwards. In both regions, however, the eddy flux is directed upwards and is numerically much larger than the circulation flux.

It should be stressed that a reduction of Q_s and a corresponding increase of LP, to bring

[1] This ratio, Q_s/LE, where E is the rate of evaporation, may somewhat differ from Q_s/LP, but for the mean values in the whole region the difference is not essential.

these quantities in closer agreement with Budyko's values of Q_s/LE, would somewhat modify the vertical eddy fluxes through the 500-mb surface. However, the energy budget in Fig. 2 would not essentially be changed through this correction.

The physical nature of the vertical eddy flux of heat is largely different from the nature of the meridional heat flux. In the former case a large spectrum of eddies of very different scales has to be considered, whereas the lateral or meridional eddy flux almost entirely is caused by large-scale eddies. Hence the vertical eddy flux of heat cannot directly be computed from regular synoptic data, as was the meridional eddy flux. In the case of the vertical heat flux, the regular small-scale turbulence and the convection associated with vertical instability plays an important role. This is especially true in the tropical region with its strong Cu convection (RIEHL & MALKUS, 1958). In the extratropical region, where Cu convection in winter is strongly reduced by the greater atmospheric stability, the importance of large-scale eddies (cyclonic disturbances and different types of waves in the westerlies) become dominating. The importance of this type of vertical heat flux will be discussed next.

3. The mechanism of the vertical heat flux in extratropical regions

MINTZ (1955) has presented values of the poleward eddy heat flux for standard isobaric layers up to 100 mb. By using his data for the eddy flux of heat through 32° N and the mean Q_s value for the northern zone and by redistributing the other terms, R_a and LP, in the equation for energy balance the scheme in Fig. 3 was achieved. Here the horizontal arrows, with the corresponding figures, mark the lateral influx of heat measured per unit area of the layers. The vertical arrows represent the total vertical flux of sensible heat across standard isobaric surfaces, starting from 1000 mb up to 100 mb where the heat flux is assumed to vanish. All numerical values are again expressed in ly/min.

The vertical heat flux results from the combined effects of physically quite different processes, as stated before. In the vicinity of the 1000-mb surface the heat flux, Q_s, is almost entirely a result of very small-scale turbulence

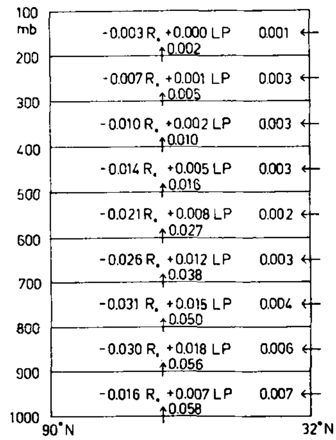

FIG. 3. Mean energy budget of standard isobaric layers in the cap north of 32° N. Arrows show horizontal and vertical flux of sensible heat (units: ly/min).

through which heat is carried upwards. In winter this mechanism dominates over the relatively warm oceans, whereas over the colder continents the turbulent heat flux in some regions even is directed downwards. The turbulent flux of latent heat measured by evaporation from the ground furnishes the lower atmosphere with water vapor that higher up in the atmosphere condensates and contributes to the heating of the atmosphere. This latter heating occurs above the condensation level and determines the distribution of LP in Fig. 3, as does also the (generally positive) lateral influx of water vapor from the south. Above the condensation level the liberation of latent heat strongly contributes to Cu convection especially over warm ocean surfaces.

The turbulent heat flux due to small-scale turbulence is directed upwards in the lower atmosphere in spite of the fact that this, on the average, shows vertical stability. The importance of this type of turbulent heat flux, referred to as "convective turbulence", has been stressed by ERTEL (1942) and by PRIESTLEY, SHEPPARD & SWINBANK (1947, 1952).

Eq. (5) shows that the vertical heat flux is measured by the average value of the product $[\omega'' T'']$. Here ω'', T'' comprise fluctuations of very different scale including the largest ones associated with synoptic disturbances, planetary waves and even mean meridional circulations. At the ground, where the ω'' values of

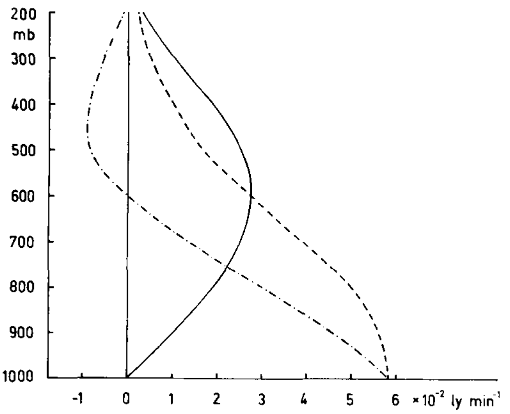

FIG. 4. Separation between small-scale vertical heat flux (dash-dotted curve) and synoptic-scale vertical heat flux (solid curve). The dashed curve represents the mean total flux derived from Fig. 3.

synoptic-scale motions approach zero, the large-scale turbulent heat flux approximately vanishes, but higher up in the atmosphere the large-scale turbulence takes over a considerable part of the vertical heat flux. In winter convective clouds appear over large regions, especially over the oceans, but these convective clouds mostly don't reach very high altitudes (PETTERSSEN *et al.*, 1962), except for limited regions of active cyclonic systems. In the upper clear regions of the atmosphere the regular dynamic turbulence obviously tends to transport heat downwards owing to the prevailing stability of the atmosphere. Only in regions of strong and high reaching Cu convection this tendency of downward small-scale turbulent heat flux may be overcome by the influence of strong Cu convection.

In the extratropical region it is not possible to estimate exactly the contribution of the different parts of the large turbulent spectrum to the mean vertical heat flux. Only a crude attempt to do so may here be made. For this purpose we make the following assumptions:

(1) The vertical heat flux is divided into two parts: the flux caused by small-scale turbulence, including Cu convection, and the flux caused by large processes which can be studied by aid of synoptic charts.

(2) At the earth's surface the small-scale turbulence determines the vertical heat flux, and the large-scale turbulent heat flux is zero. With increasing height the latter component of the heat flux increases in importance and reaches a maximum value somewhere in the middle troposphere, where the fluctuations in the synoptic-scale vertical velocity and temperature are considerable.

(3) In the free atmosphere above the general level of Cu activity the small-scale turbulence tends to carry heat downward and the large-scale turbulence upward. This tendency prevails up to the vicinity of the tropopause, where the synoptic-scale disturbances become less effective due to decreasing fluctuations in ω and T. Above the tropopause the synoptic-scale disturbances possibly even carry heat downward.

Referring to the above assumptions, founded on synoptic experience, Fig. 4 has been constructed. Here the dashed curve represents the total vertical heat flux taken from Fig. 3, the dash-dotted curve the heat flux caused by different types of small-scale processes, and the solid curve the remaining heat flux due to large-scale (synoptic-scale) processes, including mean meridional circulations. According to this scheme the large-scale heat flux increases from zero at the ground level to a maximum value around 600 mb and approaches again zero somewhere above 200 mb. Correspondingly, the small-scale heat flux has its highest value at the ground and becomes negative above 600 mb.

It should once more be stressed that the model presented in Fig. 4 is only tentative and could be subjected to reasonable criticism. For lower latitudes and other seasons of the year the components of the heat flux would deviate considerably from the above scheme.

4. Vertical heat flux and generation of kinetic energy

An interesting relationship between the large-scale vertical heat flux in Fig. 4 and the generation of kinetic energy can be derived in the following way. If k denotes the kinetic energy per unit mass the local change of kinetic energy in the region A is determined by (WHITE & SALTZMAN, 1956)

$$\frac{A}{g}\int_0^{p_0}\frac{\partial[k]}{\partial t}\,dp = \frac{2\pi a\cos\varphi}{g}\int^{p_0}\overline{(k+\phi)v}\,dp - \frac{AR}{g}\int_0^{p_0}\frac{[\omega''T'']}{p}\,dp + \frac{A}{g}\int_0^{p_0}[\mathbf{v}\cdot\mathbf{F}]\,dp, \qquad (6)$$

where R is the gas constant, $\mathbf{v}$ denotes the horizontal wind vector and $\mathbf{F}$ is the frictional force per unit mass. By using the same approximations as in deriving Eq. (3) and considering that $\bar{v}$ vanishes at the boundary between the Hadley and Ferrel cells at latitude 32° N the following formula for the maintenance of kinetic energy can be derived:

$$\frac{2\pi a \cos\varphi}{g}\int_{100\,\mathrm{mb}}^{1000\,\mathrm{mb}} \overline{k' v_g'}\,dp - \frac{AR}{g}\int_{100\,\mathrm{mb}}^{1000\,\mathrm{mb}} \frac{[\omega'' T'']}{p}\,dp + \frac{A}{g}\int_{100\,\mathrm{mb}}^{1000\,\mathrm{mb}} [\mathbf{v}\cdot\mathbf{F}]\,dp = 0. \tag{7}$$

In the above equation the first term represents the net (geostrophic) influx of kinetic energy through 32° N and the second term the conversion between available potential and kinetic energy in the region between this latitude and the North Pole. In deriving this formula only the kinetic energy of large-scale horizontal motion is considered. When all small-scale motions are disregarded the contribution of the vertical velocities to the total kinetic energy is negligibly small and can therefore be disregarded. In Eq. (7) the atmosphere above 100 mb was, as in the previous discussion, disregarded which possibly is not quite permissible, but hardly will appreciably influence the discussion.

For constant pressure the vertical heat flux (W_z) in Eq. (5) is proportional to generation of kinetic energy per unit area according to Eq. (7). Since in the latter term only large-scale motions are considered this term should therefore only be compared with the vertical heat flux owing to large-scale processes, or with the heat flux marked by the solid curve in Fig. 4. By multiplying this part of $[W_z]$ with $R/c_p p$ the solid curve can be transformed into the corresponding curve in Fig. 5 which is an expression for the increase of kinetic energy per unit area and pressure. In the figure $[W_z]$ has been converted into kilojoules per m² and sec, or kilowatts/m². Graphically evaluated from 1000 to 200 mb (average tropopause level) it gives a mean kinetic energy generation of about 5.5 watts/m² in the troposphere.

Estimates of the mean frictional dissipation of kinetic energy in the atmosphere are not very reliable, especially because very little is known about the energy dissipation in the free atmosphere above the friction layer. According to Brunt (1934) the rate of energy dissipation in the Northern Hemisphere amounts to about 5 watts/m². For a complete balance in the region considered here the first term in Eq. (7) should be taken in consideration. According to Pisharoty (1955) the mean flux of kinetic energy across the southern boundary amounts to about $20\cdot10^{10}$ kj/sec which, considering the total area A, represents an increase of mean kinetic energy of about 1.8 watts/m². Hence the total rate of dissipation north of 32° N in winter should be roughly 7.5 watts/m². This is considerably more than the dissipation estimated by Brunt, but in our estimate only winter conditions are considered when both generation and dissipation strongly exceed their annual means.

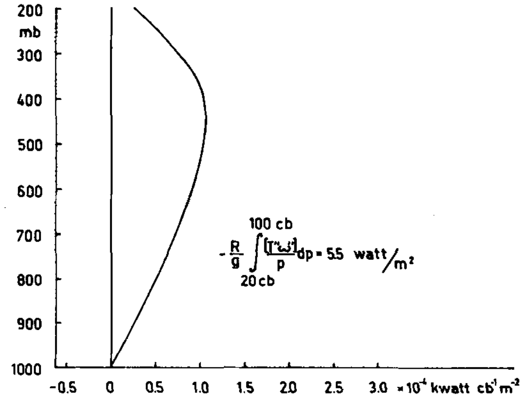

Fig. 5. Curve representing the mean generation of kinetic energy per m², cb and sec ($-R/gp[T''\omega'']$) computed from the solid curve in Fig. 4.

The mean rate of generation of kinetic energy computed from Fig. 5 includes also the negative generation resulting from the mean Ferrel cell. This was by Holopainen (1965) estimated at $-11\cdot10^{10}$ kj/sec in December–February for the cap north of 30° N or at -0.9 watts/m². Hence the large-scale eddies (cyclones and longer waves in the westerlies) would, on the average, produce kinetic energy at a rate of about 6.5 watts/m².

5. Synoptic evidence of the eddy heat flux and energy coversion in polar-front cyclones

Several attempts have been made to compute the rate of generation of kinetic energy in polar-front cyclones. Because these cannot

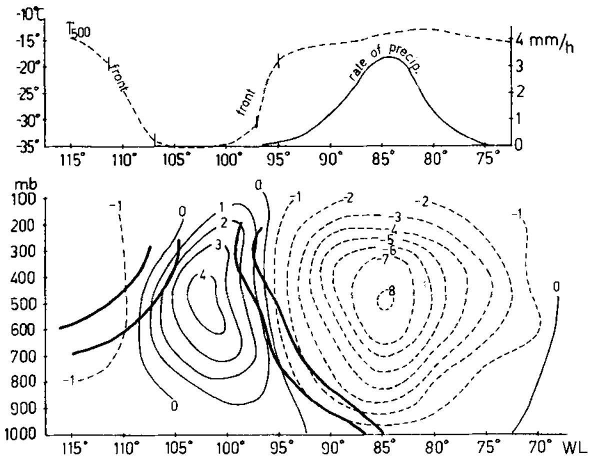

FIG. 6. Zonal cross-section along latitude 40° N through a typical winter cyclone in the U.S., showing the frontal layers and the vertical wind component (unit 10^{-4} cb/sec) after Danard. The upper part gives the distribution of the 500-mb temperature and the rate of observed precipitation (in mm/h) along the same section.

strictly be treated as dynamically closed systems the results largely depend on the boundaries selected. In three cases studied by PALMÉN (1958), PALMÉN & HOLOPAINEN (1962) and DANARD (1964) the rate of kinetic energy generation was estimated at 52, 21 and 18 watts/m², respectively. The first of these synoptic cases was a very extreme one, where the energy conversion largely exceeded the average value in cyclones. The two other cases represented more normal conditions. We may therefore estimate the rate of kinetic energy generation in an average polar-front cyclone at about 20 watts/m². Since the area used for the above computations was about 1/20 of the total area of the cap north of 32° N, five such cyclones would suffice to produce the amount of kinetic energy computed from Fig. 5.

The close correlation between temperature and vertical velocity resulting in upward heat flux and simultaneous generation of kinetic energy is illustrated by Fig. 6, constructed by use of Danard's data for January 21, 1959. Here the frontal layers separating the cold air from the warmer air masses on both sides along a zonal cross-section at latitude 40° N and the isotachs for the vertical wind component are presented. The upper part of the figure shows the 500-mb temperature in the same section and the rate of measured rainfall. The close relationship between rain intensity and vertical velocity and between the latter and the temperature is very clear. The region of strongest ascending motion ($\omega < 0$) occurs on the east side of the cold air where southerly winds prevail, whereas the strongest subsidence again occurs in the cold air with wind components from north. Hence the resulting heat flux is directed northward and upward. Fig. 6 may be considered typical of extratropical cyclones during their most active phase of development. The correlation between the temperature distribution and the general flow pattern was very similar in the two other synoptic cases mentioned before. The same type of correlation resulting in large-scale upward and poleward flux of energy and in simultaneous conversion between potential and kinetic energy characterizes most polar-front disturbances before they become occluded (PALMÉN & NEWTON, 1951) and indicates, that they represent a very important mechanism for the budget of heat and kinetic energy in the extratropical regions.

The "eddy kinetic energy" generated in this way due to conversion of "eddy available potential energy" is partly transformed into "mean zonal kinetic energy" as described e.g by WIIN-NIELSEN (1959). Hence the cyclonic disturbances do not derive their surplus energy from the pre-existing zonal circulation, but appear as active cells for the maintenance of this against the frictional dissipation. This fundamental role of the extratropical disturbances was already emphasized by the "Bergen School" at a time when, due to lack of upper air data, it was not possible to compute the effect quantitatively.

REFERENCES

BRUNT, M. A., 1934, *Physical and dynamical meteorology*. University Press, Cambridge, 411 pp.

BUDYKO, M. I., 1963, Atlas of the heat balance of the earth (in Russian). Moscow.

DANARD, M. B., 1964, On the influence of release of latent heat on cyclone development. *J. of Appl. Meteor.*, **3**, 27–37.

ERTEL, H., 1942, Der vertikale Turbulenz-Wärmestrom in der Atmosphäre. *Meteor. Zeitschr.*, **59**, 250–253.

HOLOPAINEN, E. O., 1965, On the role of mean meridional circulations in the energy balance of the atmosphere. *Tellus*, **17**, 285–294.

LONDON, J., 1957, A study of the atmospheric heat balance. Final Rep., Contr. AF 19 (122)-165 (AFCRC-TR-57-287), 99 pp., New York Univ., Dept. Meteor. and Oceanogr.

MANABE, S., SMAGORINSKY, J., and STRICKLER, R. F., 1965, Simulated climatology of a general circulation model with a hydrologic cycle. *Monthly Weather Rev.*, **93**, 769–798.

MEINARDUS, W., 1934, Die Niederschlagsverteilung auf der Erde. *Meteor. Zeitschr.*, **51**, 345–350.

MINTZ, Y., 1955, The total energy budget of the atmosphere. Final Rep., Gen. Circ. Proj., Contr. AF 19 (122)-48, Dept. Meteor., Univ. California, Los Angeles, XIII, 1–6.

MINTZ, Y., and LANG, J., 1955, A model of the mean meridional circulation. Final Rep., Gen. Circ. Proj., Contr. AF 19 (122)-48, Dept. Meteor., Univ. California, Los Angeles, VI, 1–9.

MÖLLER, F., 1951, Viertelsjahrskarten des Niederschlages für die ganze Erde. *Petermanns Geogr. Mitteil.*, **95**, p. 1.

PALMÉN, E., 1958, Vertical circulation and release of kinetic energy during the development of hurricane "Hazel" into an extratropical storm. *Tellus*, **10**, 1–23.

PALMÉN, E., and HOLOPAINEN, E. O., 1962, Divergence, vertical velocity and conversion between potential and kinetic energy in an extratropical cyclone. *Geophysica*, 8, 89–113.

PALMÉN, E., and NEWTON, C. W., 1951, On the three-dimensional motions in an outbreak of polar air. *J. Meteor.*, 8, 25–39.

PALMÉN, E., and VUORELA, L. A., 1963, On the mean meridional circulations in the Northern Hemisphere during the winter season. *Quart. J. Roy. Meteor. Soc.*, **89**, 131–138.

PETTERSSEN, S., BRADBURY, D. L., and PEDERSEN, K., 1962, The Norwegian cyclone models in relation to heat and cold sources. *Geof. Publ.*, **24**, 243–280.

PISHAROTY, P. R., 1955, The kinetic energy of the atmosphere. Final Rep., Gen. Circ. Proj., Contr. AF 19 (122)-48, Dept. Meteor., Univ. California, Los Angeles, XIV, 1–61.

PRIESTLEY, C. H. B., and SHEPPARD, P. A., 1952, Turbulence and transfer processes in the atmosphere. *Quart. J. Roy. Meteor. Soc.*, **78**, 489–529.

PRIESTLEY, C. H. B., and SWINBANK, W. C., 1947, Vertical transport of heat by turbulence in the atmosphere. *Proc. Roy. Soc.* A, **189**, 543–561.

RIEHL, H., and MALKUS, J. S., 1958, On the heat balance in the equatorial trough zone. *Geophysica*, **6**, 503–537.

WHITE, R. M., and SALTZMAN, B., 1956, On conversion between potential and kinetic energy in the atmosphere. *Tellus*, 8, 357–363.

WIIN-NIELSEN, A., 1959, A study of energy conversion and meridional circulation for the large-scale motion in the atmosphere. *Month. Weath. Rev.*, **87**, 319–332.

О МЕХАНИЗМЕ ВЕРТИКАЛЬНЫХ ПОТОКОВ ТЕПЛА И О ГЕНЕРАЦИИ КИНЕТИЧЕСКОЙ ЭНЕРГИИ В АТМОСФЕРЕ

Были достигнуты удовлетворительные результаты при установлении баланса между зональным распределением источников и стоков тепла и меридиональным потоком энергии, вычисленным из наблюдений. Однако большинство исследований баланса тепла в атмосфере до сих пор ограничивалось рассмотрением лишь некоторых зональных поясов всей атмосферы или, по крайней мере, только всей тропосферы, в то время как было сделано мало попыток для установления теплового баланса для слоев, находящихся на разных высотах. В настоящей статье представлена предварительная модель для описания вертикального потока тепла для северного полушария, исключая тропики. Показано, что вертикальный поток тепла, необходимый для баланса тепла по высоте, существенно является результатом мелко- и крупномасштабной турбулентности. В последнем типе процессов большую роль играют циклоны полярного фронта. Показано также, что вертикальный поток тепла, обусловленный крупно-масштабными турбулентными процессами, тесно коррелирует с генерацией кинетической энергии.