Should light wind and windy nights have the same temperature trends at individual levels even if the boundary layer averaged heat content change is the same?

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[1] Long-term climate trends of surface air temperature should not be expected to have the same trends for light wind and stronger wind nights, even if the trends in the boundary layer heat fluxes were the same. Parker (2004) segmented observed surface temperature data into lighter and stronger wind terciles in order to assess whether the reported large-scale global-averaged temperature increases are attributable to urban warming. We conclude, however, that trends at an individual height depend on wind speed, thermodynamic stability, aerodynamic roughness, and the vertical gradient of absolute humidity. We present an analysis to illustrate why temperature values at specific levels will depend on wind speed, and with the same boundary layer heat content change, trends in temperature should be expected to be different at every height near the surface when the winds are light, as well as different between light wind and stronger wind nights. This introduces a complexity into the assessment of long-term surface temperature trends that has not been previously recognized. **Citation:** Pielke, R. A., Sr., and T. Matsui (2005), Should light wind and windy nights have the same temperature trends at individual levels even if the boundary layer averaged heat content change is the same?, *Geophys. Res. Lett.*, *32*, L21813, doi:10.1029/2005GL024407.

1. Introduction

[2] *Parker* [2004] published an interesting study in which he segmented observed surface temperature data into "calm" (defined as the lower tercile of daily-averaged wind speeds) and "windy" (defined as the upper tercile of daily-averaged wind speeds) in order to assess whether the reported large-scale global-averaged temperature increases are attributable to urban warming. Our paper, however, questions, whether trends of surface layer air temperature should even be expected to have the same trends for these different sets of days.

2. Background

[3] *Parker* [2004] focused on minimum temperatures that are most likely to occur at night before sunrise. At this time of the night, it is well understood that the temperature change with height in the lowest tens of meters can be quite large, particularly in light wind, clear sky conditions [e.g., *Stull*, 1988; *Oke*, 1987]. For relatively windy night-time conditions, surface similarity theory has been developed which can be used to describe the vertical temperature lapse rate in the lowest few tens of meters. *Pielke* [1984, Figure 7-4], for example, illustrates the vertical lapse rate as a function of the intensity of mechanical generation and convective suppression of turbulence. For strong winds, the vertical lapse rate becomes nearly adiabatic (a temperature lapse rate of 0.1°C per 10 meters). For light wind conditions, long-wave radiative flux divergence becomes dominant [*Gopalakrishnan et al.*, 1998] and the lapse rates can become quite large (e.g., 10°C per 10 m or even larger).

[4] Thus, if the upper tercile of wind conditions, as defined by *Parker* [2004], is well represented by an adiabatic lapse rate, a temperature at 1 m, for instance, would be essentially the same as at 2 m. However, when the large lapse rates typical of lighter wind conditions occur, a temperature at 1 m would be significantly different at 2 m.

[5] This effect can be illustrated as follows:

$$\begin{aligned} J_1 &= C_p T_1 + L q_1 \\ J_2 &= C_p T_2 + L q_1 \end{aligned} \tag{1}$$

where J_1 and J_2 (J kg^{-1}) represent the heat at a level in the surface layer (which is on the order of 10 m thick) for the "windy" and the "calm" terciles, respectively. C_p (J kg^{-1} K^{-1}) is the specific heat of air at constant pressure, T is temperature (K), L (J kg^{-1}) is latent heat of vaporization, and q (kg kg^{-1}) is specific humidity. Heat is measured by moist enthalpy as shown by *Pielke et al.* [2004].

[6] In the surface layer, the vertical temperature and absolute humidity gradient associated with turbulent mixing can be estimated as

$$\frac{\partial T_1}{\partial z} \approx 0 \text{ and } \frac{\partial q_1}{\partial z} \approx 0 \left(\text{and thus } \frac{\partial J_1}{\partial z} \approx 0\right) \tag{2}$$

for the strongest wind tercile (what *Parker* [2004] refers to as "windy" nights).

[7] For the "calm nights" (but when mechanical mixing by turbulence is still dominant) using the formula given by *Pielke* [1984, p. 153]

$$\begin{aligned} \frac{\partial T_2}{\partial z} &= \frac{T^* \cdot (0.74 + 4.7 \cdot z/L)}{k \cdot z} \\ \frac{\partial q_2}{\partial z} &= \frac{q^* \cdot (0.74 + 4.7 \cdot z/L)}{k \cdot z} \\ \frac{\partial J_2}{\partial z} &= \frac{(C_p T^* + L q^*) \cdot (0.74 + 4.7 \cdot z/L)}{k \cdot z} \end{aligned} \tag{3}$$

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Here T^* is a scaling temperature, q^* is a scaling absolute humidity value, k is the von Karmen constant, and z/L is a measure of the stability with L referred to as the Monin length and z is the height above the ground.

[8] Equations (2) and (3) illustrate that the vertical profile of temperature, water vapor and heat content within the surface layer are expected, in general, to be different between the two terciles of data. Thus the observation by *Parker* [2004] that the trends of temperature at a single level are the same on "calm" and "windy" nights requires that these profiles are invariant over time and that the temperature change is a constant with altitude in the surface layer. The analysis in the next section examines quantitatively the behavior of vertical temperature trends as related to changes over time in the cooling rate of the nocturnal boundary layer (such as due to increases in the atmospheric concentrations of the greenhouse gases, carbon dioxide and methane, over the last several decades, such that we expect long-term changes in the cooling rate of the nocturnal stable boundary layer).

3. Analysis

[9] To illustrate the response of the temperature profile to nocturnal cooling, we present the following analysis by using an idealized model for the potential temperature profile. For a continuously-turbulent stable clear night boundary layer over a flat surface, the potential temperature profile can be approximated by [*Stull*, 1988, equation 12.1.3.e].

$$\Delta\theta(z) = \theta(z) - \theta(z_{nrl}) = \Delta\theta_s \cdot e^{-z/H_e}$$

where $\Delta\theta(z)$ is stable boundary layer (SBL) strength, defined by the potential temperature difference between the air at height $z(m)$ and the air in the statically neutral residual layer $z_{nrl}(m)$ (which is a remnant from the previous daytime well-mixed boundary layer). H_e is the scale height for the exponential curve [*Stull*, 2000, equation 4.4];

$$H_e \approx a(V_{RL})^{0.75} t^{0.5}$$

where, for our example, we chose $a = 0.15$ for flow over a flat prairie (a is in units of $m^{1/4} \cdot s^{1/4}$), V_{RL} is wind speed in the residual layer (m s^{-1}), and t is the cumulative time in seconds from the time the heat flux divergence becomes negative.

[10] In our analysis, we assume that the wind speeds in the residual layer (e.g., 200 m to 500 m) are the same as in the near surface layer (e.g., 2 m to 10 m), that the equations are valid in the near surface layer, and that the findings based on this description of the boundary layer are applicable to the minimum temperature, $T_{\min}$ with respect to profile differences and the trends. The assumption that $T_{\min}$ and potential temperature vary in nearly the same way with height is justified by the relationship between theta and temperature [e.g., see *Pielke*, 2002, equations 4-17 and 4-19].

[11] $\Delta\theta_s$ is $\Delta\theta(z)$ at the surface, and parameterized as by *Stull* [2000, equation 4.5]

$$\Delta\theta_s = \frac{Q_{AK}}{H_e}$$

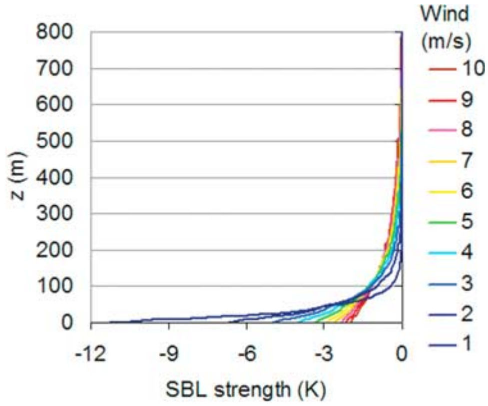

Figure 1. $\Delta\theta(z)$ (SBL strength) profile in different wind conditions for cases of -10 W m^{-2} constant cooling rate over night.

Q_{AK} is the kinematic form for the cumulative heating (in units of K m) where

$$Q_{AK} = \frac{F}{\rho_{air} C_p} t$$

where F is an assumed constant heat flux divergence at night (W m^{-2}), ρ_{air} is the density of air (Kg m^{-3}), and C_p is specific heat of air (J kg^{-1} K^{-1}).

[12] With the above set of equations, we estimate the potential temperature profile at the end of a 12-hour night for different wind cases (10 m s^{-1} to 1 m s^{-1}), with a given constant heat loss: $F = -10$ W m^{-2}, $\rho_{air} \cdot C_p = 1231$(J m^{-3} K^{-1}) at sea level, and $t = 43200$ s. Note that the near-surface wind and wind in the residual layer become similar at night.

[13] In the daytime, the potential temperature is higher near the surface due to solar insolation, while after sunset, long-wave cooling gradually lowers the near-surface temperature as the boundary layer stabilizes.

[14] The $\Delta\theta(z)$ profiles for the different (1$\sim$10 m s^{-1}) wind conditions are computed at every 10 m vertical increment (Figure 1). The windy cases are warmer near the ground, while the cooling extends over a greater depth of the boundary layer than for the calm cases. This means that the potential temperature lapse rate in the surface layer is greater in the low-wind cases than in the high-wind cases, while the potential temperature lapse rate in the statically neutral residual layer is near zero in the low-wind cases.

[15] Figure 2 shows the lapse rate of potential temperature in the surface layer for the different wind cases and also for different constant heat fluxes (-50, -40, -30, -20, and -10 W m^{-2}). The lapse rate of potential temperature is simply defined as

$$\frac{\partial\theta}{\partial z} = \frac{\theta(z_0) - \theta(z_{10})}{z_{10}},$$

where $\theta(z_0)$ is the potential temperature at the surface (in units of degrees K), $\theta(z_{10})$ is the potential temperature at 10 m AGL (in units of degrees K), and z_{10} is 10 m. As discussed earlier, the lapse rate in the light wind cases is more negative; e.g., more than -1 K m^{-1} in the 1 m s^{-1} wind case with -40 and -50 W m^{-2} cooling rate. The larger constant cooling rate tends to have a slightly more

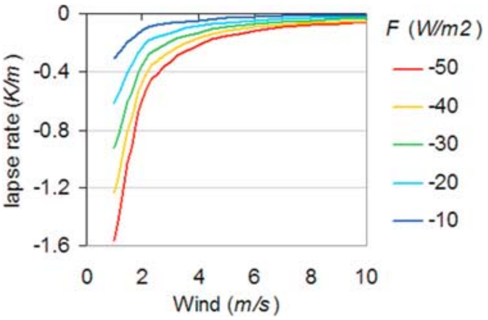

Figure 2. Lapse rate of potential temperature profile for the lowest 0 $\sim$ 10 m for different wind conditions and five different values of the flux divergence.

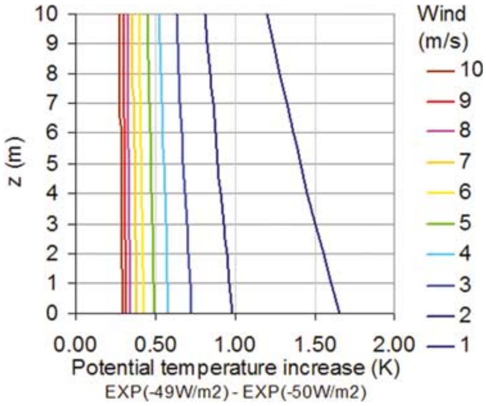

Figure 3. Potential temperature increase at different levels from the experiment with -49 W m^{-2} cooling to the experiment with -50 W m^{-2} cooling.

negative lapse rate; however, the wind more strongly controls the lapse rate rather than the constant heat flux.

[16] This example shows that the different wind condition result in different profiles of potential temperature (and thus profiles of the temperature), e.g., even though same cooling rate is added in the boundary layer, the lighter winds have a lower near-surface temperature due to the more inefficient vertical mixing and thus a larger lapse rate. The influence of windiness to produce different vertical profiles of absolute humidity will result in even greater differences in the vertical profiles of moist enthalpy.

4. Relation to Long-Term Surface Layer Temperature Trends

[17] Over the last several decades, the atmospheric concentration of carbon dioxide has changed. As shown, for example, by *Eastman et al.* [2001] the addition of carbon dioxide to the atmosphere alters the downwelling long-wave radiative fluxes such that, in their idealized sensitivity experiments, the nighttime minimum temperature was increased, although the daytime maximum (and thus the temperature above the surface in the "residual layer" was essentially unchanged). In the results reported in *Eastman et al.*, a doubling of the carbon dioxide concentrations in the model resulted in a growing season, central Great Plains area-averaged increase of minimum temperature about 0.1°C. While this experiment was idealized and used a doubling of carbon dioxide, it raises the issue that we should expect long-term changes in the cooling rate of the boundary layer at night over time as the atmospheric concentration of greenhouse gases, including water vapor, changes. This change of heat loss would occur even without a change in the temperatures above the surface layer. Changes in the lapse rate, however, above the boundary layer, if they occur, would also alter the cooling rate.

[18] To illustrate the effect of a change in cooling rate on the temperatures in the surface layer, the change in potential temperatures at individual layers for the lower 10 m due to a reduction in the heat flux loss by 1 W m^{-2} is presented in Figure 3 and Table 1. This reduction in loss could be for any reason as mentioned above, but one reason we should expect this change over the last several decades is due to the observed increase in the atmospheric concentration of carbon dioxide. We chose a value of 1 W m^{-2} to illustrate the effect of even a small change in the boundary layer heat flux.

[19] As seen in Figure 3 and Table 1, for the windy cases, the temperature increase due to the reduced boundary layer cooling is nearly uniform with altitude. However, there are significant changes with height in the surface layer, and the temperature increase is higher with the lighter wind cases. While the heat change averaged throughout the boundary layer is the same in each case, the temperature change that results in the surface layer is a function of the wind speed.

[20] Aerodynamic roughness changes over time also alter the surface layer temperature profile. This occurs due to the different mechanically-forced turbulent mixing over the rougher surfaces with a given wind speed. Even with the same trend of boundary layer heat flux, there would be different trends at specific levels within that layer when the aerodynamic roughness change effect is included.

5. Conclusions

[21] This paper does not address the actual trends in surface layer heat content over time. However, it does indicate that if the nocturnal boundary layer heat fluxes change over time, the trends of temperature under light winds in the surface layer will be a function of height, and that the same trends of temperature will not occur in the surface layer on windy and light wind nights.

[22] *Parker*'s [2004] conclusions, therefore, need further analysis and interpretation before they can be used to conclude whether or not there is an influence of urban

Table 1. Tabulated Version of Figure 3[a]

	Wind Speed, m s^{-1}									
Z, m	10	9	8	7	6	5	4	3	2	1
10	0.28	0.30	0.33	0.36	0.40	0.45	0.52	0.63	0.81	1.20
9	0.28	0.30	0.33	0.36	0.40	0.45	0.53	0.64	0.83	1.24
8	0.28	0.30	0.33	0.36	0.40	0.46	0.53	0.65	0.85	1.28
7	0.28	0.31	0.33	0.37	0.41	0.46	0.54	0.66	0.86	1.32
6	0.28	0.31	0.33	0.37	0.41	0.47	0.55	0.67	0.88	1.37
5	0.29	0.31	0.34	0.37	0.41	0.47	0.55	0.68	0.89	1.41
4	0.29	0.31	0.34	0.37	0.42	0.48	0.56	0.69	0.91	1.46
3	0.29	0.31	0.34	0.38	0.42	0.48	0.57	0.70	0.93	1.50
2	0.29	0.31	0.34	0.38	0.42	0.49	0.57	0.71	0.95	1.55
1	0.29	0.32	0.35	0.38	0.43	0.49	0.58	0.72	0.97	1.60
0	0.29	0.32	0.35	0.38	0.43	0.50	0.59	0.73	0.98	1.66

[a]Potential temperature increase at different levels from the experiment with -49 W m^{-2} cooling to the experiment with -50 W m^{-2} cooling.

warming on the large-scale temperature trends. More broadly, the issue of the influence of winds on the vertical temperature stratification with respect to the temperature trends raises the issue as what is actually meant by the term ''surface temperature trend.'' Along with the issues of surface temperature changes as related to surface moist enthalpy changes [*Pielke et al.*, 2004] and microclimate station exposure changes [*Davey and Pielke*, 2005], the reported regionally- and globally-averaged surface temperatures trends have unresolved uncertainties.

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