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SHORT CONTRIBUTION

Vertical exchange of momentum and energy in the subcloud layer of Australian Hurricane Kerry (1979)

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ABSTRACT

Vertical turbulent energy and momentum transfers gathered by aircraft below cloud base in a hurricane east of Australia are examined. Spectrum analysis shows that the humidity instrumentation was unsuitable for this purpose; only fluxes of sensible heat and of momentum could be computed. Maximum energy of the spectra is in the range of 600–1000 m wavelength. Both fluxes are directed downward, but their magnitude is only about 10% of air–sea exchanges.

Therefore the heat transfer from above does not reduce the requirement of "large" sensible heat flux from the surface. The small shearing stress was unexpected; a new model for the vertical wind structure observed in the central portion of hurricanes is proposed.

1. Introduction

A note was appended to the proof of the paper "Energy transformations in the subcloud layer of Hurricane Gladys in 1975" (Riehl and Meitin, 1979) referring to preliminary turbulent flux calculations in hurricane Kerry, off the Australian east coast on February 21–22, 1979. The present short contribution will communicate the final calculations made from the observations of the P-3 Orion aircraft of the National Oceanic and Atmospheric Administration (U.S.A.).

On February 21 and 22 the aircraft executed a cloverleaf at a radar altitude of about 500 m above the ocean, which was still about 100 m below cloud base. Perhaps this pattern is not the most meaningful for flux calculations, but it is the only available one and should give a reasonable first approximation. I chose six 8–10 min pieces of radial track on each day outward from the radius of strongest winds where T (temperature) and T_d (dewpoint) were almost constant (Fig. 1). The number of samples therefore is twelve; the average distance of each leg of 60–70 km covers the most important part of the hurricane.

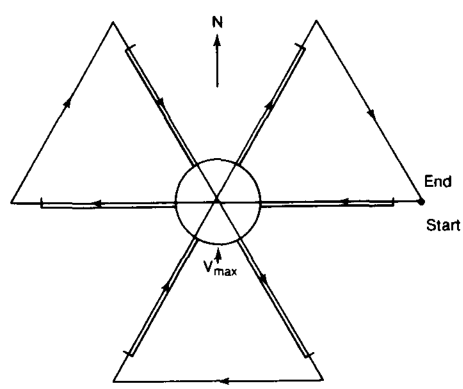

Fig. 1. Typical ideal cloverleaf flight pattern used for exploration of many hurricanes. Arrows depicting aircraft directions are drawn for Southern Hemisphere storm. Length of radial flight legs from center averages about 100 km. Length of legs used for computation in this paper (double lines) averages about 60 km. Location of "start" and "end" points is arbitrary.

2. Observations and analysis

Data were recorded at 1-s intervals or very nearly every 100 m ground distance on the radial

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legs on this low altitude part of the flight mission. The variables here to be considered are V (total wind speed), MR (mixing ratio of water vapor), and potential temperature (θ) chosen in preference to temperature to compensate for irregular up and down motions of the aircraft in the near-adiabatic subcloud layer. Vertical air motion (w) was computed from the inertial navigation system; a gust probe was not carried by the aircraft.

All quantities were detrended linearly for each sample, a procedure held justifiable in view of the small size of turbulent elements expected even for wind speed. Denoting deviations from the detrended lines by primes, the quantities $\overline{\theta' w'}$, $\overline{MR' w'}$ and $\overline{V' w'}$ were computed for each sample. Here the overbar denotes averaging over the individual sample. Power spectra also were calculated for the individual variables and for the products. Further, result were plotted in map form for the mean vertical motion and the fluxes, in order to determine asymmetries about the center. This attempt failed except for V which was 5–10 m/s stronger on the south than on the north side. The hurricane was moving eastward at the very slow rate of 2–3 m/s, so that the asymmetry in V in this case is not in agreement with the hurricane propagation vector. The other individual variables and their products were irregularly distributed about the center, so that no meaningful charts could be drawn.

3. Vertical motion

The vertical motion averaged over each of the twelve legs ranged from -0.03 to $+1.06$ m/s. Overall mean is 50 cm/s, in good agreement with previous estimates of $\overline{w}$ in the inner hurricane area. Using the mass continuity equation assuming steady state and constant density, a mean convergence of 1×10^{3} s^{-1} is obtained for the layer from the surface to flight altitude.

The spectrum showed a clear maximum of $(w')^2$ near a wavelength (L) of 900 m and a second one of equal magnitude at $L = 2.5$ km. The latter is considered inherent in the aircraft motion and not indicative of the atmosphere. No problem was introduced by the second peak for determining transports, since it was uncorrelated with the other variables.

4. Spectra

From the time analysis following the aircraft spectra are expressed in cycles/s. On the flight legs crossing the center, aircraft ground speed was computed as close to 100 m/s. Therefore, the time coordinate can be replaced by equivalent wavelength (L) of ground distance, a more meaningful quantity for this presentation. One cycle/s corresponds to 100 m ground distance, 10^{-1} cycles to 1000 m etc.

Temperature. Since the flight legs were chosen in the near-isothermal portion of the hurricane, fluctuations were small, irregular and without distributional meaning. A third day's data no doubt would have changed the whole configuration.

Wind speed. A very large peak of $(V')^2$ appears near $L = 700$ m, with steep slopes on both sides.

Mixing ratio. This was the great disappointment. A huge peak, 1–2 orders of magnitude greater than all other fluctuations, existed near $L = 1$ km. The peak was found at exactly the same place in all twelve mixing ratio spectra. According to best judgment it represents the recovery time (10 s) of the mirror of the dewpoint hygrometer. The response of this instrument additionally is regarded as too slow for water vapor transport calculations at high frequencies. Therefore all moisture calculations are invalidated, to the author's great regret.

Turbulent transports. The flux-spectra $(\theta' w')^2$ and $(V' w')^2$ show a gradual ascent from low to high frequencies. Peaks are reached near $L = 600$–800 m. Some uncertainty prevails at the highest frequencies since data were reported at 1-s intervals. Turbulent elements smaller than 100 m may play a role. In future experiments data should be taken perhaps every 10 m instead of 100 m.

Vertical heat transport

The product $c_p \rho \overline{\Theta' w'}$ is shown in Fig. 2. Here c_p is specific heat at constant pressure and ρ is air density. In Fig. 2 the values accumulate from low to high frequencies. Alternately, the value at a particular frequency or wavelength indicates the transport which occurs omitting all higher frequencies or smaller wavelengths. The heat transport is here defined as positive toward the subcloud layer, i.e., it is directed downward across the flight level near cloud base. This reversal of direction compared to the flux from the ocean surface agrees with

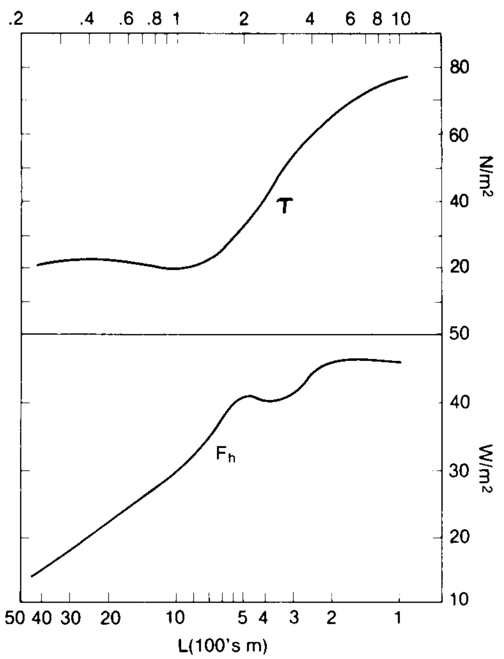

Fig. 2. Profiles of shearing stress and of vertical heat transport (defined positive toward subcloud layer) against frequency (cycles/s) and approximate wavelength composited for airplane explorations of hurricane east of Australia on February, 21 and 22, 1979.

other measurements made in convective situations. While an additional contribution at wavelengths below 100 m cannot be completely discounted, the probability is high that a downward flux F_h in the range of 45–50 W/m² is a good indicator of the upper limit of such transport which then has the order of magnitude of 10–15% of the estimated surface sensible heat source (Q_s) near 300 W/m² Clearly, such transport is within the range of error of computed Q_s, hence cannot be considered a factor of importance in generation and maintenance of hurricanes. It was a main objective of the Gladys study as well as of this investigation to determine whether downward heat flow near cloud base could reduce the requirement on Q_s substantially, since computed values of Q_s have sometimes been termed "too large". As is evident from Fig. 2, this is not the case as far as downward transport of released condensation heat is concerned.

Because of the MR problem, additions to heat transport arising from the difference of the specific heats of air and water vapor (Brook, 1978) could not be computed.

5. Vertical momentum transport

The shearing stress $\tau = -\rho\overline{V'w'}$ is seen in the upper part of Fig. 2. Here also, the transport is downward; it is mainly due to eddies with $L < 1$ km; and total stress may attain 8 dyn/cm² if eddies smaller than 100 m could have been included; these are very noticeable during many flights as a "washboard effect". As in the case of F_h the magnitude of τ is about 15% of the computed surface stress. Nearly all of the momentum abstracted from atmosphere to ocean is taken from the main inflow layer.

This is a surprising result. Since the wind, especially the dominant tangential component v_θ ($V \approx v_\theta$) is constant with height to 500–400 mb, very far above the inflow layer, a large downward transport from the middle troposphere had been expected. As suggested by Ooyama (1969) the deep tropospheric layer not participating in the inflow must contract inward somewhat in order for the high velocities observed in the layer to occur. If not hindered by friction at the surface or in the free atmosphere, the contraction occurs at constant angular momentum (Ω), closely given at small radii without the Coriolis term by

$$v_\theta r = \text{const.} \tag{1}$$

where r is radius positive outward from the center. In the inflow layer, because of the momentum abstraction to the ocean, the relation

$$v_\theta r^{0.5} = \text{const.} \tag{2}$$

holds closely (Riehl, 1963). Since the wind does not increase with height (cold-core type) the co-existence of relations (1) and (2) is not compatible. Since we have seen that large momentum abstraction downward does not occur from the middle troposphere, one may fall back on turbulent exchange of momentum with the high troposphere. However, another possibility presents itself at least as partial explanation of the observed constant v_θ profiles with height and vertical eye walls often extending to the high troposphere.

The thermal wind is closely given by the cyclostrophic relation

$$r^{-1}\,\partial v_\theta^2/\partial z = gT_m^{-1}\,\partial T_m/\partial r \tag{3}$$

where T_m is the mean virtual temperature over the vertical distance ∂z chosen. Since the hurricane is warm-cored above the isothermal inflow layer, it follows from eqn. (3) that the radial pressure gradient decreases upward. Therefore an outward radial wind component v_r should appear almost from cloud base up; correspondingly v_θ should decrease upward and the eyewall should start leaning outward. Thus the effect of the inward moving mass in the middle troposphere at constant Ω is opposed by the thermal wind, one tending to increase, the other to decrease, v_θ with height. We shall now calculate whether a reasonable balance can be achieved between these conflicting influences with a simple example.

Consider a boundary inflow layer with $v_\theta = 30$ m/s at $r = 100$ km. Upon reaching $r = 60$ km, $v_\theta =$ 40 m/s from eqn. (2), and this is the wind speed actually observed through the greater depth of the troposphere. Above this boundary layer a slow contraction may be taking place, say, to 7 km, so that, at the 60-km radius, a linear increase of v_θ with height would be observed to $v_\theta = 50$ m/s if the inflow takes place following eqn. (1). We now inquire, using eqn. (3), what radial temperature gradient is necessary to counteract the increase from 40 m/s at the top of the boundary layer, roughly 1 km height, to 50 m/s at 7 km. For this purpose $r = 80$ km in eqn. (3), $\delta r = 40$ km, $T_m =$ 270 K and $\delta z = 6$ km. For a linear increase of v_θ from 40 to 50 m/s, the increase of v_θ^2 is also almost linear, so that the temperature difference is constant in the volume. Its value is -2 °C, which is the magnitude observed in the warm-core cyclone outside the eye. If δz is raised to 8 or 10 km, the requirement on the temperature gradient is reduced, whereas this gradient in actuality increases upward. Here then the cyclostrophic term becomes dominant, the radial wind component turns outward and v_θ begins to decrease with height. With this simple model a complex dependence on turbulent exchanges is avoided. The expected large downward momentum transfer is not observed. Instead we obtain an equilibrium between inward acceleration toward lower pressure at constant momentum and a temperature gradient which reduces the inward velocity increase through flattening of the isobaric surfaces.

6. Conclusion

The analysis of the hurricane Kerry fluxes has brought some disappointment, especially with regard to latent heat fluxes which could not be measured. Maps of the vertical fluxes could not be drawn because of irregular distributions. However, it is shown that the eddy sizes for vertical transport of heat and momentum are small, with L mainly in the range of 100–2000 m. Both fluxes are directed downward, but their magnitude is rather insignificant compared to the interface transfers. Thus downward heat transport does not relieve the requirement of "large" Q_s at the ocean surface. The fact that downward momentum transport is also small, leads to the suggestion of a combined dynamic-thermal steady-state model to explain the observed constant tangential wind speed with height in hurricanes to great altitudes. With these results and residual problems in mind it is clear that the research airplane can remain a first-rate tool for exploration of the principal questions still associated with hurricanes.

7. Acknowledgements

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REFERENCES

Brook, R. R. 1978. The influence of water vapor fluctuations on turbulent fluxes. *Bound. Layer Meteor.* **15**, 481–489.

Ooyama, K. 1969. Numerical simulation of the life cycle of tropical cyclones. *J. Atmos. Sci. 26*, 3–40.

Riehl, H. 1963. Some relations between wind and thermal structure of steady-state hurricanes. *J. Atmos. Sci. 20*, 276–287.

Riehl, H. and Meitin, J. 1979. Energy transformations in the subcloud layer of hurricane Gladys (1975). *Tellus 31*, 505–514.