Sensible Heat Fluxes in the Nearly Neutral Boundary Layer: The Impact of Frictional Heating within the Surface Layer

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ABSTRACT

The effect of frictional dissipative heating on the calculation of surface fluxes in the atmospheric boundary layer using bulk flux formulas is considered. Although the importance of frictional dissipation in intense storms has been widely recognized, it is suggested here that its impact is also to be seen at more moderate wind speeds in apparently enhanced heat transfer coefficients and countergradient fluxes in nearly neutral conditions. A simple modification to the bulk flux formula can be made to account for its impact within the surface layer. This modification is consistent with an interpretation of the surface layer as one across which the flux of total energy is constant. The effect of this modification on tropical cyclones is assessed in an idealized model, where it is shown to reduce the predicted maximum wind speed by about 4%. In numerical simulations of three individual storms, the impacts are more subtle but indicate a reduction of the sensible heat flux into the storm and a cooling of the surface layer.

1. Introduction

Sensible and latent heat fluxes at the land or sea surface are usually calculated using bulk flux formulas derived from surface layer similarity theory (Garratt 1992) that yields expressions for the requisite exchange coefficients. However, published measurements of sensible heat fluxes in the nearly neutral regime at high wind speeds are not completely in accord with the predicted values of the exchange coefficient for heat C_H. Smedman et al. (2007b) report enhanced sensible and latent heat fluxes in slightly unstable conditions over the sea when the wind speed is high and suggest that these enhanced fluxes are due to the presence of roll vortices in very nearly neutral conditions (Smedman et al. 2007a). They propose that a distinct regime of thermodynamic stability should be recognized to encompass such cases, for which they propose the term ''unstable very close to neutral'' (UVCN). Subsequent modeling studies (Rutgersson et al. 2007; Sahlée et al. 2008) have suggested that these effects may be important in storms and, on annual time scales, in the surface flux budget of regions of consistently high wind speeds, such as the Southern Ocean.

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In a separate study, Mahrt et al. (2012) review measurements of the sensible heat flux in nearly neutral conditions over the sea and note that upward fluxes have often been measured when the surface is slightly colder than the air. Such results are frequently dismissed as a result of experimental error in situations where the sensible heat flux and temperature difference are small. Mahrt et al. (2012) suggest a number of possible explanations for such countergradient fluxes and show that the measurements indicate that the effect becomes more pronounced at higher wind speeds.

Mahrt et al. (2012) do not include frictional dissipation among their possible explanations for such countergradient heat fluxes, but it will be argued below that frictional dissipation in the nearly neutral surface layer should be considered as a significant contributor to apparently enhanced heat transfer coefficients and countergradient fluxes (though not necessarily to the exclusion of other mechanisms.) Accounting for frictional dissipation within the surface layer requires a small but simple modification of the bulk flux formula.

The importance of frictional dissipation in tropical cyclones (TCs) has been emphasized by Bister and Emanuel (1998), who conclude, by modifying the axisymmetric model of Emanuel (1986), that the maximum wind speed in a TC is increased by a factor of about $\sqrt{3/2}$ if frictional dissipative heating is included in the model. It is argued

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below that the adjustment to the bulk flux formula should also be included in this model and that it acts to reduce the maximum wind speed in this idealized model.

The aim of this paper is to explore the impact of frictional dissipation in the surface layer insofar as it affects surface exchange. In section 2 a basic version of the theory in nearly neutral boundary layers is discussed: this is sufficient to explain the occurrence of apparently enhanced transfer coefficients and countergradient fluxes in these conditions. It is shown that a small modification should be made to the bulk flux formula for the sensible heat flux. In section 3 the theory is presented from the more general perspective of conservation of total energy, while its relationship to surface layer similarity theory is considered in section 4. Implications for the modeling of TCs are discussed in section 5. First, the axisymmetric model of Emanuel (1986) is considered and a slightly reduced estimate of the maximum wind speed is derived. This resolves the paradox about the role of frictional dissipation in TCs noted by Kieu (2015). Numerical simulations of three intense TCs are then performed to assess the impact in less idealized situations. In section 6 the climatological impact is assessed. Finally, some conclusions are drawn.

2. Basic theory in nearly neutral conditions

Although a discussion of the surface layer and the role of frictional dissipative heating based on the concept of total energy will be presented later, it is instructive to begin with a simpler analysis of the role of dissipative heating in the nearly neutral boundary layer, where the impact of atmospheric stability on turbulent diffusion is neglected.

The surface layer is defined as the lower portion of the atmospheric boundary layer, across which the mean upward fluxes of momentum $-\overline{\tau}$, heat $\overline{H}$, and moisture $\overline{Q}$ do not vary substantially from their values at the surface. Here, overbars denote mean quantities. In the region of high shear close to the surface, the rate of production of turbulent kinetic energy (TKE) from the mean flow is large and the budget of TKE is dominantly a balance between shear production and dissipation (Businger and Businger 2001). However, see Smedman et al. (2007a) for some observations suggesting that deviations from this balance might be significant. Higher above the surface, on a scale known as the Obukhov length, commonly denoted by L, buoyancy becomes a significant component of the budget of TKE and modifies the characteristics of the turbulence. The surface layer is traditionally modeled using a first-order closure expressed in terms of Monin–Obukhov similarity theory (Garratt 1992).

Initially, we consider the nearly neutral limit, where the effect of buoyancy is weak and L is large. The TKE that is generated from the mean flow by shear production is dissipated locally, constituting a source of heat within the surface layer. In a steady state, by conservation of energy, this must be balanced by a divergence of the heat flux $\overline{H}$:

$$\frac{\partial \overline{H}}{\partial z} = \rho \overline{\varepsilon}(z), \tag{1}$$

where ρ is the density and $\overline{\varepsilon}(z)$ is the mean rate of dissipation of TKE, customarily defined as a quantity per unit mass. Following standard practice in surface layer theory, ρ is taken as uniform except insofar as it affects the buoyancy. Within the surface layer the upward flux of momentum remains close to its surface value, $-\rho u_*^2$, where $u_* = \sqrt{-|\overline{\tau}|/\rho}$ is the friction velocity. The integrated production of TKE up to a height z is $\rho \overline{u}_*^2 u(z)$, where $\overline{u}(z)$ is the mean velocity. Assuming a logarithmic profile, $\overline{u}(z) = (u_*/k)\ln(z/z_{0m})$, where z_{0m} is the roughness length for momentum and k is the von Kármán's constant, the total energy extracted from the mean flow up to a height z is $\rho(u_*^3/k)\ln(z/z_{0m})$. On the assumption of a local balance between production and dissipation, this energy must be dissipated locally and converted into heat. By conservation of energy for the layer of the atmosphere up to z, it follows that

$$\overline{H}(z) = \overline{H}(0) + \rho \frac{u_*^3}{k} \ln\left(\frac{z}{z_{0m}}\right). \tag{2}$$

Equivalently, this equation may be derived by integrating the standard expression for dissipation,

$$\overline{\varepsilon} = \frac{u_*^3}{kz}, \tag{3}$$

between z_{0m} and z.

We assume that conditions are sufficiently close to neutral that that the turbulent diffusivity for heat is unaffected by atmospheric stability and so may be set equal to $u_* kz/\sigma_N$, where σ_N is the neutral turbulent Prandtl number. Hence,

$$-\rho c_p \frac{u_* kz}{\sigma_N} \frac{\partial \overline{\theta}}{\partial z} = H(0) + \rho \frac{u_*^3}{k} \ln\left(\frac{z}{z_{0m}}\right), \tag{4}$$

where θ is the potential temperature. This equation may be integrated to give

$$\begin{aligned} \overline{\theta}(z) - \overline{\theta}(z_{0m}) &= \sigma_N \frac{\theta_{*0}}{k} \ln\left(\frac{z}{z_{0m}}\right) - \frac{\sigma_N u_*^2}{2c_p k^2}\left[\ln\left(\frac{z}{z_{0m}}\right)\right]^2 \\ &= \sigma_N \frac{\theta_{*0}}{k} \ln\left(\frac{z}{z_{0m}}\right) - \frac{\sigma_N \overline{u}^2(z)}{2c_p}, \end{aligned} \tag{5}$$

where $\theta_{*0} = -\overline{H}(0)/(\rho c_p u_*)$ is the turbulent scale for temperature defined from the surface heat flux $\overline{H}(0)$ and

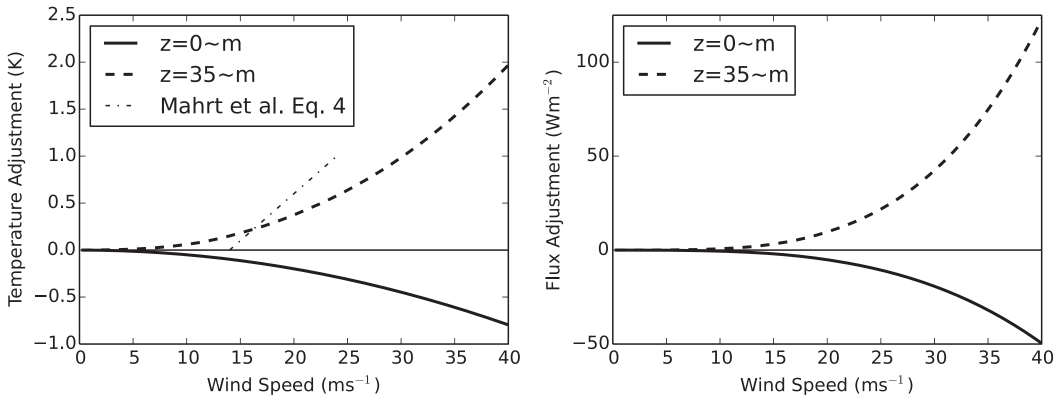

FIG. 1. The implied offsets to the surface temperature required to calculate (left) the true fluxes at $z = 0$ and 35 m and (right) the corresponding adjustments to the fluxes.

the assumption of a logarithmic velocity profile has again been invoked. The surface skin temperature $\overline{\theta}_s$ is normally related to the aerodynamic surface temperature $\theta(z_{0m})$ using an inverse Stanton number B^{-1}:

$$\frac{\overline{\theta}(z_{0m}) - \overline{\theta}_s}{\theta_{*0}} = B^{-1}, \tag{6}$$

which may be expressed in terms of a thermal roughness length z_{0h} by setting $B^{-1} = (\sigma_N/k)\ln(z_{0m}/z_{0h})$. Hence,

$$\overline{\theta}(z) - \overline{\theta}_s = \sigma_N \frac{\theta_{*0}}{k} \ln\left(\frac{z}{z_{0h}}\right) - \frac{\sigma_N \overline{u}^2(z)}{2c_p}. \tag{7}$$

The surface heat flux is therefore

$$\overline{H}(0) = \frac{\rho c_p}{\sigma_N} \frac{k^2}{\ln\left(\frac{z}{z_{0m}}\right)\ln\left(\frac{z}{z_{0h}}\right)} \overline{u}(z)\left[\overline{\theta}_s - \frac{\sigma_N \overline{u}(z)^2}{2c_p} - \overline{\theta}(z)\right]. \tag{8}$$

Similarly, from Eqs. (2) and (8), the heat flux at a height z becomes

$$\overline{H}(z) = \frac{\rho c_p}{\sigma_N} \frac{k^2}{\ln\left(\frac{z}{z_{0m}}\right)\ln\left(\frac{z}{z_{0h}}\right)} \overline{u}(z) \times \left\{\overline{\theta}_s + \frac{\sigma_N \overline{u}(z)^2}{c_p}\left[\frac{\ln\left(\frac{z}{z_{0h}}\right)}{\ln\left(\frac{z}{z_{0m}}\right)} - \frac{1}{2}\right] - \overline{\theta}(z)\right\}. \tag{9}$$

Thus, the actual surface heat flux at $z = 0$ will be more negative than that predicted by a standard bulk flux formula, where the flux is proportional only to $\overline{\theta}_s - \overline{\theta}(z)$. Conversely, for realistic values of the roughness lengths, the flux at the height of observation will be more positive than would be predicted from a standard bulk flux formula, and countergradient fluxes may be observed at this height. To obtain the actual surface flux from the bulk flux formula, an adjusted surface temperature should be used, as implied by Eq. (8). Similarly, to obtain the actual flux at the height of observation, a different adjustment is required, as implied by Eq. (9).

Before these formulas may be applied, the neutral turbulent Prandtl number must be specified. Several different values are in common use. A value of 0.74 was proposed by Businger et al. (1971), while Dyer and Bradley (1982) proposed a value of 1.0 and Högström (1988) concluded that $\sigma_N \approx 0.95$. If a value of 1.0 is adopted, the speed-dependent term in Eq. (8) becomes the kinetic energy density of the mean flow and this seems theoretically attractive, since then the flux is related to the difference in total energy, which must be conserved during fluid motion. This motivates the following consideration of total energy, but we shall find that the analysis does not actually constrain the turbulent Prandtl number to be unity.

The left panel of Fig. 1 shows these adjustments to the surface temperature [i.e., the middle terms within the brackets in Eqs. (8) and (9)] to obtain the fluxes at $z = 0$ m (solid line) and at $z = 35$ m (dashed line) as a function of wind speed. A height of 35 m has been chosen because Mahrt et al. (2012) used measurements made between 30 and 40 m in their analysis. Roughness lengths have been calculated using the COARE3.0 algorithm (Fairall et al. 2003) and the turbulent Prandtl number has been set to 1. These adjustments are small for wind speeds less than 10 m s^{-1}, but become more significant at higher wind speeds. As an illustrative device, Mahrt et al. (2012) showed that an adjustment of the surface temperature of the form

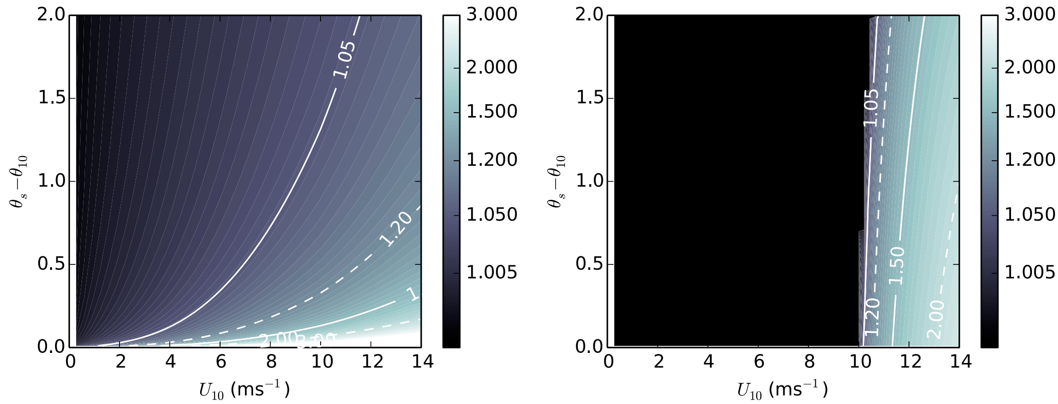

FIG. 2. The enhancement of the exchange coefficient for heat C_H at 10 m as a function of wind speed and temperature difference, calculated using (left) the present theory and (right) the fit to observational data proposed by Rutgersson et al. (2007).

$$\overline{\theta}_s \leftarrow \overline{\theta}_s + 0.1(\overline{u} - 14)\mathscr{H}(\overline{u} - 14), \tag{10}$$

where $\mathscr{H}$ is the Heaviside function, caused their experimental data, measured between 30 and 40 m, to fall more nearly along a single curve. For comparison, their illustrative adjustment is shown as the dotted line in Fig. 1 for wind speeds between 14 and 24 m s^{-1}, roughly matching the range of wind speeds they considered. This adjustment is broadly consistent with that derived above, although larger in magnitude above 16 m s^{-1}. The associated adjustments to the fluxes are shown in the right panel. These are very small below 15 m s^{-1}, when the adjustment to the flux at 35 m is only 3 W m^{-2}, but increase cubically with wind speed, disregarding the slight variation in the exchange coefficient with the wind speed.

Alternatively, we may define an apparent exchange coefficient C'_H that gives the actual sensible heat flux at the height of observation, if used in conjunction with the actual difference in temperature between the air and the surface. Immediately,

$$C'_H = C_H\left\{1 + \frac{\overline{u}^2(z)}{c_p[\overline{\theta}_s - \overline{\theta}(z)]}\left[\frac{\ln(z/z_{0h})}{\ln(z/z_{0m})} - \frac{1}{2}\right]\right\}. \tag{11}$$

In general $C'_H > C_H$, showing that there is an apparent enhancement of the transfer coefficient for realistic heights of observation and roughness lengths. We may term the ratio C'_H/C_H an enhancement factor. Likewise, an enhancement factor for the UVCN regime of Rutgersson et al. (2007) may be derived from their Eqs. (8) and (10). Contour plots of the two enhancement factors are shown in Fig. 2 for the weakly unstable regime with $\overline{\theta}_s - \overline{\theta}(10\,\mathrm{m}) > 0.1$ K. Peak values of the enhancement factors, at the highest wind speeds and least unstable conditions are similar, although the formulas of Rutgersson et al. (2007) show greater enhancement for stronger temperature contrasts.[1] It is important to note that the enhancement factor derived above is a theoretically based one for a single process, whereas the enhancement factor of Rutgersson et al. (2007) was derived observationally and may be influenced by several processes. Nevertheless, we suggest that some of the enhancement observed by Rutgersson et al. (2007) should be explained by frictional dissipative heating. Conversely, it should be noted that dissipative heating provides no mechanism to enhancement of the latent heat flux that is also reported by Rutgersson et al. (2007).

3. The surface layer from the perspective of total energy

As already noted, the foregoing results suggest that it might be useful to consider the surface layer in terms of total energy. This seems a natural approach within the framework of a diffusive closure that should be formulated in terms of transport equations for conserved quantities, where any representation of turbulent mixing should be consistent with the requirements of conservation.

Indeed, the principle of conservation is inherent within the traditional picture of the surface layer. Specifically, conservation of the mass of water implies that the moisture flux $\overline{Q}$ must be constant with height in a steady state. Conservation of momentum implies that $\overline{\tau}$ must be constant. Similarly, conservation of energy implies that the total energy flux should be constant. At low wind speeds,

[1] In calculating the Obukhov length, the sensible heat flux at 10 m has been calculated from the temperature difference adjusted for dissipation, but using the exchange coefficients of Rutgersson et al. (2007).

where the kinetic energy is negligible, this is equivalent to the asserting that the heat flux $\overline{H}$ should be constant, as is traditionally done; but as the wind speed increases kinetic energy becomes important and its impact should be understood within the framework of total energy.

Air may be treated as an ideal gas possessing potential, kinetic, and internal energy. The potential and kinetic energies per unit mass are gz and $\mathbf{u}^2/2$ respectively. The internal energy per unit mass may be written as $c_v T$, where c_v is the specific heat at constant volume and is taken as constant. A complete representation of the thermodynamics of moist air would account for the separate contributions of dry air and the three phases of water to the specific heat and could be included following Iribane and Godson (1981), but would obscure the point being made here. The latent energy carried by water vapor is significant, even though its concentration is low. However, we shall concentrate on the simple case when there is no phase change within the surface layer. As the flux of water vapor is constant, so is the flux of latent energy and we may omit explicit latent contributions from the total energy. The vertical flux of total energy across a horizontal element of surface area is $\rho(c_v wT + gz + w\mathbf{u}^2/2)$. Work $pw = R_a \rho T w$ is also done on the fluid above the surface by that below, where R_a is the gas constant for air. Since $c_p = c_v + R_a$, the pressure work is generally absorbed into the first term by replacing c_v by c_p. Work done by an expanding air parcel is therefore consistently included if the flux is taken as $\rho(c_p wT + gz + w\mathbf{u}^2/2)$. Moreover, as a parcel of air moves through the fluid, its total energy is changed by the work done against pressure and by diabatic processes, which we denote as $\dot{\mathcal{Q}}$. Hence, we have (Batchelor 1967),

$$\frac{D}{Dt}\left(c_p T + \frac{\mathbf{u}^2}{2} + gz\right) = \frac{1}{\rho}\frac{\partial p}{\partial t} + \dot{\mathcal{Q}}. \tag{12}$$

It is therefore useful to introduce the quantity $E = c_p T + \mathbf{u}^2/2 + gz$.

Making the Reynolds decomposition, $\mathbf{u} = \overline{\mathbf{u}} + \mathbf{u}'$, and forming horizontal averages, we obtain

$$\overline{E} = c_p \overline{T} + gz + \frac{1}{2}\overline{\mathbf{u}}^2 + \overline{e}, \tag{13}$$

where $e = \mathbf{u}'^2/2$ is the TKE. Also,

$$\overline{w'E'} = c_p \overline{w'T'} + \overline{u}\,\overline{u'w'} + \overline{w'e'}. \tag{14}$$

The first and third terms represent the heat flux and the flux of TKE respectively, while the second is the work done on the fluid above the surface by that below. By considering a horizontal slab of fluid and neglecting any internal heating due to the absorption of radiation, we see that $\overline{w'E'}$ must be constant with height in a steady state. As an illustrative example, consider a slightly unstable surface layer. The surface layer gains energy from the heat flux at the surface and from the work done on it by the faster overlying flow. In a steady state there must be an equal loss due to the net upward flux of heat or TKE at the top of the layer.

To make further progress, a closure must be introduced, it is standard practice in surface layer theory to adopt a first-order closure. The turbulent diffusivity can be pictured physically as representing the mixing of fluid parcels originating from above and below the level where the fluxes are to be determined. The process of mixing must respect the principles of conservation. It is therefore reasonable to formulate an expression for the momentum flux in terms of the gradient of momentum. This gives that standard expression,

$$\overline{u'w'} = -K_M \frac{\partial \overline{u}}{\partial z}. \tag{15}$$

An equation for energy transport should similarly embody the principle that energy is conserved. Traditionally, the closure for the heat flux is formulated as

$$\overline{w'T'} = -K_H \frac{\partial}{\partial z}(\overline{T} + gz/c_p). \tag{16}$$

In that $T + gz/c_p$ is conserved as a parcel of fluid is moved adiabatically in a hydrostatically balanced atmosphere, this is a reasonable closure. More generally, however, when kinetic energy is significant, we suggest that the closure is better formulated in terms of E. Consider a parcel of fluid that is brought down from a region of fast mean flow toward the surface. It will experience warming by subsidence, as represented in the traditional closure, but as it interacts with the surrounding fluid its mean kinetic energy will be converted first into TKE and then into heat. This is process is more faithfully represented by formulating the closure as

$$\overline{w'E'} = -K_E \frac{\partial \overline{E}}{\partial z}. \tag{17}$$

More formally, we argue from Eq. (12) that a vigorously turbulent fluid will evolve to a state in which vertical variations in $\overline{E}$ are small. From Eq. (12), the equation for the variance of E is

$$\frac{\partial}{\partial z}\left(\frac{\overline{E'^2}}{2}\right) + \overline{w'E'}\frac{\partial \overline{E}}{\partial z} + \overline{w'\frac{\partial}{\partial z}\left(\frac{E'^2}{2}\right)} = \frac{1}{\rho}\overline{E'\frac{\partial p'}{\partial t}} + \overline{E'\dot{\mathcal{Q}}'}. \tag{18}$$

We expect that $E' = O(u_*^2)$ and that $p' = O(\rho u_*^2)$ and write $\partial\overline{E}/\partial z = U^2/L_E$, where U is a characteristic magnitude of $\overline{u}$ and L_E is the length scale for variations in $\overline{E}$. If we suppose that the length scale for turbulence is z and the time scale is z/u_*, we expect that $L_E = O(U^2 z/u_*^2) = O(z/C_D)$, where C_D is the drag coefficient. Over the sea, the neutral drag coefficient is about 0.001–0.0025. Consequently, the gradient of $\overline{E}$ must be small and we argue that a slight disequilibrium may be maintained by the forcing of the system by surface fluxes, making it plausible to introduce the closure above.

Substituting the expressions for $\overline{w'E'}$ and $\overline{E}$ into Eq. (17), we obtain

$$c_p\overline{w'T'} + \overline{w'e'} = -K_E\left[\frac{\partial}{\partial z}(c_p\overline{T} + gz) + \frac{\partial \overline{e}}{\partial z} + \overline{u}\frac{\partial \overline{u}}{\partial z}\left(1 - \frac{K_M}{K_E}\right)\right]. \quad (19)$$

This equation does not separate the flux of heat from that of TKE. At a general level, we would not expect a temperature gradient to drive a flux of TKE, which represents a more organized form of energy, but this yields no further simplification of Eq. (19). Instead, we turn to observations for guidance. Wyngaard and Coté (1971) measured turbulence statistics in the surface layer and interpreted them within the framework of Monin–Obukhov similarity. In stable conditions they concluded that turbulent transport of TKE was small, while in unstable conditions $\partial\overline{w'e'}/\partial z$ approximately balanced the buoyancy flux. We therefore expect the transport of TKE to be small relative to the heat flux in the surface layer by a factor of $\beta z/c_p$, where β is the buoyancy parameters: this will typically not exceed 0.0035 below 100 m. Moreover, $\overline{e}$ varies little with height in the surface layer and is of order u_*^2. Even in tropical cyclones (see section 5) u_* does not much exceed 3 m s^{-1}, so the TKE is negligible compared to the kinetic energy of the mean flow. It is therefore permissible to drop the terms involving the TKE.

If we also assume that $K_M = K_E$, Eq. (19) reduces to the traditional closure with $K_E = K_H$. However, if $K_M \neq K_E$, then we expect an additional contribution to the heat flux from the mean flow that is lacking in the traditional closure. Observations in nearly neutral flow suggest that the turbulent Prandtl number, K_M/K_H, is close to 1; but K_M is found to exceed K_H in stable flows and to be less than K_H in unstable flows (Li et al. 2015). Nevertheless, this contribution is unlikely to be significant in realistic conditions, where a combination of very high wind speeds and high instability would be difficult to achieve. It should also be recalled that in section 2, we adopted the traditional closure for the heat flux and then obtained an expression involving total energy only by assuming that the turbulent Prandtl number was unity. The analysis here shows that we are not constrained to assume this, but if we do not, an additional contribution should be included in the heat flux.

4. Application to similarity theory

The diffusivities, K_M and K_H are traditionally specified by applying Monin–Obukhov similarity theory (Garratt 1992). Close to the surface, where there is a local balance between production and dissipation, the relevant length scale is z, but further from the surface, where these processes are weaker, buoyancy is relatively more important and becomes significant on the scale of the Obukhov length, $L = -u_*^3/(k\beta\overline{w'\theta_v'}|_0)$, where θ_v is the virtual potential temperature and the buoyancy flux is nominally evaluated at the surface. All variables then become functions of the dimensionless variable z/L. While L is defined from the surface fluxes, when the fluxes do vary slowly with height, it is sometimes useful to define a local Obukhov length $\Lambda(z)$ from the local fluxes at a height z and similarity in $z/\Lambda(z)$ is assumed. The diffusivities are then written as

$$K_{\{M,H\}} = \frac{u_* kz}{\phi_{\{M,H\}}[z/\Lambda(z)]}, \quad (20)$$

where $\phi_{\{M,H\}}$ are the stability functions for momentum and heat. As the heat flux is taken as constant in the traditional version of the theory, a turbulent temperature scale is defined by $\theta_* = -\overline{w'\theta'}/u_*$, so that

$$\frac{\partial\theta}{\partial z} = \frac{\theta_*}{kz}\phi_H. \quad (21)$$

Following the discussion above, we assert that it is more appropriate to replace this equation by one relating to total energy. Because $\overline{w'E'}$ is constant over the surface layer, the turbulent energy scale, $E_* = -\overline{w'E'}/u_*$, is constant, unlike θ_*, and we have argued that $\overline{E}$ is a more accurately conserved variable than $\overline{T} + gz/c_p$. We therefore replace Eq. (21) with

$$\frac{\partial\overline{E}}{\partial z} = \frac{E_*}{kz}\phi_E. \quad (22)$$

Integrating the similarity equations for $\overline{u}$ and $\overline{E}$ between a reference height z_r and some other height z in the usual way, we obtain

$$\overline{u}(z) - \overline{u}(z_r) = u_*\int_{z_r}^{z}\frac{\phi_M}{kz}dz = \frac{u_*}{k}\Phi_M \quad (23)$$

and

$$\overline{E}(z)-\overline{E}(z_r)=E_*\int_{z_r}^{z}\frac{\phi_E}{kz}dz=\frac{E_*}{k}\Phi_E, \tag{24}$$

where these equations define Φ_M and Φ_E. Hence

$$\overline{w'E'}=-\frac{k^2}{\Phi_M\Phi_E}[\overline{u}(z)-\overline{u}(z_r)][\overline{E}(z)-\overline{E}(z_r)]. \tag{25}$$

Ignoring the contributions from the TKE, we have $\overline{w'E'}|_r=c_p\overline{w'T'}|_r+u_r\overline{u'w'}$, so that

$$\begin{aligned}c_p\overline{w'T'}&=-\frac{k^2}{\Phi_M\Phi_E}[\overline{u}(z)-\overline{u}(z_r)][\overline{E}(z)-\overline{E}(z_r)]\\&\quad+\frac{k^2}{\Phi_M^2}\overline{u}(z_r)[\overline{u}(z)-\overline{u}(z_r)]^2\\&=-c_p\frac{k^2}{\Phi_M\Phi_E}[\overline{u}(z)-\overline{u}(z_r)]\Big\{\overline{\theta}(z)-\overline{\theta}(z_r)\\&\quad+\frac{1}{2c_p}[u(z)-u(z_r)]^2\\&\quad+\left(1-\frac{\Phi_E}{\Phi_M}\right)\frac{\overline{u}(z_r)}{c_p}[\overline{u}(z)-\overline{u}(z_r)]\Big\}.\end{aligned} \tag{26}$$

Diffusivities are assumed to be determined by local scaling and thus depend on the local buoyancy flux. Following Stull (1988), we have

$$\begin{aligned}\overline{w'\theta'_v}&\approx\overline{w'\theta'}+(\varepsilon_M^{-1}-1)\overline{\theta}\,\overline{w'q'}\\&=\frac{1}{c_p}(\overline{w'E'}-\overline{u}\,\overline{u'w'}-\overline{w'e'})\\&\quad+(\varepsilon_M^{-1}-1)\overline{\theta}\,\overline{w'q'},\end{aligned} \tag{27}$$

where ε_M is the ratio of the molecular weights of water and dry air. The fluxes of total energy and moisture are constant, so we obtain

$$\frac{\partial}{\partial z}\left(\frac{1}{\Lambda}\right)=\frac{k\beta}{c_pu_*^3}\frac{\partial}{\partial z}(\overline{w'e'}-\overline{u}\,\overline{u'w'}). \tag{28}$$

Employing the budget of TKE (Wyngaard and Coté 1971), we may rewrite this equation as

$$\frac{\partial}{\partial z}\left(\frac{1}{\Lambda}\right)=-\frac{\beta}{c_p}\left(\frac{\phi_\varepsilon}{z}+\frac{\phi_p}{z}+\frac{1}{\Lambda}\right), \tag{29}$$

where ϕ_ε and ϕ_p are the similarity functions for dissipation and pressure transport. Hence,

$$\frac{1}{\Lambda}=\frac{1}{\Lambda_r}-\frac{\beta}{c_p}G\left(\frac{z}{\Lambda_r}\right), \tag{30}$$

for some function G, where $G(z_r/\Lambda_r)=0$. The length scale for variations of Λ^{-1} is therefore c_p/β, which is on the order of 30 km and much larger than the depth of the surface layer, so it suffices to evaluate Λ^{-1} at the reference height z_r and to take the similarity variable as $\zeta=z/\Lambda_r$. This means that we relate the effect of frictional heating to its effect only on the fluxes and not to a modification of the stability functions. We shall also assume that ϕ_E may be taken as identical to ϕ_H, being a common similarity function for scalars.

a. Surface fluxes and the exchange coefficient

If we specifically set $z_r=z_{0m}$ in Eq. (26) and introduce a thermal roughness length following Eq. (6), we may derive the bulk flux formula incorporating frictional heating as

$$\begin{aligned}\overline{H}(0)&=\rho c_p\frac{k^2}{\Phi_M(\zeta,\zeta_{0M})\Phi_E(\zeta,\zeta_{0H})}\\&\quad\times\overline{u}(z)\left[\overline{\theta}_s-\frac{\overline{u}^2(z)}{2c_p}-\overline{\theta}(z)\right].\end{aligned} \tag{31}$$

This is equivalent to Eq. (8) in neutral conditions, except that σ_N no longer multiplies the middle term in brackets. It differs from the traditional form in two ways. First, a term in $\overline{u}^2(z)$ appears alongside the temperature difference, exactly as in section 2. Second, and less importantly, the inclusion of the term in $\overline{u}^2(z)$ slightly modifies the Obukhov length and the exchange coefficient. The impact is illustrated in Fig. 3 which shows the turbulent virtual temperature scale based on the surface flux as a function of the difference in potential temperature between the surface and 10 m for a wind speed of 10 m s^{-1}. The thick line shows the result obtained from traditional surface layer similarity theory and the dash–dotted line the impact of accounting for frictional dissipation, calculated as the difference between the result obtained using Eq. (31), allowing for the effect of frictional dissipation on the Obukhov length, and the standard result. The dotted line shows the effect of neglecting the impact of dissipative heating on the Obukhov length and including only the adjustment to the temperature difference; this is small relative to the impact of adjusting the temperature difference. At this wind speed, the effect of frictional dissipation is very small, but it becomes more significant at higher wind speeds. However, at such higher wind speeds, the surface layer becomes more nearly neutral and the impact of dissipation on the Obukhov length becomes smaller. In practice, therefore, it is adequate to include only the adjustment to the temperature difference.

b. The apparent similarity function

Taking z_r as a height of observation several meters above the surface, we can consider how frictional heating

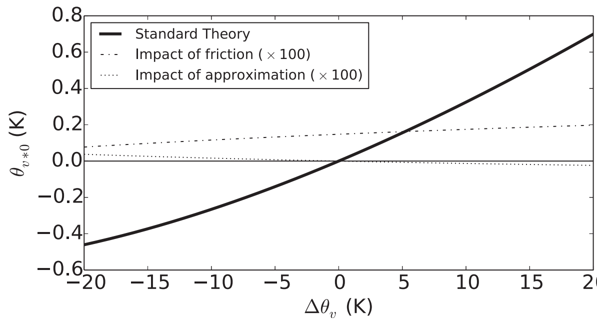

FIG. 3. The turbulent virtual temperature scale defined from the surface flux θ_{v*0} as function of the difference in potential temperature between the surface and 10 m for a wind speed of 10 m s^{-1}. The thick solid line shows the result from standard similarity theory, the dash–dotted line the impact of accounting for frictional dissipation, and the dotted line the impact of the approximation wherein the impact of frictional dissipation on the exchange coefficient is neglected.

might affect observational determinations of the similarity function.

Indeed, suppose that an apparent similarity function $\tilde{\phi}_H(z/L)$ is deduced from the fluxes at a height z_r and profiles of the virtual potential temperature. If the standard assumption that z_r lay within the constant flux layer were made, the surface Obukhov length L would be assumed to be Λ_r and ζ, as defined here, would equal the assumed value of z/L. For simplicity, since the neutral turbulent Prandtl number is close to unity, and frictional heating is significant only in nearly neutral cases, we shall take $\phi_H = \phi_M$; we shall also neglect the gradient and flux of TKE relative to the gradient and flux of potential temperature, so that Eq. (19) reduces to the standard form $\partial\overline{\theta}/\partial z = \theta_*/(kz)$. From Eq. (14),

$$E_* = c_p\theta_* + \overline{u}(z)u_* = c_p\theta_{*_r} + \overline{u}(z_r)u_*, \qquad (32)$$

where the subscript r on θ_* indicates that it is evaluated at the reference height; u_* is constant with height.

Standard experimental practice is to approximate vertical gradients using finite differences between two levels, z_2 and z_1, above and below z_r respectively; $\overline{\theta}_v$ is a convenient choice because of its direct link to buoyancy. Writing $\Delta X = X(z_2) - X(z_1)$ for the difference in any quantity X evaluated at z_2 and z_1, we have

$$\tilde{\phi}_H(\zeta_r) = \frac{kz_r}{\theta_{v*_r}}\frac{\partial\theta_v}{\partial z} \approx \frac{kz_r}{\theta_{v*_r}}\frac{\Delta\theta_v}{\Delta z} = \frac{z_r}{\Delta z}\Delta\Phi_H - \frac{u_*^2}{k\theta_{v*_r}}\frac{1}{2c_p}\frac{z_r}{\Delta z} \times \left\{\left[\ln\left(\frac{\zeta_2}{\zeta_r}\right)\right]^2 - \left[\ln\left(\frac{\zeta_1}{\zeta_r}\right)\right]^2\right\}, \qquad (33)$$

where, since the frictional correction is small, we have retained only the neutral contributions to the similarity functions. Supposing, for illustration, that the underlying similarity function is $\phi_H(\zeta) = 1 + \gamma_H\zeta$, we obtain

$$\tilde{\phi}_H(\zeta_r) = \frac{1}{\Delta z}\ln\left(\frac{z_2}{z_1}\right) + \gamma_H\zeta_r - \frac{\beta z_r^2}{2c_p}\mathscr{G}\frac{1}{\zeta_r}, \qquad (34)$$

where $\mathscr{G} = \{[\ln(z_2/z_r)]^2 - [\ln(z_1/z_r)]^2\}/(z_2 - z_1)$. For fixed levels z_1, z_2, and z_r, the right-hand side of this equation is a function of ζ_r only. Considered as such, $\phi_H \to \pm\infty$ as $\zeta_r \to 0-$, with the sign depending on that of $\mathscr{G}$. The last two terms are equal when $\zeta_r = -\sqrt{\sigma_N\beta|\mathscr{G}|z_r^2/(2c_p\gamma_H)}$, which will typically be less than about 0.01, and the singularity will become apparent if ζ_r is closer to 0. Typically, this is too small to affect standard determinations. However, note that if the similarity function is determined from observations of θ, not θ_v, the singularity will be shifted to $\zeta = z_r k\beta(\varepsilon_M^{-1} - 1)\overline{\theta}q_*/u_*^2$, which may be rather larger.

Qualitatively, this behavior is similar to that observed by Smedman et al. (2007a,b), who explained it in terms of their UVCN regime. Indeed, a singularity should be expected in the presence of any process affecting the temperature profile that does not scale with θ_* alone and in practice it may be hard to separate frictional effects from those of other processes. Nevertheless, this indicates the need for care in the interpretation of experimental data in nearly neutral conditions and to account for frictional dissipative heating.

5. Implications for tropical cyclones

Because the impact of the adjustment to the bulk flux formula increases with wind speed, it is natural to consider how it might affect models of TCs. It is important to note that in the extreme conditions of a TC, a traditional surface layer may not exist. The point is controversial and for a criticism of the concept the reader is referred to Smith and Montgomery (2014), who, while conceding that a shallow "logarithmic" layer may exist, assert that such a layer cannot extend to depths of 100–200 m. Nevertheless, the traditional concepts of surface layer similarity theory are typically adopted in conceptual and numerical models of TCs, where they provide the connection between the atmospheric state and the surface fluxes that drive the storm. By modifying that connection, the proposed modification to the bulk flux formula may therefore affect the evolution of the simulated TC. We consider first a widely discussed idealized model of a TC and then some numerical simulations. As we need not consider fluctuations in this section, we shall drop the overbars on mean quantities.

a. Emanuel's axisymmetric model

The flux of enthalpy from a warm sea surface is the motive force in tropical cyclones. Emanuel (1986) developed an axisymmetric model of a tropical cyclone, incorporating surface fluxes, that leads to an expression for the azimuthal velocity in the eyewall of the form

$$V^2 = \frac{T_s - T_o}{T_s}\frac{C_K}{C_D}(k^* - k), \tag{35}$$

where T_s is the temperature of the sea surface and T_o is the outflow temperature in the upper troposphere; C_K and C_D are the exchange coefficients for momentum and enthalpy, k^* is the specific enthalpy at the sea surface, and k that of the boundary layer. Thus, the maximum wind speed is proportional to $\sqrt{C_K/C_D}$ so there is considerable interest in the value of this ratio at wind speeds typical in intense tropical storms. It has, for example, been suggested that sea spray may greatly enhance sensible heat fluxes (Andreas 2011), causing C_K/C_D to be significantly greater than 1 in such systems. Nevertheless, the effect of sea spray remains uncertain (Veron 2015).

In the original version of the model, dissipative heating was neglected, but it was included by Bister and Emanuel (1998), who showed that Eq. (35) should be modified to

$$V^2 = \frac{T_s - T_o}{T_o}\frac{C_K}{C_D}(k^* - k), \tag{36}$$

that is, that T_o should replace T_s in the denominator on the right-hand side. They commented that since $T_s/T_o \approx 3/2$ in such systems, dissipative heating should play a significant role in developing the storm.

It has been argued above that, in the presence of dissipation, the sensible heat flux at the surface should be calculated using an adjusted temperature that is cooler than the actual one. This reduces the flux of enthalpy into the system. Consequently, Eq. (17) of Bister and Emanuel (1998), namely

$$-\frac{1}{R}\frac{\partial \ln\theta_e}{\partial R} = \frac{f}{c_p T_s}\frac{1}{rV}\left[\frac{C_K}{C_D}(k^* - k) + |\mathbf{V}|^2\right], \tag{37}$$

should be modified to

$$-\frac{1}{R}\frac{\partial \ln\theta_e}{\partial R} = \frac{f}{c_p T_s}\frac{1}{rV}\left[\frac{C_K}{C_D}(k^* - k) + \left(1 - \frac{C_K}{2C_D}\right)|\mathbf{V}|^2\right]. \tag{38}$$

The equation for the maximum wind speed then becomes

$$V^2 = \frac{T_s - T_o}{T_o}\frac{C_K}{C_D}(k^* - k)\left[1 + \frac{C_K}{2C_D}\left(\frac{T_s}{T_o} - 1\right)\right]^{-1}. \tag{39}$$

Although Eqs. (35) and (36) were derived by Emanuel (1986) and Bister and Emanuel (1998) respectively using dynamical reasoning, they may also be derived less formally by viewing the tropical cyclone as a Carnot cycle operating between the temperatures T_s and T_o (Emanuel 1986, 2003; Bister et al. 2011). However, by recasting these derivations in terms of the total energy budget of the cyclone, Kieu (2015) has recently shown that the difference between Eqs. (35) and (36) is somewhat paradoxical.

Kieu (2015) presents the energy budget in two ways. On the one hand, we may consider a cylindrical volume extending from the surface to the tropopause, centered on the cyclone and enclosing the whole circulation. Following Emanuel (2003) and Bister et al. (2011), we assume that the surface flux budget is dominated by a region of area A around the radius of maximum wind. Using the traditional bulk flux formula, the input of enthalpy to the cyclone is $Q_i = A\rho C_K V(k^* - k)$. The kinetic energy of the circulation is dissipated in the surface layer and the dissipation is $D = A\rho C_D V^3$. No work is performed on the environment, so the loss of energy from the upper troposphere Q_o is equal to Q_i. The dissipation D is not an extra source of energy, but does function as a source of entropy at a temperature T_s (Bister et al. 2011), so, if the system does behave as a Carnot cycle,

$$\frac{Q_i + D}{T_s} = \frac{Q_o}{T_o}, \tag{40}$$

leading to Eq. (36).

Alternatively, we may consider a second cylinder enclosing the cyclone, identical to the first but excluding the surface layer. As the surface layer is still viewed as a constant flux layer for heat, the traditional bulk flux formula gives $Q_i = A\rho C_K V(k^* - k)$ as the input of enthalpy to the system, but now the circulation above the surface layer does work $W = AV\rho C_D V^2$ on the surface layer itself, so $Q_o = Q_i - W$ and, if the system behaves as a Carnot cycle,

$$\frac{Q_i}{T_s} = \frac{Q_o}{T_o}, \tag{41}$$

because dissipation takes place only in the surface layer below and outside the cylinder: this leads to Eq. (35). However, as Kieu (2015) notes this derivation is physically equally as valid as that leading to Eq. (36): indeed dissipation is not discussed, but it is not relevant to the energy and entropy budgets of this domain and the derivation is energetically consistent. Kieu (2015) suggests that this paradox may indicate that some of the assumptions made by Bister and Emanuel (1998), while appropriate in the well-mixed layer, may fail in the surface layer.

The results obtained in sections 2 and 3 suggest a simpler resolution of the paradox: the expressions for Q_i used in the derivations just presented are incomplete. In the first case, where we consider a cylinder extending to the surface, the frictional adjustment implied by Eq. (8) gives $Q_i = A\rho C_K V(k^* - k - V^2/2)$. In the second case, using Eq. (9) and noting that $\ln(z/z_{0h})/\ln(z/z_{0m}) = C_D/C_K$, we have $Q_i = A\rho C_K V[k^* - k + V^2(C_D/C_K - 1/2)]$. Using these expressions, we obtain the same equation for V^2 for each cylinder and that equation is the same as Eq. (39).

The estimate of V^2 from Eq. (36) is greater than that from Eq. (35) by a factor of T_s/T_o, which is 1.5 for the typical values of T_s = 300 K and T_o = 200 K (Bister and Emanuel 1998), but the ratio between the estimates from Eqs. (39) and (35) depends on the value of C_K/C_D. Bell et al. (2012) have estimated C_D and C_K at high wind speeds by considering the budgets of angular momentum and total energy in a toroidal region of rectangular cross section around the center of a tropical cyclone. An azimuthally averaged model incorporating data assimilation is used to obtain the fluxes of angular momentum and total energy across the vertical bounding cylinders and the upper annular boundary of this region. The fluxes across the lower annular boundary, 10–100 m above the surface, are then obtained by considerations of conservation. They estimate that the most likely value of C_K/C_D at high wind speeds is 0.4, so the value of V^2 predicted from Eq. (39) will be about 36% greater than that from Eq. (35), rather than 50% greater.

As Bell et al. (2012) note, frictional dissipation is a process of conversion of energy and need not be considered explicitly within their toroidal control volume, so the analysis presented here does not affect their estimates of the enthalpy flux across the lower annulus. However, their estimates of C_K are obtained using a traditional bulk flux formula, neglecting the effect of frictional dissipation below the lower annular boundary of the control volume. Therefore a modification of their estimates of C_K is potentially required. If frictional dissipation in the surface layer were neglected, for a given flux of enthalpy F_{zk} at a height z, one would deduce an exchange coefficient,

$$C_{KT} = F_{zk}/\{\rho|\mathbf{u}_h(z)|[k^* - k(z)]\}, \qquad (42)$$

where, following their notation, $|\mathbf{u}_h|$ is the horizontal wind speed. Including frictional dissipation, one obtains

$$C_{KF} = F_{zk}/\{\rho|\mathbf{u}_h(z)|[k^* - k(z) - (1/2)u_h^2(z)]\}, \qquad (43)$$

so that

$$C_{KF}/C_{KT} = 1 - \frac{u^2(z)/2}{k^* - k(z)}. \qquad (44)$$

Bell et al. (2012) obtain $F_{zk} \approx 2000\,\mathrm{W\,m^{-2}}$ at wind speeds of about 70 m s^{-1}, from which they estimate that $C_{KT} \approx 0.001$. This implies that $k^* - k(z) \approx 28.5\,\mathrm{kJ\,kg^{-1}}$, so that $C_{KF}/C_{KT} \approx 0.91$, thus implying that only a small reduction in their inferred value of C_K is indicated. The ratio of V^2 predicted from Eq. (39) to that from Eq. (35) is increased very slightly from 1.36 to 1.375. Alternatively, comparing Eqs. (36) and (39), we see that the maximum wind speed is 4% lower in the latter case.

On the basis of this model, we might expect the modification of the bulk flux formula to lead to a small reduction in the intensity of tropical cyclones in numerical simulations: this we now assess.

b. *Simulated impact on intense tropical cyclones*

Three intense tropical cyclones were selected for numerical study on the basis of their high peak winds speeds. Supertyphoon Vongfong occurred in the northwestern Pacific early in October 2014 and attained an estimated peak wind speed of 155 kt (80 m s^{-1}), as described in the Annual Tropical Cyclone Report for 2014 of the Joint Typhoon Warning Center (JTWC), available online (from https://metoc.ndbc.noaa.gov). Hurricane Genevieve occurred in the central Pacific in August of the same year and attained wind speeds of 140 kt (72 m s^{-1}), while Supertyphoon Meranti occurred in the western Pacific in September 2016.

For each storm, two simulations have been carried out using a nested configuration of the Met Office Unified Model (MetUM). Scientifically, the model follows the configuration Global Atmosphere 6.0 (GA6.0) described by Walters et al. (2017). The control simulations (CNTL) use the standard configuration of the model, while, in the second simulation of each pair (FRIC), the bulk flux formula for the sensible heat flux is adjusted as described above. The parameterization of surface exchange in the model is based on the stability functions of Beljaars and Holtslag (1991), so the neutral turbulent Prandtl number is equal to 1 by default. Charnock's coefficient is set to 0.018, independent of wind speed, and, as an indicative value, the ratio of exchange coefficients C_K/C_D is 0.31 at 60 m s^{-1}. Frictional dissipation is included in the model following the parameterization of Zhang and Altshuler (1999), except that, to ensure the stability of the model in a range of forecasting applications, the total heating increment is applied with an

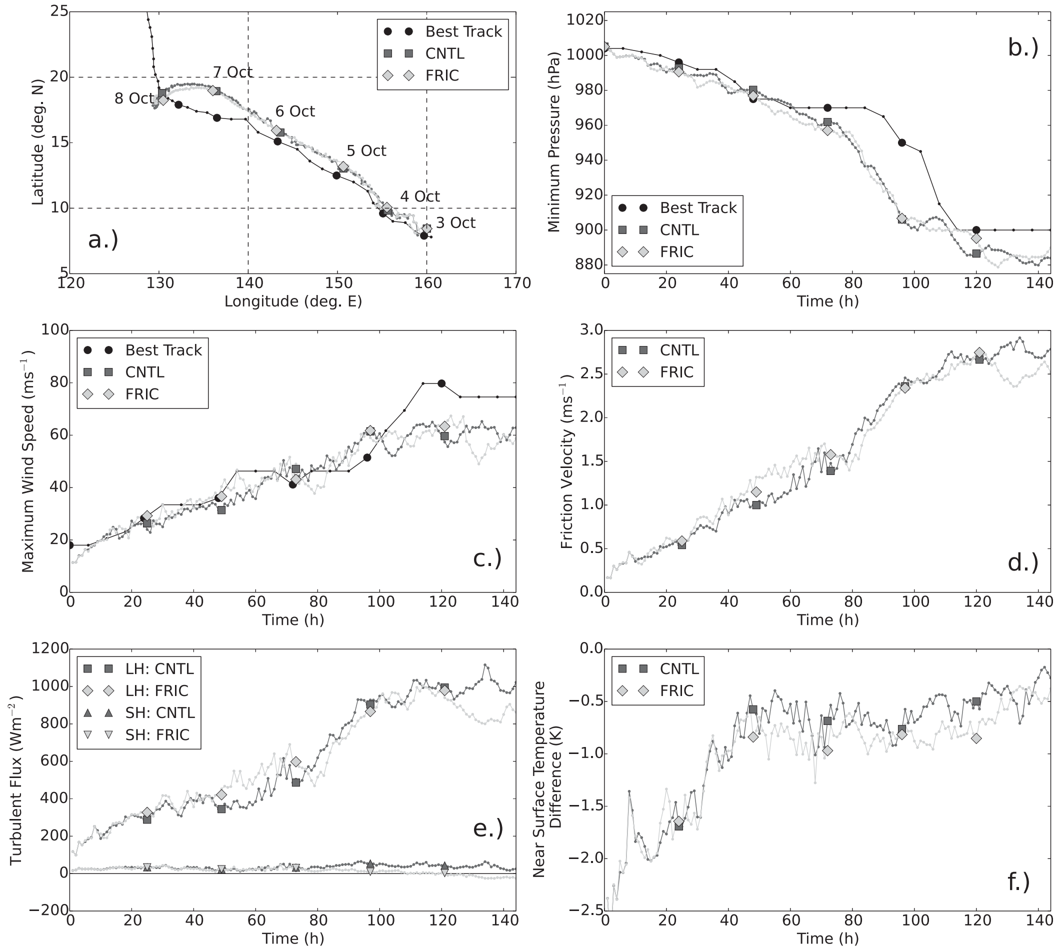

FIG. 4. Supertyphoon Vongfong. (a) The analyzed track and the simulated tracks generated in CNTL and FRIC. (b) The evolution of the minimum pressure as analyzed and as simulated in CNTL and FRIC. (c) As (b), but for the maximum 10-m wind speed. (d) The evolution of the friction velocity, calculated from the average magnitude of the surface stress in a 1° × 1° box moving with the storm, as simulated in CNTL and FRIC. (e) As (d), but showing the average values of the simulated latent and sensible heat fluxes at the surface within the comoving 1° box. (f) As (e), but for the difference between the near-surface temperature at 1.5 m and the surface temperature.

imposed linear profile across the boundary layer. We emphasize that the purpose of the simulations is to assess the modification of the bulk flux formula, not the inclusion of frictional heating itself.

The simulations are initialized from operational analyses valid at 0000 UTC some 3–5 days before the time of maximum wind speed and cover a period of 6 days. Global simulations at a resolution of 0.5625° × 0.375°, reinitialized from analyses at 0000 UTC each day, provide lateral boundary conditions for simulations at a resolution of 0.02° × 0.02° (approximately 2.1 km east–west × 2.2 km north–south) within an inner domain covering the region bounded by 5°–25°N, 125°–165°E. While the global model is reinitialized each day, no reinitialization of fields within the inner domain is performed, allowing cumulative differences between CNTL and FRIC to develop. Note also that an interactive ocean model is not included and sea surface temperatures are persisted through the simulations. A time step of 1 min is used in the inner domain and time step values of wind and pressure are compared with observations. Best track data (Knapp et al. 2010) are used to assess simulations.

Figure 4a shows the observed track and the tracks forecast in the inner domains of the simulations CNTL and FRIC in the case of Vongfong. Differences in the forecast tracks are very small compared to the errors in either forecast, though the track of FRIC is slightly to the left of that in CNTL. Figure 4b shows the forecast minimum pressure in the inner domains as a function of time measured from 0000 UTC 3 October 2014, together with the best track data, while Fig. 4c shows the corresponding maximum wind speeds. Relative to the best

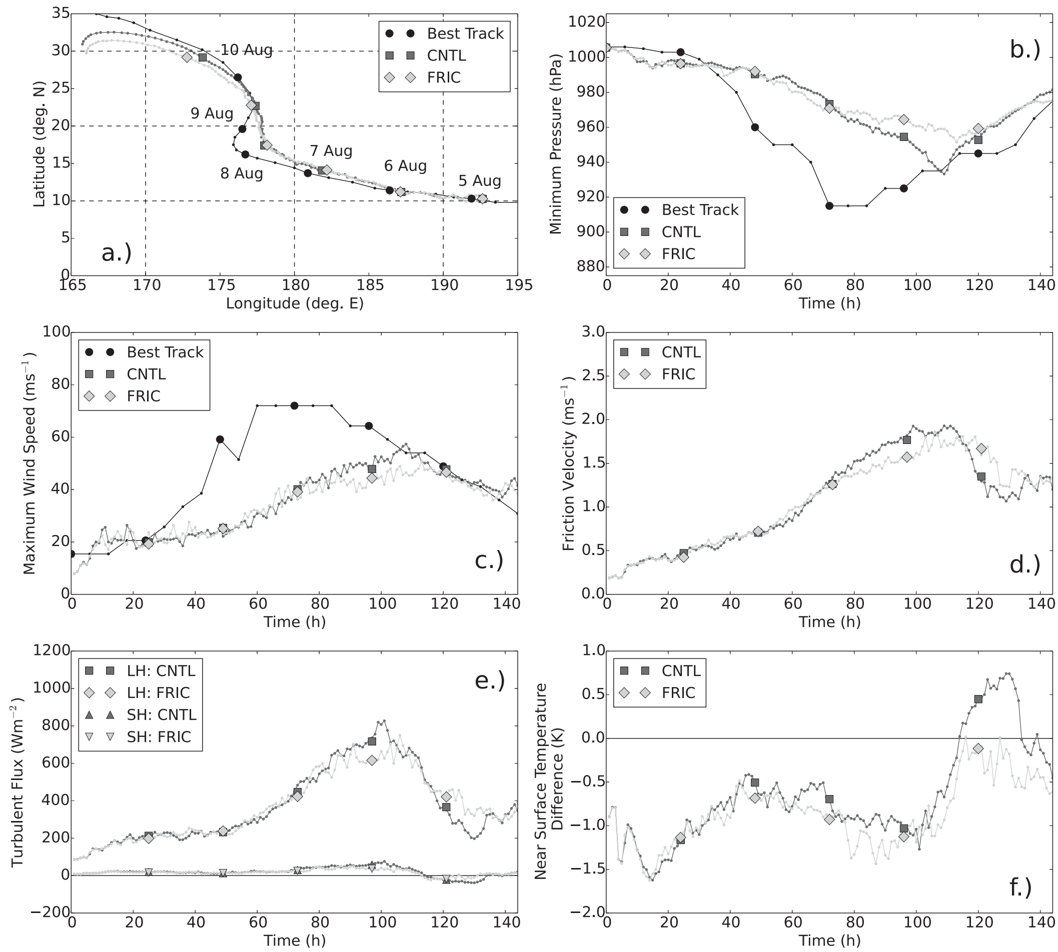

FIG. 5. As in Fig. 4, but showing Hurricane Genevieve.

track data, both simulations exhibit excessively low central pressures after 70 h. The maximum wind speeds are relatively well forecast until 100 h, after which time they are low. While there are differences of detail between the two simulations, they are small relative to the differences from the observational estimates.

Turning to the surface fluxes, Fig. 4d shows the evolution of the simulated friction velocity, while Fig. 4e shows the corresponding latent and sensible heat fluxes. These have been computed as averages over a $1° \times 1°$ box centered on the location of the minimum pressure. In FRIC, the friction velocity and latent heat flux are slightly higher around 60 h and slightly lower around 120 h (when the storm is most intense) than in CNTL, though in relative terms the differences are modest. Correspondingly, the accumulated precipitation in FRIC is about 17% larger then in CNTL around 70 h (not shown), but thereafter the difference becomes smaller. The increase in the latent heat flux in FRIC reflects changes in the boundary layer profiles of wind and humidity, because, as noted in section 2, in nearly neutral conditions and for given profiles, accounting for dissipative heating does not enhance the moisture flux.

Larger relative differences are seen in the sensible heat flux, which appears systematically lower in FRIC than in CNTL after 70 h, as would be expected; it becomes negative after 120 h. The sensible heat flux, averaged over the whole simulation, is reduced from 34 W m^{-2} in CNTL to 13 W m^{-2} in FRIC, a relative reduction of 38%, although it is the latent contribution that dominates the surface fluxes and a reduction of only 2% is seen in the average latent heat flux. Although the sensible heat flux is reduced in FRIC, the near-surface temperature gradient becomes more negative (Fig. 4f) and relative to the sea surface temperatures the near-surface air temperature is about 0.2 K lower.

Results for the equivalent pairs of simulations for the cases of Genevieve and Meranti are shown in Figs. 5 and 6 respectively. In the main, these also show only subtle differences between the simulations. In the case of

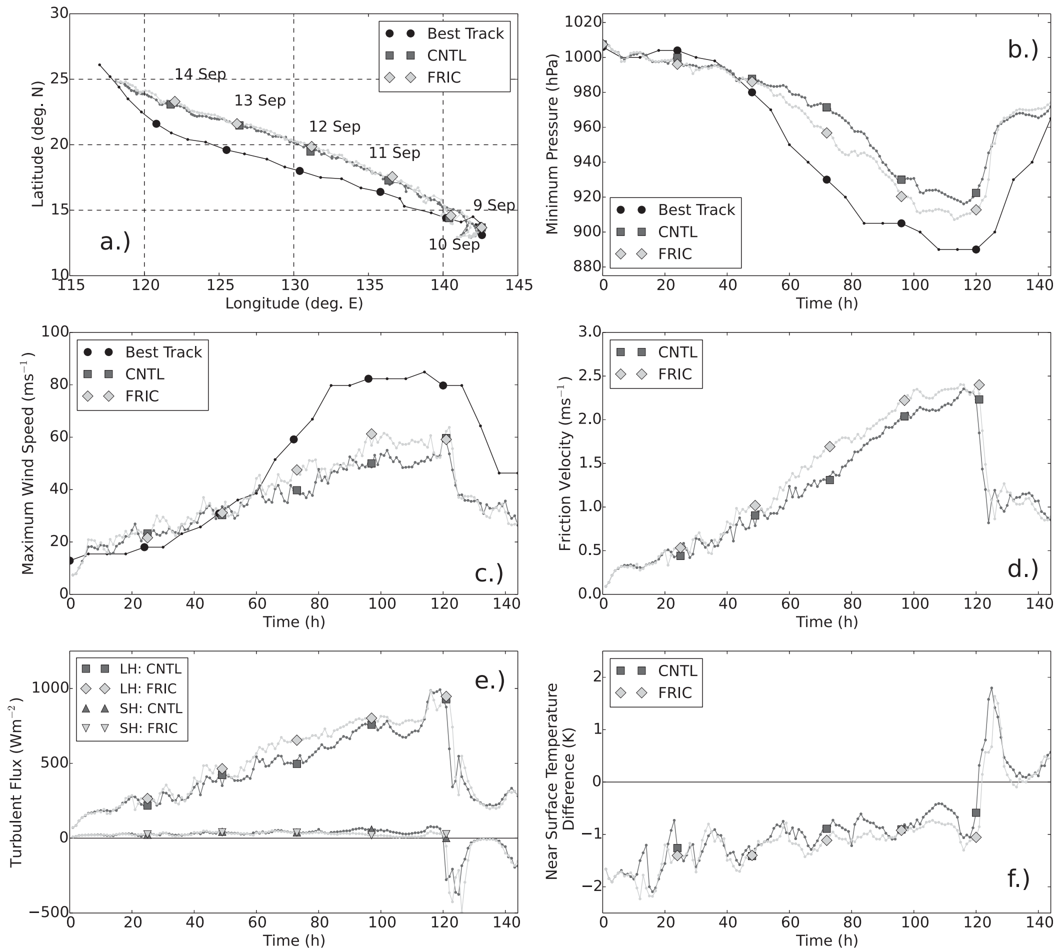

FIG. 6. As in Fig. 4, but showing Supertyphoon Meranti.

Genevieve, the track of FRIC is slightly to the left of that of CNTL, but the displacement is to the right in the case of Meranti. Both simulations fail to simulate the rapid intensification of the Genevieve, but CNTL yields a lower minimum pressure around 110 h. In the case of Meranti, which made landfall on Taiwan at $T + 120$ h, FRIC is actually slightly deeper than CNTL. In both cases, the near-surface air temperatures are slightly colder relative to the underlying surface in FRIC. There is possibly an indication that FRIC intensifies slightly more rapidly than CNTL at around 70 h in the cases of Vongfong and Meranti, when the wind speed and latent heat fluxes are higher, but the equivalent difference is less prominent than in the case of Genevieve.

Overall, only small differences are found, and it is hard to separate systematic effects from random differences between the simulations, although a reduction in the sensible heat during the periods of highest winds and cooling of the surface layer relative to the surface appear consistent between the simulations.

6. Climatological impact

The differences in surface fluxes just obtained for TCs are relatively modest, but perhaps large enough to suggest a minor systematic climatological effect. Indeed, when Sahlée et al. (2008) assessed the impact of the enhanced exchange coefficients proposed by Rutgersson et al. (2007) on a global scale, they found an enhancement of the sensible heat flux from the Southern Ocean (defined as 30°–60°S) of 2.3 W m^{-2} in the annual mean and an enhancement of the latent heat flux of 6.5 W m^{-2}. They obtained these values by direct calculation of the sensible and latent heat fluxes from the 6-hourly fields in an atmospheric reanalysis, using bulk flux formulas with and without adjustment for their UVCN regime. Note that these estimates represent the direct effect of their change and, in practice, conservation of energy within the atmospheric column will require adjustments of other fields in the model, so the eventual resultant impact is likely to be different.

In the Southern Ocean, where the annual-mean wind speed is about 12 m s^{-1}, the direct effect of the adjustment to the bulk flux formula proposed here would be to make the sensible heat flux more negative by about 1 W m^{-2} (disregarding the effect of temporal variations that will increase its magnitude). The state of the model will then adjust to balance the surface flux budget. To assess the overall impact, two 20-yr Atmospheric Model Intercomparison Project (AMIP) climate simulations have been performed using the global climate configuration of the MetUM at GA6.0 (Walters et al. 2017), one being the control and the other including the adjustment to the bulk flux formula. Figure 7 shows the impact on the components of the zonal-mean surface flux budget. In the Southern Ocean, and also at the latitudes of the storm tracks in the Northern Hemisphere, the sensible flux is indeed more negative by about 0.4 W m^{-2}. Thus, the effect is small relative to the annual-mean sensible heat flux of about 10 W m^{-2} and the annual mean latent heat flux of about 40 W m^{-2}. However, the standard deviation of the annual averages of the zonal-mean sensible heat flux is about 0.8 W m^{-2} in these regions, so we cannot safely conclude from these simulations that the effect is significant.

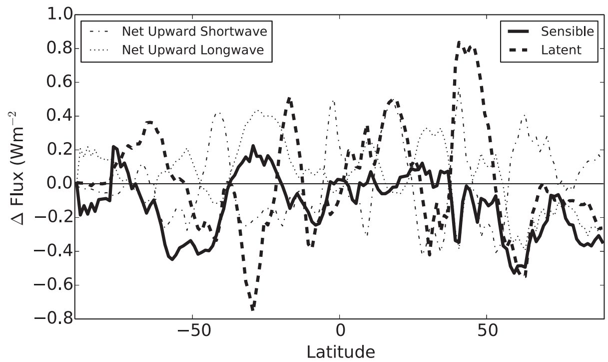

FIG. 7. The impact of the frictional adjustment of the bulk flux formula on the zonal-mean surface fluxes in a 20-yr AMIP simulation.

7. Summary

Frictional heating in the atmospheric boundary layer is recognized as significant at the wind speeds attained in tropical cyclones, but it is usually neglected at moderate wind speeds. Here, it has been argued that its impact in the surface layer should become apparent at wind speeds above about 10 m s^{-1} and that it may be a significant factor in observations of enhanced heat transfer and countergradient fluxes in nearly neutral conditions (although these are subtle effects). Other explanations for such effects have also been suggested (Smedman et al. 2007a; Mahrt et al. 2012) and an emphasis on the role of frictional dissipation should not be interpreted as discounting the possibility that multiple mechanisms are in play. An interesting topic of future research might be to examine the residual discrepancy in observations, having accounted for frictional heating. Nevertheless, at low or moderate wind speeds these effects are small and the observational challenges should not be underestimated.

However, the bulk flux formula for the surface heat flux should be modified to account for frictional dissipation in the surface layer in such a way that the surface sensible heat flux is actually less positive than that obtained from an unmodified formula. This adjustment is consistent with an interpretation of the surface layer as one across which the total energy flux is constant. Indeed this seems a more natural way to view the surface layer. Consideration of the total energy does not constrain the turbulent Prandtl number, but does suggest that, if this differs from unity, velocity gradients should drive a flux of heat or TKE, though the effect will be small.

Because the magnitude of the adjustment increases with wind speed, it is of potentially greater significance in the case of tropical cyclones. This possibility has been assessed first in the case of an idealized axisymmetric model, where it was shown to resolve a paradox regarding the impact of frictional dissipation, and in numerical simulations of three intense tropical cyclones. The impact is subtle in all cases, but a slight reduction of the sensible heat flux into the storm and a cooling of the surface layer during the period of highest winds appear consistent features of the three simulations presented.

In summary, while the magnitude of the effect is small, it is of relevance to the conceptual understanding of the very nearly neutral boundary layer and to the explanation of apparently anomalous behavior observed in such conditions.

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