NOTES AND CORRESPONDENCE

Viscous Dissipation of Turbulence Kinetic Energy in Storms*

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ABSTRACT

In this note the magnitude of the viscous dissipation of turbulence kinetic energy in the surface layer of storms is investigated. It is shown that the layer-integrated dissipative heating is a cubic function of the wind speed. The magnitude of the estimated heating at higher wind speeds confirms the importance to storm evolution of this term in the turbulence kinetic energy equation and suggests that dissipative energy should be included in numerical weather prediction models, particularly in models that resolve mesoscale structures in storms. A general discussion of the implications of the results for the energetics of a range of storm systems is provided.

1. Introduction

In a notable development in our understanding of the energetics of hurricanes, Bister and Emanuel (1998) recognized that viscous dissipation of turbulence kinetic energy (TKE) can be a significant source of heat in hurricanes. They also showed that this source of heat increases the efficiency of the hurricane.[1] This work stimulated us to look at the details of the dissipation of TKE near the surface.

Part of the stress in the atmosphere goes into wave production. This energy is not available for frictional dissipation. The remaining kinetic energy is ultimately dissipated on molecular scales, with the dissipation largely taking place in the surface layer. The magnitude of this dissipation can be estimated with reference to the TKE equation. Under steady state and horizontal uniformity, the TKE equation can be expressed by (e.g., Kraus and Businger 1994, p. 148)

$$-\overline{uw}\frac{\partial U}{\partial z} + \frac{g}{\Theta_v}\overline{w\theta_v} - \frac{\partial}{\partial z}\left(\frac{\overline{wp}}{\rho} + \frac{\overline{wu_i^2}}{2}\right) - \varepsilon = 0. \quad (1)$$

The convection used here is to write the average components over time in uppercase and the fluctuating components in lowercase. Thus, U is the average wind velocity and u, υ, and w are the fluctuating components, Θ_v and θ_v are average and fluctuating virtual potential temperature. P is pressure, ρ is density, z is height, g is acceleration of gravity, and ε is the viscous dissipation of TKE. The first term of Eq. (1) represents shear production, the second term represents buoyant production, and the third term contains a pressure work term and a turbulent transport term.

Because the shear production term is very large under high wind conditions, near-neutral conditions exist in the surface layer. Under these conditions, the TKE equation [Eq. (1)] reduces to a simple balance between the shear production term and the viscous dissipation at the high end of the turbulence spectrum (Wyngaard and Coté 1971), and can be written

$$-\overline{uw}\frac{\partial U}{\partial z} = \varepsilon. \quad (2)$$

Because $-\overline{uw} = \tau/\rho = u_*^2$ and $\partial U/\partial z = u_*/(kz)$, upon integration we have

$$U = \frac{u_*}{k}\ln\frac{z}{z_0}, \quad (3)$$

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[1] The added heat and/or moisture resulting from the dissipative heating leads hydrostatically to a lower sea level pressure in the area of enhanced winds.

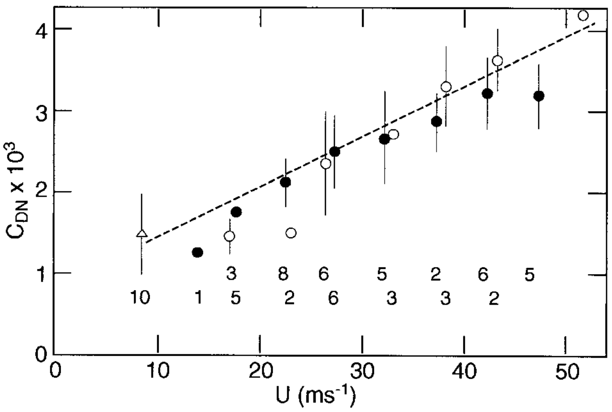

FIG. 1. Mean values of the neutral drag coefficient as a function of wind speed at a 10-m height based on hurricane studies (○), wind flux experiments (●), and vorticity/mass budget analysis (△). Vertical bars refer to std dev of individual data for each mean, with the number of data points used in each 5 m s^{-1} shown above the abscissa. The dashed line refers to Charnock's relation with $a = 0.014$ [after Garratt (1977)].

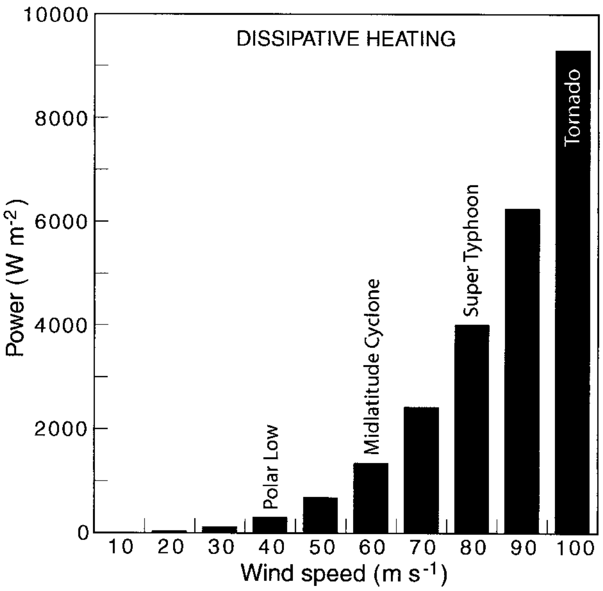

FIG. 2. Histogram of dissipative heating as a function of wind speed. Bar labels indicate highest surface wind speeds attained for each kind of storm.

where u_* is the friction velocity, k is the von Kármán constant, z_0 is the roughness length, and z is the height of wind observations, commonly referenced to about 19.5 m in ship observations. Equation (2) may be rewritten

$$\frac{u_*^3}{kz} = \varepsilon. \tag{4}$$

Per unit volume, the power generated is $\rho u_*^3/kz$. Over the depth of the surface layer z_1, the power produced per unit area can be expressed by

$$\frac{\rho u_*^3}{k}\ln\left(\frac{z_1}{z_0}\right) = \rho\overline{\varepsilon}z_1, \tag{5}$$

where $\overline{\varepsilon}$ is the average over the surface layer and $z_1 \gg z_0$. The roughness length is not well known under high wind conditions. However, indirect evidence (Garratt 1977) shows that the Charnock relationship (1955) remains valid for winds up to 50 m s^{-1} (Fig. 1). The Charnock relationship can be expressed by

$$z_0 = \frac{au_*^2}{g}, \tag{6}$$

where a is a constant of about 0.014.

The friction velocity is related to the wind speed U by $u_*^2 = C_D U^2$, where C_D is the drag coefficient. Upon substitution in Eq. (5), the dissipation of TKE in the surface layer becomes

$$\rho\overline{\varepsilon}z_1 = \frac{\rho(C_D U^2)^{3/2}}{k}\ln\left(\frac{z_1}{z_0}\right). \tag{7}$$

Equation (7) gives an estimate of the magnitude of the layer-integrated dissipation as a function of U. A graph of the dissipation calculated over a range of wind speeds illustrates the cubic dependence of the dissipative heating on the wind strength (Fig. 2).

2. Discussion and conclusions

In this note, we investigate the magnitude of the viscous dissipation of TKE in the surface layer of storms. A lack of observational data on sea surface exchange coefficients of momentum at high wind speeds necessitates a number of assumptions in the analysis and therefore leaves room for significant errors in the estimates presented in Fig. 2.

The Charnock relationship works well for quasi-steady-state conditions in which there is a balance between the wind stress and a mature wave state. However, this balance is rarely met in the rapidly changing boundary layer of a moving cyclone. As a consequence, more kinetic energy is transferred to the waves and wave production, and this energy is not available for dissipative heating. The approach taken here, to extrapolate the Charnock relationship to higher wind speeds, seems justified because it likely underestimates the actual values of u_* and C_D, thus at least partially compensating for the loss of energy to wave production under these conditions.

To obtain direct measurements of the stress over the ocean in high winds is a major challenge to the boundary layer community. Sea spray prevents us from using the dissipation technique and probably also prevents the use of sonic anemometers, although it is possible that the noise from the spray may be filtered out. The use of pressure transducers may be more promising, because under high-wind conditions the signal is strong. There also is the challenge of setting up an observation platform in a hostile environment.

To obtain the layer-integrated dissipation, Eq. (2) was integrated from the sea surface to 19.5 m, the height of reference ship observations. This height is likely an underestimate of the depth of the surface layer. An increase of the surface-layer depth from 19.5 to 100 m increases the energy estimate by 18% at 50 m s^{-1}, indicating that the dissipative heating is not overly sensitive to the prescription of the surface-layer depth. In sum, the estimates of the dissipative heating provided in this paper may reasonably be considered conservative estimates of the actual values.

Implications and speculation

Figure 2 illustrates the fact that the dissipative heating is a cubic function of the wind speed. At lower wind speeds typical of conditions for which direct observations of the friction velocity are available, the dissipative heating is negligibly small. This fact helps to explain the lack of attention paid historically to this term in the TKE equation and the absence until recently of dissipative heating in mesoscale numerical weather prediction models (Zhang and Altshuler 1999). Moreover, the signature of dissipative heating has been masked in part by the increased uncertainty in the magnitude of the surface fluxes at high wind speeds. The rapid increase in the dissipative heating with increasing wind speed may raise the question as to whether a positive feedback cycle is possible. However, the loss of kinetic energy through friction will always have a larger limiting role in the storm dynamics than will the mitigating influence of the dissipative heating. Instead, the heating in the surface layer should be viewed as an ameliorating effect that reduces slightly the large energy sink to the storm system represented by frictional dissipation of kinetic energy.

In Fig. 2 it is seen that the heating becomes a significant fraction of the solar constant at wind speeds greater than 40 m s^{-1}. Winds of this magnitude and higher produce a considerable amount of sea spray in the surface layer and are often accompanied by precipitation. As a consequence, the dissipative heating may be largely converted to latent heat through evaporation of ubiquitous spray (or raindrops) under high-wind conditions. The details of the transfer of energy within the surface layer and to and from the ocean surface under these conditions are complex and remain the subject of ongoing research (e.g., Fairall et al. 1994; Andreas 1995).

The highest winds in storm systems tend to be confined to mesoscale regions (e.g., eyewall or low-level jet). Nevertheless, wind speeds in the range of 40–50 m s^{-1} have been observed in strong midlatitude cyclones, with winds of 30–40 m s^{-1} extending over relatively long ocean fetches (e.g., NOAA 1991; Neiman et al. 1993). In these cases, the dissipative heating is significant in sum, and its impact on storm evolution needs to be considered. In general, the impact of dissipative heating on the intensity and depth of stronger midlatitude cyclones and other baroclinic systems[2] will depend on the sector of the storm in which the higher winds are occurring. For example, if the heating takes place primarily in the warm sector, an increase in the efficiency of the larger cyclone could result. Enhanced surface winds typically occur in a relatively narrow swath ahead (to the east) of the surface trough. This flow is often associated with a low-level jet that is sometimes referred to as the warm conveyor belt (Browning and Pardoe 1973). Dissipative heating in this airstream is advected toward the center of the low where the added latent-heat release would contribute hydrostatically to a reduction in the central surface pressure of the low. In contrast, if the highest winds occur on the cold side of the storm, differential frictional heating will act to reduce baroclinity in the system. However, a decrease of the static stability in the cold air will tend to increase the depth of convective clouds and may potentially have an impact on the strength of postfrontal rainbands.

Bister and Emanuel (1998) point out that because the determining factor for hurricane intensity potential is the enthalpy of the air, hurricane intensity potential is not sensitive to how the dissipative heating is partitioned into latent and sensible contributions. However, this point does not preclude the convective structure of a hurricane to be impacted by the details of the energy partition. Using two independent theoretical derivations, Bister and Emanuel show that inclusion of dissipative heating in the boundary layer increases the maximum wind speed by 20%. In a study using a full primitive equation model, Zhang and Altshuler (1999) attribute a 10% increase in the simulated wind strength of Hurricane Andrew to the inclusion of the dissipative heating term in the model's boundary layer parameterization.

The centers of some midlatitude cyclones and polar lows show evidence of a warm core and axisymmetric structure (Bosart 1981; Rasmussen and Zick 1987; Businger and Baik 1991; Albright et al. 1995). In these systems the resulting dissipative heating would contribute directly to the moist-static energy available in the core and thus would improve the efficiency of the core circulation in a way analogous to that predicted in their tropical cousins. Measurements in strong polar lows show latent and sensible heat fluxes of about 500 W m^{-2} each, which is the same order of magnitude as the dissipative heating estimate for wind speeds of about 40–50 m s^{-1} observed in these storms (Shapiro et al. 1987).

In a study by Black and Holland (1995), the energy budget of the marine boundary layer in Tropical Cyclone Kerry was investigated. The authors find that a down-

[2] Some polar lows (Harrold and Browning 1969) and subtropical storm systems (Businger et al. 1998) display baroclinic characteristics.

ward energy flux of 100 W m^{-2} was required to balance the budget in the area of strongest winds (see their Fig. 19). The maximum wind speeds analyzed for that time were about 40 m s^{-1}. At that wind speed, the dissipative heating is estimated to be about 290 W m^{-2}. An areal average wind speed might be approximately, 30 m s^{-1}, giving a dissipative heating of about 100 W m^{-2}, which is clearly sufficient to account for the budget deficit.

The large magnitude of the dissipative heating at higher wind speeds seen in Fig. 2 raises the possibility of measuring its impact on the energy budget of the inflow layer of hurricanes indirectly. As noted previously, very challenging conditions exist near the surface in the core of mature tropical cyclones. Nevertheless, the significance of dissipative heating in tropical storm systems needs to be investigated both observationally, for example, in a coordinated Lagrangian field experiment[3] (Johnson et al. 2000), and numerically. The impact of the heating in storm systems will likely depend on storm structures, stage of development, and the duration of model simulations. Once empirical data on the magnitude and partition of the dissipative heating are available, accurate representation of this term in the parameterization of energy exchange at the surface should improve the performance of numerical models that can resolve mesoscale structure in storms (e.g., Zhang and Altshuler 1999).

Last, Doppler radar observations reveal wind speeds in the strongest tornadoes that reach 150 m s^{-1} (Wurman et al. 1996; Bluestein et al. 1997). Estimating the magnitude of the dissipative heating at these velocities is beyond the scope of this paper. However, if the heating follows the trend suggested by Eq. (7) it will contribute significant heat locally to the base of a tornado and may again provide an opportunity for investigation.

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[3] Field experiments that will include deployment of a balloon instrument platform that is hurricane-hardened are planned for the next several Atlantic hurricane seasons.

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