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Energy transformations in the subcloud layer of hurricane Gladys (1975)

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ABSTRACT

The question of the energy balance of air penetrating toward a hurricane core in the subcloud layer is examined at hand of research flight observations, when a whole hurricane mission was carried out at and below cloud base. The pressure drop in this model is taken as due to energy supplied by oceanic sensible heat transfer to the atmosphere. The subcloud layer depth so calculated agrees well with the observed one. All sensible heat supply is retained in the subcloud layer while about half of the evaporated moisture must be transported by turbulent exchange to higher layers. Kinetic energy balance is achieved in the subcloud layer without substantial turbulent transfer from above. But, this result does not hold at hurricanes of greater intensity when heat or kinetic energy transport from above is required or when, alternatively, the surface sensible heat source must be intensified. The kinetic energy balance of inflow above the subcloud layer must be achieved largely by turbulent exchange in cloud towers penetrating to the upper troposphere.

1. Introduction

Observations in hurricanes made during the years since 1955 when Doppler-radar wind measurements became available have shown these storms to be quasi-balanced vortices. The local oceanic heat source underneath creates a warm core in the rain area—outside the still warmer eye—and abstraction of absolute angular momentum from atmosphere to ocean takes place. Because of the latter the tangential wind component decreases upward from inflow to outflow layers. The balanced vortex is achieved when energy transfer from the ocean and momentum transfer to the ocean are so adjusted that the cyclostrophic thermal wind around the core will maintain the interior temperature field so that at most the surface energy source is exported from the hurricane by a requisite small transverse circulation. Otherwise the outflow could increase without limit, far beyond the capability of any surface interaction process to maintain the temperature field, and the vortex would fall apart. Such imbalance may indeed occur in many initial cores; these, therefore, vanish again after a life on the time scale of only hours.

While the cyclostrophic wind is rather well established by the observations, the mechanism for producing the central warm core outside the eye has remained a more contentious subject. Land data are too sparse and obviously unsatisfactory for an adequate description; almost no aircraft data exist below cloud base for obvious reasons. Yet it is precisely the subcloud layer that requires exploration. As has been emphasized repeatedly, the amount of convergence into a rain area and attendant precipitation is generally unrelated to vortex development or even just development of kinetic energy in the atmosphere. As long as air keeps moving up the same moist adiabat, there will be no change in the mass distribution which would lead to pressure changes, even when the rain amount reaches 50 cm/day and more. The requirement is for the ascending motion to shift to warmer moist adiabats with decreasing radius from the center, in order to establish and maintain the

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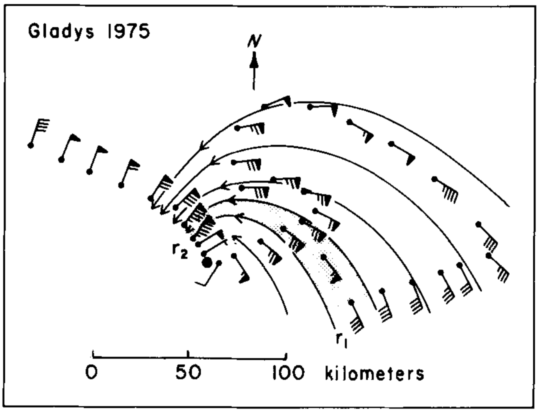

Fig. 1. Winds in hurricane Gladys measured on September 29, 1975, in the Western Atlantic by aircraft flying near 300 m height except at 150 m in the north-western portion in the outskirts. Shaded: inflow path for calculations.

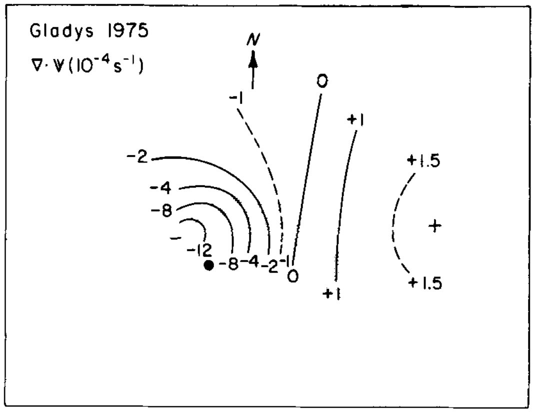

Fig. 2. Horizontal velocity divergence computed from Fig. 1.

interior warm temperature field requisite for positive solenoidal circulation and maintenance of kinetic energy of a vortex.

The transfer of latent heat from ocean to atmosphere can contribute to this shift only through lowering cloud base, a process of importance in the outskirts where cloud base lowers from 600–700 m to around 300 m; roughly at this height it remains constant all the way to the eye, largely at constant dewpoint. One must, therefore, look for sensible heat transfer, either from the ocean or other potential sources, to provide the increase in thermal energy requisite for the ascent to shift to higher energy levels. In general, air temperature at the ocean surface becomes almost constant when ocean temperature (T_w) minus air temperature near the ocean (T_a) attains 2–3 °C. Using the concept of isothermal expansion, the central potential temperature can be computed given central pressure (Palmén and Riehl, 1956). In a simple one-dimensional calculation (Malkus and Riehl, 1960) it was shown that the necessary heat could be supplied from the ocean surface without any unusual forcing of exchange processes. Nevertheless, the argument has been advanced that the computed sensible heat transfer is too large. In this paper it will be shown that the previous position not only can be well maintained for hurricanes up to medium intensity but that for very intense storms this energy source becomes insufficient, unless increases of $T_w - T_a$ are permitted beyond 2–3 °C.

2. Hurricane Gladys (1975)

The first opportunity to obtain a view on the subcloud layer, at least near cloud base in the inner portions, arose in hurricane Gladys of 1975, when on September 29 the northern semi-circle of this storm was probed with NOAA research aircraft in the western Atlantic. The plane entered from northwest. After crossing the eye it made a large clockwise loop to the east and then exited toward north. Fig. 1 shows a representative sample of winds. Whatever the reasons for this choice of flight plan, the data can be used to establish a trajectory backward from the maximum wind, shaded in Fig. 1. The maximum wind of almost 100 knots or 185 km/h occurred along the vortex displacement which was 290°, 13 km/h. The maximum wind, therefore, was not in its usual place to the right of the direction of motion, possibly a transitory state. One must assume that the pattern was maintained for the, roughly, 1 h required by the air to make the traverse along the shaded trajectory which, because of the slow rate of storm motion, had only slightly larger amplitude than the trajectory.

Computation of the horizontal divergence from Fig. 1 is difficult as usual, since the two terms in the definition of divergence in natural coordinates are opposed. Wind speed more than doubles from r_1 to r_2 (r radius) in Fig. 1, while the perpendicular cross section between streamlines shrinks. As initially drawn, its distance, say unity at r_1, decreases to 0.2 or less at r_2, so that this term predominates strongly in the second half of the shaded path. Accordingly, in Fig. 2, convergence is strongly concentrated in the right front quadrant, with slight divergence farther out. Such a pattern has often been computed, including divergence in averaged hurricanes (Riehl, 1979). The asymmetry of divergence along the low-level air path must be used in the calculations to follow.

Establishment of cyclostrophic wind balance demands hydrostatic balance as a side condition: the latter has been shown to hold closely for the observed mean pressure field near sea level. However, for a balanced vortex to persist, it must be possible to derive the same pressure from the laws of thermodynamics and motion along inflow trajectories. It will be the main problem of this report to examine this problem considered in earlier work (Riehl, 1963) in the light of the aircraft data near cloud base, where heating from condensation is not yet a factor in thermodynamic energy balances. At first the depth of the layer with sensible heat source will be computed which essentially gives cloud base height due to change of temperature lapse rate there.

3. Depth of layer with sensible heat source

We consider the first law of thermodynamics for isothermal expansion per unit mass

$$\frac{dh}{dt} = -\frac{1}{\rho}\frac{dp}{dt} \tag{1}$$

where h is the heat source per unit mass, ρ air density, p pressure, and t time. We consider only the vertical energy source for motion with the trajectory. If F_z is the vertical energy transport per unit area

$$\left(\text{units } \frac{m1^2}{t^2}/1^2\, t\right), \quad -\frac{1}{\rho}\frac{dF_z}{dz} = \frac{dh}{dt},$$

so that eq. (1) becomes

$$\frac{dF_z}{dz} = \frac{dp}{dt} \tag{2}$$

Integrating with respect to t and z,

$$\int (F_{z\,\text{upper}} - F_{z\,\text{lower}})\, dt = \int (p_{t_2} - p_{t_1}) dz$$

where t_1 and t_2 correspond to r_1 and r_2. We now will assume that $F_{z\,\text{upper}} = 0$ for this discussion and that $F_{z\,\text{lower}} = Q_s$, the sensible heat source. Further, the pressure difference $p_2 - p_1$ may be independent of height over the small distance dz. Then

$$\int Q_s\, dt = (\hat{p}_1 - \hat{p}_2)\, H$$

where H denotes the thickness of the subcloud layer and the caret denotes means over this layer. Usually, the sensible heat source, when referred to the ocean, is expressed by

$$Q_s = \rho c_p\, C_d (T_w - T_a)\, V \tag{3}$$

where c_p is specific heat at constant pressure, C_d a drag coefficient and V wind speed at a nominal level of 10 m above the surface. The word "nominal" applies here since, in the inner hurricane area, there is no solid or definite boundary between air and sea on account of wave heights and spray. The nominal level will be taken as 50 m above mean sea level where the aircraft-recorded winds can be used in eq. (3). The drag coefficient may be varied with wind speed (Garratt, 1977) but will be assumed as 2.5×10^{-3} for the present calculation, slightly lower than Garratt's mean value would be at 10 m height over the trajectory of Fig. 1. Whether the whole formula (3) has any meaning in the hurricane setting, remains questionable; it continues to be used pending development of a better formulation. Thus $Q_s\, dt = \rho c_p\, C_d (T_w - T_a) \int V\, dt$, since only wind speed will show a large variation over the trajectory. Noting that $V = ds/dt$, where s is horizontal distance along the trajectory, $\int V\, dt = l$, the trajectory length, and

$$H = \frac{\rho c_p\, C_d (T_w - T_a)\, l}{(\hat{p}_1 - \hat{p}_2)} \tag{4}$$

From the low-level flight data $p_1 - p_2 = 1005 - 987$ mb $= 18$ mb. Evaluating, $H = 360$ m, very close to the aircraft-measured depth of the subcloud layer. One may also take a deeper layer of, say, 500 m, since there are variations in subcloud layer depth which occasionally attains 500 m height. Since the base of the cloud layer is the top of the adiabatic subcloud layer, one should expect F_z to become zero at that level.

4. Budgets over the trajectory

The initial objective of the subsequent computations was to determine requirements for turbulent vertical transport through cloud base in order to maintain the observed subcloud layer structure.

A different situation was expected from regions without net convergence or with low-level divergence such as the trades. There, turbulent motions must carry out all upward transport, especially of moisture, against the mean descending motion. In the convergent region, some of the oceanic evaporation would be picked up by the mean motion. But, a residual turbulent transport was expected to exist. In contrast, the sensible heat flux up was expected to stop or even reverse at cloud base; a large downward flux of kinetic energy was thought to be needed to maintain the energy of the subcloud layer windfield against the transport of energy to the ocean through very large shearing stresses.

None of the foregoing computations materialized. Turbulent transports had been programmed for measurement by a gust probe installation in the aircraft. Unfortunately, the instrument did not function reliably in the very humid environment. Computation of turbulent vertical motion was attempted at length from the aircraft performance data, resorting in the end to Gray's (1962) formula over distances from 100 m (1 s flight time) to 2000 m (20 s). Over all these intervals computed fluxes proved to be small and quite unsystematic. Details of the extensive efforts will not be presented, since the results are not regarded as reliable. Calculations are, therefore, reduced to shed light on the question what residual turbulent transports are requisite for balances.

For the computations it must be assumed that the divergence field of Fig. 2 and the winds of Fig. 1 may be regarded as representative for the whole subcloud layer from the height of about 50 m upward. It is implied that wind direction remains constant and that nearly all vertical shear of wind speed is concentrated below 50 m over the moving and indefinite boundary between air and ocean, in contrast to measurements over land, where a logarithmic wind profile to about 300 m height becomes quickly established above the rigid land surface. Scattered bits of information support this position; for instance, waves arranged in wind rows over ocean have been found to lie precisely along aircraft-measured wind directions at 300–600 m height (unpublished data). In hurricane Donna (1960) lighthouse wind speed observations offshore from the South Florida coast agreed with earlier aircraft measurements. It remains clear that subcloud-layer wind profiles are badly needed, though very difficult to obtain with conventional measuring devices.

With the foregoing assumptions computations were made with subcloud-layer depth of 360 and 500 m. Here the 360 m height, determined earlier, will be used.

4.1. *Mass flow*

The horizontal mass flow is given by

$$M_r = V\,\delta n \rho\,\delta z \tag{5}$$

where V is wind speed, δn width of the cross section taken as 1 cm at r_1 and 0.2 cm at r_2 and $\rho\,\delta z = \delta p/g$. With these definitions the values of M_r, shown in Table 1, were obtained at r_1 and r_2. The difference, the mean upward mass transport M_z, is given in the third column. From Fig. 2 it occurs entirely in the second half of the trajectory. The slow hurricane movement has been omitted.

4.2. *Moisture*

Multiplying the mass flows by the specific humidity, 17 g/kg at the entrance and 19.7 g/kg at the exit, the horizontal moisture transports of the second line of Table 1 are obtained, also shown in energy units after multiplying by the latent heat of condensation, $L = 2.5 \times 10^3$ j/g. The vertical transport was computed assuming q = 19 g/kg in the second half of the trajectory.

Ocean evaporation (E) was determined with the conventional formula

$$E = \rho C_d (q_w - q_a)\, V \times \text{area} \tag{6}$$

where q is specific humidity and the other symbols have the same meaning as in eq. (3). The area A = 500 m^2 from Fig. 1. Using the moisture values of Fig. 3, E = 5.4 cm/d, up one order of magnitude from evaporation normally observed over the oceans, to which some evaporation of spray in the air should be added. Not knowing the latter it must be omitted. The moisture budget shows a residual of almost 50% of the evaporation to be transported by turbulence through cloud base. If a 500 m depth of the subcloud layer is used, the excess reduces to 25%. Thus, the mean motion picks up somewhere between 50 and 75% of the evaporated water but it does not replace turbulent transport through cloud base fully as anticipated. Unfortunately, the independent measurement of this transport cannot be offered as explained earlier; it would serve as

Table 1. *Budget of mass, heat, moisture, and kinetic energy using mean motion only*

	Inflow	Outflow	Upflow	Source	Sink	Balance
Mass						
10^4 g/s	7.2	3.6	3.6	/	/	0
Moisture				Evap.		
(10^2 g/s)	12.3	7.1	6.8	3.1	/	+1.5
(10^5 j/s)	31	18	17	8	/	+4
Heat				Q_s	Rad.	
(10^4 j/s)	0 (base)	5.8	2.9	10.0	0.7	+0.6
KE				Generation	Ground friction	
(10^4 j/s)	1.5	4.0	2.2	12.0	6.7	≈0.6

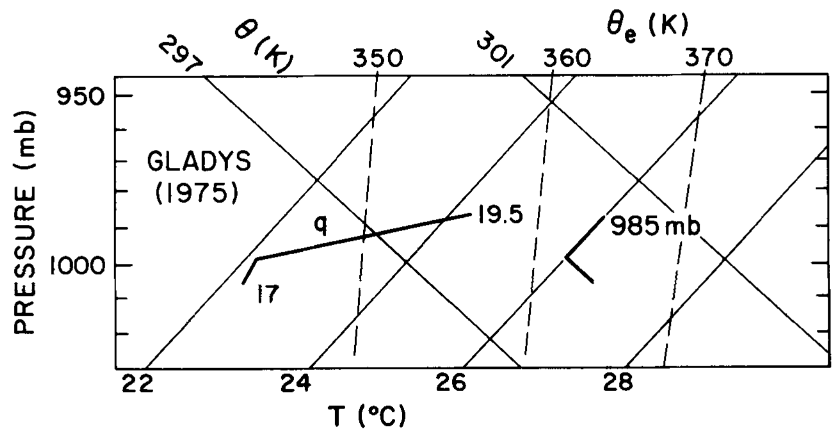

Fig. 3. Tephigram showing variation of temperature and specific humidity along the inflow trajectory of Fig. 1.

control on the evaporation calculation as well. The latter would have to be overestimated by a factor of two if the turbulent transport were to vanish entirely, an unlikely situation.

4.3. *Sensible heat*

The measure for the sensible heat budget is taken as $c_p \theta$ where θ is potential temperature. As in other calculations of tropical heat balance, a base value has been subtracted in order to avoid very large quantities of heat to be carried around. From Fig. 3, $\theta = 298.7$ K at r_1, about 300.3 K at r_2 and the estimate for the upflow is 299.5 K. For simplicity the base is taken as 298.7 K, so that the inflow becomes zero. Then outflow and upflow are as given in the third line of Table 1. The oceanic source Q_s is determined from eq. (3); hence, the Bowen ratio $Q_s/Q_e = 12.5\%$, not large. Additional heat transfer deduced by Brook (1978) due to the high specific humidity of water vapor is not included, since this heat is carried away from the subcloud layer and would have to be added in equal amounts to the transports. After allowance for radiation cooling of 1 C/d in the layer, a slight excess is left, well within the limits of computational accuracy. For subcloud layer depth of 500 m, there would be a corresponding deficit. Therewith the postulate of no turbulent heat transport through cloud base is well supported by the calculations.

4.4 *Kinetic energy*

The kinetic energy equation to be evaluated is, conventionally,

$$\rho \frac{dK}{dt} = -\mathbb{V} \cdot \nabla p + \frac{\mathbb{V} \cdot \partial \pi}{\partial z} \tag{7}$$

where $K = V^2/2$ and π is taken as $\pi_{s,z}$ only, the shearing stress in the s-direction along the vertical. After transforming the left side into a flux expression and integrating over δz, assuming steady state, the transport terms are computed as given in the last line of Table 1. As usual, these terms are very small, even in the hurricane, compared to, say, the latent heat flux; they are brought in line with the sensible heat transport through the introduction of the base into the latter.

The generation term is $-\int \mathbb{V}\cdot\nabla p\, \delta A\, \delta z$, evaluated by numerical integration over the distance s with $\mathbb{V}\cdot\nabla p = V\,\partial p/\partial s$. This term overbalances the divergence of transport greatly, so that about 50% is left for work against friction. However, no less than 50% are used for the actual increase of the kinetic energy of the moving air, just as computed for other hurricanes (Riehl, 1963). Thus, the hurricane is a highly efficient engine from this point of view, probably the most efficient one in the atmosphere.

The friction term, transformed as found in the literature, $\mathbb{V}\cdot\partial\pi/\partial z = \partial/\partial z(\mathbb{V}\cdot\pi) - \pi\cdot\partial\mathbb{V}/\partial z$. After integration with height and area one obtains $[\mathbb{V}\cdot\pi)_H - \mathbb{V}\cdot\pi)_0]A + D$, where the latter usually denotes internal dissipation. Lacking the gust probe measurements only the surface term can be evaluated with conventional methods. Thus,

$$-(\mathbb{V}\cdot\pi)_0 A = -C_d \rho \tilde{V}_0^3 A \tag{8}$$

where $\sim$ denotes area average, obtained by numerical averaging over portions of the inflow layer. The value so obtained is slightly more than half of the generation term. Balance is achieved for kinetic energy without turbulent transport through cloud base and with hardly any residual for D as internal dissipation. Of course, because of the appearance of V_0^3 in eq. (8), the computation is very sensitive to the choice of V_0. With somewhat lower values at 50 m height, assuming a small vertical wind shear above, the computed friction term would be smaller, leaving a residual for D. One cannot go farther with the present data. It appears that the calculation takes care of ocean interaction and D underneath. Assuming that ocean spray acts to transfer kinetic energy into the ocean, a mechanism usually not considered, the actual internal dissipation may indeed be small and the calculation as performed gives the total kinetic energy transport from atmosphere to ocean. It may be added that, from inspection of V^2 on short distance scales, variations are quite small, so that very large values of turbulent vertical motion would be needed for a substantial turbulent flux of kinetic energy at cloud base. If so, it would seem that calculation with Gray's method should have given an indication, at least for 20-s intervals.

In conclusion of this section, the reader may be missing a treatment of angular momentum. This, however, cannot be provided on account of the flight plan followed. Along the trajectory the momentum source term due to $\partial p/r\,\partial\theta$, where θ here is angular measurement, does not vanish and there is no way of assessing it. For a momentum budget one must be able to make a line integral calculation about the center where the pressure term vanishes. Evidently, this was not the objective of the mission.

5. Energy balance along a horizontal trajectory

In the following the energy balance will be established for an average subcloud layer unit mass which reaches the inner core; i.e., moves horizontally. For this purpose the pressure term of eq. (2) will be expanded as is familar to

$$\frac{dp}{dt} = \frac{\partial p}{\partial t} + \mathbb{V}\cdot\nabla p + w\frac{\partial p}{\partial z}$$

In a moving hurricane the local time derivative need by no means vanish, unless one chooses coordinates attached to a steady moving center. In the case of Gladys the trajectory was placed so that its initial part was through rising pressure behind, the latter part toward falling pressure ahead of the center. Thus the local pressure change cancels over trajectory. The last term vanishes since the trajectory is to be horizontal. Then $dp/dt = \mathbb{V}\cdot\nabla p$. Inserting in eq. (2) and introducting eq. (7),

$$-\frac{dF}{dz} = \rho\frac{dK}{dt} - \mathbb{V}\cdot\frac{\partial\pi}{\partial z}$$

Upon integration with height and using the approximations of the preceding section

$$F_0 = \rho\frac{dK}{dt}H + (\mathbb{V}\cdot\pi)_0 \tag{9}$$

Dividing by density and integrating with time

$$\int\frac{1}{\rho}F_0\,dt = (K_{r_2} - K_{r_1})H + \int\frac{1}{\rho} - (\mathbb{V}\cdot\pi)_0\,dt$$

21 Sept. 1938 1600 EST

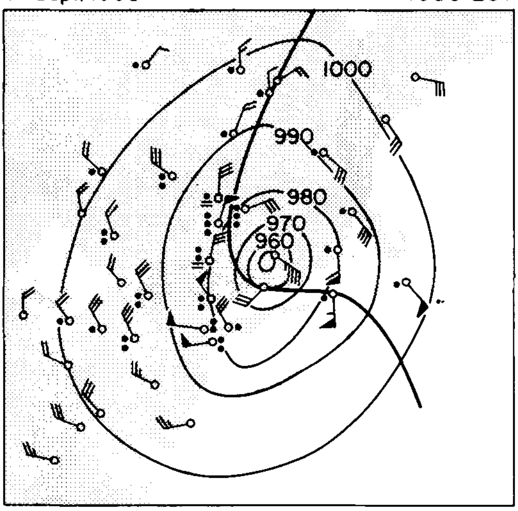

Fig. 4. Surface isobars, winds, and weather in the northeastern United States, September 21, 1938, 1600 L (data from Pierce, 1939).

Temp.-V_r 21 Sept. 1938 1600 EST

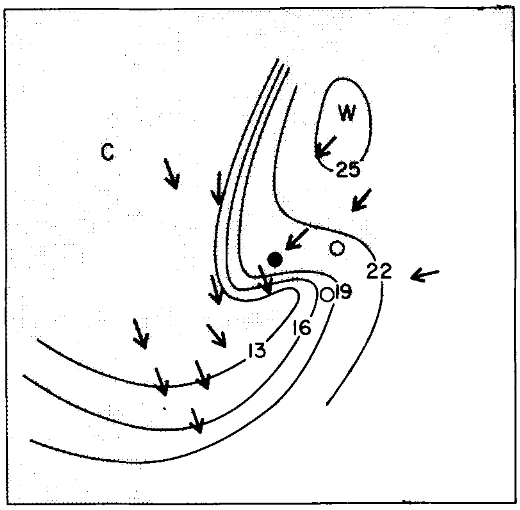

Fig. 5. Isotherms and relative wind arrows for the same period.

Inserting the turbulent exchange expressions for F_0 and π_0,

$$c_p C_d (T_w - T_a) \frac{l}{H} = K_{r_2} - K_{r_1} + C_d \tilde{V} \frac{l}{H} \tag{10}$$

where again use has been made of the relation $\int V dt = l$. Evaluating eq. (10) for particles going through the whole pressure drop, we have in j/g

Gain from oceanic heat source	1.6
Increase of kinetic energy	0.9
Frictional loss	1.0

Balance is almost achieved, certainly within the limits of accuracy of data and calculations.

Finally, solving for V^2,

$$\frac{V_2^2 - V_1^2}{\tilde{V}^2} = 2C_d \frac{l}{H} \left[\frac{c_p(T_w - T_a)}{\tilde{V}^2} - 1 \right] \tag{11}$$

It is seen that kinetic energy gain depends on the ratio $c_p(T_w - T_a)/\tilde{V}^2$. As long as $T_w - T_a$ remains constant, the capability for converting sensible heat gained from the ocean to energy of motion must decrease. It must vanish when the bracket on the right side of eq. (11) becomes zero. For $T_w - T_a = 2\,°C$, this will happen at $\tilde{V}^2_{rms} = 45$ m/s. For contrast in the strong hurricane Donna (1960), with central pressure of 960 mb, maximum wind 60 m/s, one obtains for a 4-h trajectory, starting at tangential speed of 25 m/s, which may be taken as total wind speed,

Sensible heat source	4.5 j/g
Gain of kinetic energy	1.5
Frictional loss	4.5

The sensible heat source is barely sufficient to balance the frictional loss. Nothing is left over for gain of kinetic energy unless turbulent heat transports from above is postulated, a good possibility, or unless $T_w - T_a$ is raised. In this case an increase to 3 °C would raise the sensible heat source to 6.7 units, quite sufficient for balance in hurricane Donna. A question remains just on what basis $T_w - T_a$ acquires the observed values. It is quite possible that, with increasingly strong pressure drops, the ocean cannot supply the heat to the atmosphere rapidly enough to compensate for the expansion cooling.

A similar problem was encountered in an attempt to compute a maximum rainstorm in a river basin of the Andes (Riehl and Byers, 1960). Latent heat release with an efficiency factor was taken for the left side of eq. (10). The latent heat release increases linearly with inflow whereas the friction depends on V^3. It was then found that, for the efficiency of conversion of latent into kinetic energy assumed, the maximum storm was twice as

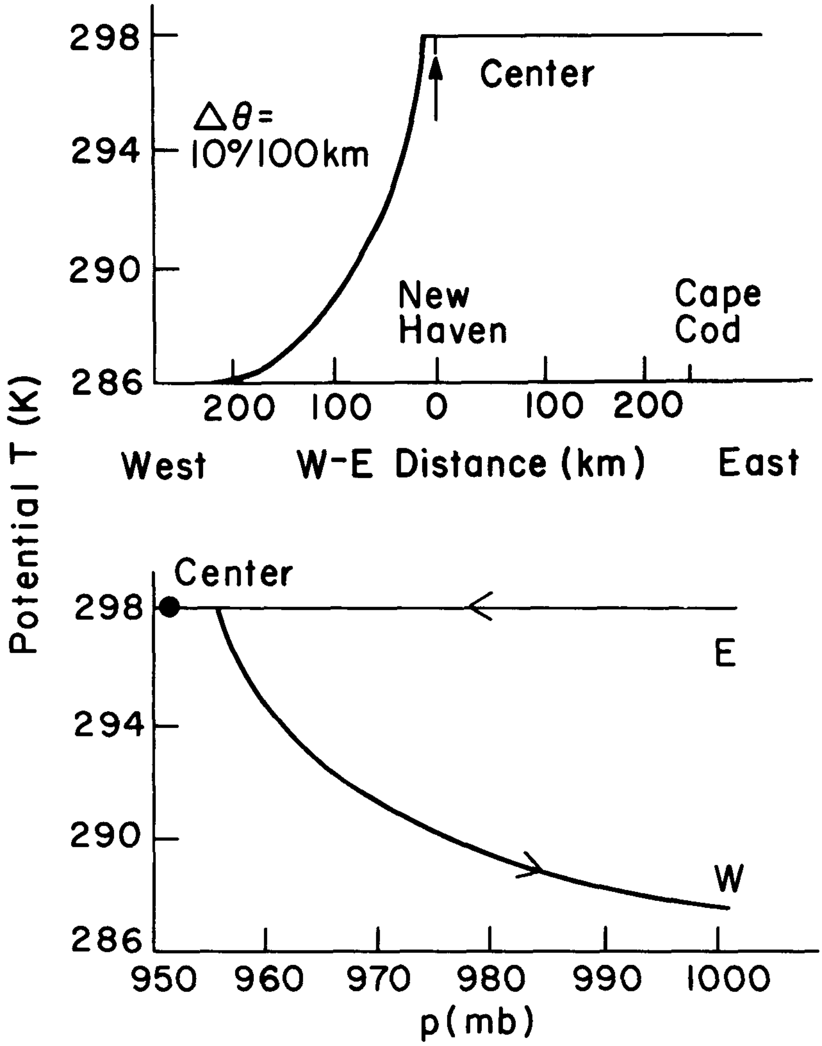

Fig. 6. Cross section of potential temperature for the same case. Upper: east-west; Lower: vs. surface pressure.

intense as the average observed one. In the present problem on hurricanes, there is an advantage in that an efficiency factor need not be assumed.

Finally, it will be of interest to glance briefly at an inflow trajectory over land, obtained from charts prepared by Pierce (1939), for the famous New England hurricane of 1938. Upon entry on the continent in Connecticut any oceanic heat source was lost, a situation shown in Fig. 4. The wind field has widened greatly from the tight tropical core. Precipitation is most abundant at the boundary to cold air over the continent which the hurricane was evading by rapid northward movement. But light rain prevails north-east of the center. There, because of the very rapid hurricane displacement (100 km/h) the relative surface winds point toward the center (Fig. 5). While the isotherms are most crowded in the frontal zone, temperature also drops more gradually along the arrows of Fig. 5. This drop corresponds closely to that of pressure, so that $\theta =$ const., shown in Fig. 6 both in horizontal cross section and against pressure. The surface energy source, of course, is zero here, but a turbulent addition to the thermal energy of the inflow is not apparent as well. Gain of kinetic energy along the trajectories is small with the weakened gradient of wind speed. Essentially, integrating along a trajectory,

$$[(c_p T + K)_2 - (c_p T + K)_1]\, \delta z = \int_{\mathrm{top}} \widetilde{w' K'}\, dt - C_d \tilde{V}_0^2 l + \int D\, dt \tag{12}$$

where, in view of a logarithmic wind profile to be expected, turbulent energy transport through the top of the layer as well as dissipation quite close to the ground cannot be neglected, since the lower boundary is now rigid. This expression should be contrasted with eq. (10). Under the circumstances, the center could not maintain itself in steady state but filled rapidly after a few devastating hours over New England.

6. Conclusion

It has been shown that the pressure drop of about 18 mb in the subcloud layer inflow toward the maximum wind can be accounted for by the first law of thermodynamics, assuming that the sensible heat source supplies the energy for isothermal expansion in the moderate hurricane Gladys. The depth of the subcloud layer is computed with considerable accuracy. Further, equivalent potential temperature increases from 348 to 356 K from beginning to end of the inflow trajectory. Using the formula for hydrostatic balance given by Malkus and Riehl (1960), namely $-dp_0 = 2.5 d\theta_e$, a closely corresponding value of pressure decrease of 20 mb is computed. Thus, the double demand is met that surface pressure can be calculated hydrostatically and by following the energy transformations in the inflow.

The oceanic moisture source must be exported in part through cloud base to higher levels. Thus, while the ratio of sensible to latent heat acquisition in the inflow is above 20%, the Bowen ratio Q_s/Q_e is only 12%, a very moderate value for hurricane inflow. The actual kinetic energy gain is produced by half of the work done by pressure forces, a very high ratio observed also in other hurricanes. The other half accounts for the frictional energy transfer to the ocean. A large turbulent kinetic energy transport downward through cloud base, which was expected, proved unnecessary and probably did not exist from inspection of the microstructure of energy variations. However, due to instrumental problems, an actual measurement which was intended cannot be supplied. It seems likely that the shearing stress at cloud base is an order of magnitude less than that on the ill-defined ocean surface containing large waves and deep penetration of spray said by observers on the flight to have encrusted the aircraft with a salt cover. The energy balance of the inflow above the subcloud layer must then be postulated as occurring largely in cloud exchange with the high troposphere, a subject discussed by Gray (1966).

Considering briefly a hurricane of greater intensity, the oceanic heat source proved insufficient to produce the requisite pressure field. However, if the moderate value of $T_w - T_a = 2\,°\mathrm{C}$ is raised to $3\,°\mathrm{C}$, balance again is achieved. Such an increase is conceivable when the pressure drop occurs so rapidly that the processes of energy transfer from the ocean are lagging so that a small adiabatic expansion of $1\,°\mathrm{C}$ takes place. Otherwise, an increasing role for turbulent heat transfer downward can achieve the same result; such reversed flow has been found in measurements of turbulent fluxes near cloud base in non-hurricane convective and in trade wind cumulus cases.

7. Acknowledgements

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ТРАНСФОРМАЦИИ ЭНЕРГИИ В ПОДОБЛАЧНОМ СЛОЕ УРАГАНА ГЛЭДИС (1975)

На основе данных наблюдений исследовательских полетов в урагане внутри облаков и под ними рассматривается вопрос о балансе энергии воздуха, проникающего к ядру урагана в подоблачном слое. Падение давления в этой модели происходит благодаря энергии, поставляемой потоком ощутимой теплоты от океана к атмосфере. Глубина подоблачного слоя, вычисленная таким образом, хорошо согласуется с наблюдаемой. Все ощутимое тепло остается в подоблачном слое, в то время как около половины испарившейся влаги должно быть перенесено турбулентным обменом к более высоким слоям. Баланс кинетической энергии в подоблачном слое достигается без существенного турбулентного переноса сверху. Однако этот результат не сохраняетсяв в ураганах большей интенсивности, когда требуется перенос сверху тепла или кинетической энергии, или когда, с другой стороны, должен усилиться приток ощутимой теплоты с поверхности океана. Баланс кинетической энергии в притекающем воздухе над подоблачным слоем должен достигаться в основном турбулентным обменом в мощных кучевых облачных башнях, проникающих до верхней тропосферы.

ADDED IN PROOF

Shortly before arrival of this proof preliminary data became available from hurricane flights off the east coast of Australia in February 1979. The aircraft was a P-3 of the National Oceanic and Atmospheric Administration (U.S.A.). Vertical motion was determined from inertial navigation equipment, a great improvement over previous instrumentation. One-second data during the flights appear suitable for a check, whether the discussion on turbulent fluxes in the paper is realistic.

On February 21 and 22 the aircraft executed a cloverleaf in a hurricane situated east of Australia at an altitude of about 500 m above the ocean, which was still below cloud base, hence comparable with Gladys. I chose six 8–10 min pieces of radial track from the cloverleaf of February 22, situated outward from the radius of strongest winds where T and T_d were almost constant. The legs gave quite variable results. Mean vertical motion ranged from -0.4 to $+1.3$ m/s; average 0.6 m/s, confirming at least the order of magnitude in Fig. 2.

All fluxes were partly up and partly down. Taking averages, sensible heat flux was downward at a mean rate of 10% of estimated surface source of 300 W/m^2; latent heat flux was upward at 20% of oceanic source near 1500 W/m^2; shearing stress nearly cancelled with downward momentum transfer on four legs and upward transfer on two legs. It appears that these new data support the deductions made from the Gladys observations quite well.

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