# Comments on "Development and Application of a Physical Approach to Estimating Wind Gusts"

STEPHEN D. BURK AND WILLIAM T. THOMPSON

*Marine Meteorology Division, Naval Research Laboratory, Monterey, California*

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In a recent study, Brasseur (2001, hereafter B01) developed a new wind gust estimate (WGE) method that is based upon dynamical considerations rather than empirical relationships. In this approach the WGE is designed to be calculated within NWP models that compute turbulent kinetic energy (TKE). The essence of the WGE method is contained in Eq. (1) and Fig. 4 of B01. For convenience, we reproduce his Eq. (1) here:

$$\frac{1}{z_p}\int_0^{z_p} E(z)\,dz \geq \int_0^{z_p} g\frac{\Delta\theta_v(z)}{\Theta_v(z)}\,dz, \quad (1)$$

where $z_p$ is the height of a given parcel, $E$ is the TKE, g is gravitational acceleration, $\Theta_v(z)$ is the mean virtual potential temperature, and $\Delta\theta_v(z)$ is the change in virtual potential temperature over a given layer. The integrals are meant to extend only over the depth of the boundary layer (BL); that is, the height $z_p$ should lie within the BL. As B01 states in relation to Eq. (1): "Estimating wind gusts is done assuming a parcel flowing at a given height will be able to reach the surface if the mean turbulent kinetic energy of large turbulent eddies is greater than the buoyant energy between the surface and the height of the parcel." Such parcels are assumed to bring their momentum to the surface as gusts.

We find the B01 approach to be very appealing (despite the usual limitations of parcel theory) because of its conceptualization of the physical processes leading to wind gusts. Our purpose in this comment is to suggest two alterations or extensions to the B01 approach, as well as a potential very different application of this method. First, we note that care must be exercised in the computation of buoyancy within cloud layers. Thus, we suggest that $\Delta\theta_v(z)$ in Eq. (1) be replaced with

$$\Delta\theta_v(z) = \alpha\Delta\theta_l(z) + \beta\Delta q_w(z), \quad (2)$$

where $\theta_l$ is liquid water potential temperature and $q_w$ is total moisture. The coefficients $\alpha$ and $\beta$, which alter the relative weighting of $\Delta\theta_l(z)$ and $\Delta q_w(z)$ inside and outside of a cloud, may be formulated as described by Yamada (1979) or Smith (1990) if cloud fraction, $R$, is an available model variable. In the absence of $R$, one can use the "all-or-nothing" assumption in which at any given time a grid volume is considered either fully saturated ($R = 1$) or clear ($R = 0$), yielding

$$\alpha = \begin{cases}\phi_t - b\phi_l & \text{cloudy}\\ \phi_t & \text{clear}\end{cases} \qquad \beta = \begin{cases}\phi_w + a\phi_l & \text{cloudy}\\ \phi_w & \text{clear,}\end{cases}$$

where

$$\phi_t = 1 + 0.609q_w - 1.609q_l$$

$$\phi_l = (1 + 0.609q_w - 3.218q_l)\frac{\langle\theta\rangle L_v}{\langle T\rangle C_p} - 1.609\theta_l$$

$$\phi_w = 0.609\left(\theta_l + \frac{\langle\theta\rangle L_v}{\langle T\rangle C_p}q_l\right)$$

$$a = \left(1 + 0.622\frac{L_v^2}{C_p R_d}\frac{q_{sl}}{T_l^2}\right)^{-1}$$

$$b = 0.622a\frac{\langle T\rangle}{\langle\Theta\rangle}\frac{L_v q_{sl}}{R_d T_l^2}.$$

Here, $T$ is temperature, $T_l$ is liquid water temperature, $q_l$ is liquid water content, $q_{sl}$ is saturation vapor pressure at temperature $T_l$, $L_v$ is latent heat of vaporization, $C_p$ is specific heat at constant pressure, and $R_d$ is the dry gas constant.

As a further modification to B01, we suggest that Eq. (1) should be considered as a necessary, but not a sufficient, condition for a parcel at level $z_p$ to reach the surface. For example, consider the case of a decoupling cloud layer in which a slightly stable layer develops just beneath cloud base and separates a subcloud region that

*Corresponding author address:* Dr. Stephen D. Burk, Marine Meteorology Division, Naval Research Laboratory, 7 Grace Hopper Ave., Monterey, CA 93943-5502.
E-mail: burk@nrlmry.navy.mil

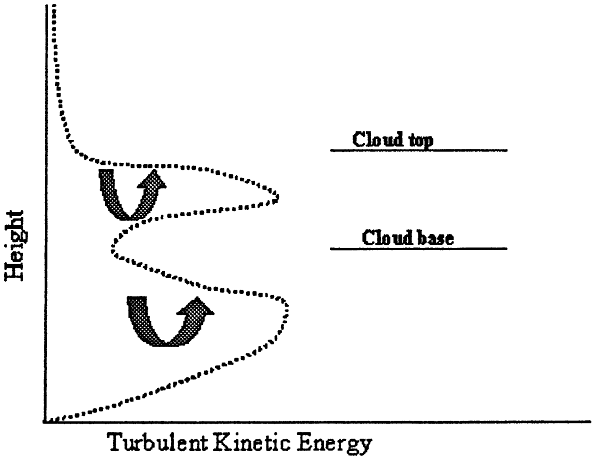


FIG. 1. Schematic TKE distribution in a decoupling cloud-topped boundary layer. Distinct cloud- and subcloud-layer circulations may develop, altering the energetics involved for parcels to reach the surface.

may be driven primarily by surface fluxes from a cloud layer where radiative forcing may dominate (Bretherton and Wyant 1997). The mean TKE from the surface to a level $z_p$ within the cloud layer may satisfy the inequality in Eq. (1); however, the mean TKE from $z_p$ to cloud base may be insufficient to overcome the potential energy of buoyancy associated with the cloud-base stability. In this case the BL may not be filled with large eddies as depicted in Fig. 4 of B01, but rather may conform more to the sketch here in Fig. 1 wherein the subcloud and cloud-layer circulations have become distinct. This situation in which the TKE distribution contains a minimum at cloud base, with local maxima in the subcloud layer and the cloud layer, is common in trade wind and stratocumulus-topped BLs. Thus, we suggest that the criterion,

$$\frac{1}{z_p - z'} \int_{z'}^{z_p} E(z)\, dz \geq \int_{z'}^{z_p} g \frac{\Delta\theta_v(z)}{\Theta_v(z)}\, dz, \qquad (3)$$

must be satisfied in the WGE computation *at all levels $z'$ beneath $z_p$*. Equation (1) then becomes a special case of Eq. (3) where $z' = 0$. Although our two proposed alterations to Eq. (1) are designed to make this expression more physically correct, we recognize that they may not routinely impact the computed gust velocity in a substantial manner.

Finally, we wish to note a potential new application of the B01 technique to computation of a scaling velocity in convective conditions such as occurs with a strongly heated BL during light wind over land. Under these conditions the ‘‘free convection scaling velocity,’’ $w^*$, is often used either to scale turbulence variables or as part of a BL parameterization. This scaling velocity derives from dimensional analysis that yields

$$w^{*3} = B_s z_i, \qquad (4)$$

where $B_s$ is the surface buoyancy flux and $z_i$ the inversion height. The fact that $w^*$ is defined in terms of $B_s$, and yet $w^*$ is often needed to compute $B_s$, can be troublesome.

For cloud-topped BLs, a generalized convective scaling velocity of the form

$$\hat{\omega}^3 = 2.5 \int_0^{z_i} B(z)\, dz \qquad (5)$$

is frequently used (Deardorff 1980) where the factor 2.5 is chosen so that $\hat{\omega} \simeq w^*$ for clear, convectively mixed layers. Note, however, that because $z_i$ is not predicted explicitly in most NWP models, it can be difficult to diagnose $z_i$ [which appears in both Eqs. (4) and (5)] in an accurate and robust manner for the myriad of BL structures that can arise.

Using our modification, Eq. (3), to the B01 WGE one can derive an alternative scaling velocity, here termed $\tilde{\omega}$, in a more physically based manner and this new scaling velocity contains no explicit dependence on the surface flux or $z_i$. As described by Eq. (3), we again find all levels wherein parcels are capable of reaching the surface. But, rather than selecting the maximum wind speed from among these parcels as in the WGE approach, we select the square root of the maximum turbulent velocity variance (maximum standard deviation) at those levels. Thus,

$$\tilde{\omega} = \max[(2E(z))^{1/2}] = \max(q), \qquad (6)$$

*where the max is selected from among levels where parcels are capable of reaching the surface* and $E = q^2/2$. In this application, it is expected to be important that the buoyancy flux in the model TKE equation, as well as $\alpha$ and $\beta$ in Eq. (2), account for fractional cloudiness rather the using the all-or-nothing assumption that was discussed earlier.

To better discriminate coupled from decoupled BLs, one may compute $\langle\tilde{\omega}\rangle = \alpha\tilde{\omega}_{\mathrm{cld}} + (1 - \varepsilon)\tilde{\omega}_{\mathrm{clr}}$, where, of parcels capable of reaching the surface according to Eq. (3), $\varepsilon$ is the fraction coming from the cloud layer, $(1 - \varepsilon)$ is the fraction from the subcloud layer, $\tilde{\omega}_{\mathrm{cld}}$ is the maximum in-cloud turbulent velocity, and $\tilde{\omega}_{\mathrm{clr}}$ is the maximum subcloud turbulent velocity. When the BL is fully decoupled $\varepsilon = 0$ and $\langle\tilde{\omega}\rangle$ is solely determined by subcloud TKE. The generalized convective velocity scale, $\hat{\omega}$ [Eq. (5)], with its integration over the full BL depth, cannot clearly distinguish coupled from decoupled BLs, whereas $\langle\tilde{\omega}\rangle$ as described here can, to some extent, address such regime transitions.

Observations of a well-mixed *clear* convective boundary layer (e.g., Caughey and Palmer 1979) show that the sum of the turbulent velocity variances, $q^2$, attains a maximum value of $\sim w^{*2}$ near mid-BL, thus, our choice of the maximum in Eq. (6). That is, $\tilde{\omega} = \tilde{\omega}_{\mathrm{clr}} \sim w^*$. Therefore, in such conditions, we do not expect $\tilde{\omega}$ to differ greatly from $w^*$, but $\tilde{\omega}$ contains the advantage of being physically based as well as not having a direct dependence on $z_i$ or surface flux. We have already outlined the potential advantages of $\tilde{\omega}$ for a cloud-topped BL.

We have coded the B01 WGE approach as well as our two modifications for testing in the navy's Coupled Ocean–Atmosphere Mesoscale Prediction System (COAMPS). Working with colleagues, we intend to perform extensive tests comparing model forecasts with surface station wind gust observations.[1] Also, we intend to conduct comparative tests of the alternative choices for convective scaling velocity, $w*$, $\hat{\omega}$, and $\tilde{\omega}$.

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[1] When we originally wrote this comment, we did not anticipate that initial publication of tests of our proposed new formulations would occur in a group reply by Brasseur et al. (2002).

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