A Simplified Model for Intermittent Turbulence in the Nocturnal Boundary Layer

FELIPE D. COSTA, OTÁVIO C. ACEVEDO, JOSÉ C. M. MOMBACH, AND GERVÁSIO A. DEGRAZIA

Departamento de Física, Universidade Federal de Santa Maria, Santa Maria, Rio Grande do Sul, Brazil

(Manuscript received 26 August 2010, in final form 16 February 2011)

ABSTRACT

A model for the exchange between the surface and the atmosphere under stable conditions is proposed. It is based on the classical scheme first suggested by Blackadar and comprises prognostic equations for the wind components and air and ground temperature. The main difference from previous works consists in the fact that the turbulent intensity is determined by a prognostic equation for turbulent kinetic energy (TKE), rather than by using stability functions that arbitrarily relate it to atmospheric stability. Results show that the model reproduces the condition of connection and disconnection between the surface and upper levels. Furthermore, it leads to periodic turbulence bursts when one level within the stable boundary layer (SBL) is considered and the use of additional levels increasingly leads to more complex solutions, characterizing the occurrence of global intermittency. Such turbulence bursts occur in the disconnected state and cause large fluctuations of the variables near the surface. The boundary layer height plays a role in the sense that for the same geostrophic winds, connection is favored for shallower layers. Although playing a role in the intermittency characteristics, soil type is not determinant to their existence, as the bursts occur even for very high values of heat capacity. Vertical profiles for both the intermittent and connected state are analyzed and in general agree with observations. It is shown that, near the surface, weak turbulence bursts favor the exchange between the air and the cooler ground, leading to a local temperature decrease, while stronger events that mix the air deeper in the SBL cause an average warming tendency. An opposite pattern occurs at the upper SBL. Intermittency is favored over a range of low geostrophic winds and clear skies, in agreement with previous suggestions. The vertical structure of the intermittent events is analyzed, and it shown that they are generated at the surface by a local shear increase above a threshold, propagating upward through the turbulence transfer term in the TKE equation. It is proposed that such events constitute a natural characteristic of the disconnected SBL, which occurs along with low large-scale winds and clear skies.

1. Introduction

The characterization of the atmospheric state near the surface is particularly complex in the very stable nocturnal boundary layer. Such conditions happen when large longwave radiative loss occurs during clear-sky nights. If, additionally, there are no major mechanisms to accelerate the winds near the surface, the mechanical turbulence generation may be almost entirely counteracted by the destruction caused by the thermal stable stratification. Although the theory indicates that no turbulence must exist under such conditions, the occurrence of localized bursts is often observed, a process known as global intermittency (Mahrt 1999). It is typically characterized by the succession of turbulent and calm periods throughout the night (Fig. 1). Observations show that, in general, the turbulent events during global intermittency are non-periodic with variable amplitudes and periods, suggesting a chaotic organization.

Nappo (1991) characterized the turbulence bursts, showing that most of them lasted for 5–20 min and that they are a consistent feature of the stable PBL over all types of terrain. Similar events have been analyzed by Coulter and Doran (2002), who noted that they may occur simultaneously at sites 1 km apart from each other, therefore having a large spatial coherence. Neither Nappo nor Coulter and Doran focused on the physical causes of the small-scale events that they analyzed. Sun et al. (2002), on the other hand, showed that a large intermittent event that caused increased turbulence for a couple of hours could be attributed to a density current passage. Other mechanisms that may lead to intermittent behavior are described by Sun et al. (2004) as Kelvin–Helmholtz...
instabilities, low-level jets, and gravity waves. In all of these cases, the events are transported downward toward the surface from an upper-level instability. A different type of intermittent event, originating at the surface layer by its interaction with a thin vegetated layer, has been simulated by van de Wiel et al. (2002a). Poulos and Burns (2003) associated the bursts with atmospheric stability, proposing stability functions that incorporate the occurrence of intermittency.

Blackadar (1979) proposed a classical model for the interaction between the ground surface and the atmosphere. In such a scheme, prognostic equations for the wind components and air and ground temperature are solved. The ground temperature is forced by a radiative budget, while the other variables depend on external forcings, as well as on the turbulent flux convergence over a layer close to the surface. Therefore, the quality of this and similar schemes relies strongly on its ability to properly represent the turbulent fluxes and their dependency on the prognostic variables. A popular solution to this problem is the use of stability functions, which directly relate turbulence intensity to a measure of atmospheric stability, most frequently the gradient Richardson number (Louis 1979; Beljaars and Holtslag 1991; Delage 1997; Poulos and Burns 2003).

McNider et al. (1995) showed that Blackadar’s model presents high sensitivity to the initial conditions in the sense that a small difference in the external forcing may lead to a completely different solution in terms of the connection state between the ground surface and the atmosphere. However, both in the connected and in the disconnected state, the solutions are constant in time after a transient period, showing that such a model is not able to reproduce the observed lack of order characteristic of intermittency. Although observations confirm the large distinction between the connected and disconnected condition (Acevedo and Fitzjarrald 2003), they also show a much larger degree of complexity. In fact, the mere initiation of turbulence after its total decay is a difficult task for the numerical schemes that simulate the stable atmospheric surface layer. Such a task has been achieved by van de Wiel et al. (2002a) by considering a thin vegetated layer with a low heat capacity, but the global intermittency simulated in this case is periodic, in disagreement with most observations. Similar periodic bursts of turbulence have been simulated by Revelle (1993). Cuxart and Jiménez (2007) simulated intermittent bursts induced by a low-level jet using a large-eddy simulation (LES) model forced by observed conditions at the surface. Similar results were achieved by Zhou and Chow (2010). Boing et al. (2010) used a direct numerical simulation (DNS) model to simulate intermittent events caused by the shear instability induced by a canopy. A simplified model that simulates nonperiodic turbulence bursts of variable amplitude in the nocturnal ABL is still missing in the literature.

Here, we hypothesize that the limitation of Blackadar’s model in reproducing intermittent turbulent bursts is...
directly caused by the use of a stability function. From a dynamical system point of view, such use reduces the number of degrees of freedom of the system, as it imposes a very simplistic relationship between turbulent intensity and atmospheric stability, two major variables of the problem. Dynamically, such reduction severely dampens the possibility of complex behavior to arise. In the present study, we propose using a turbulent kinetic energy (TKE) prognostic equation, as in the Blackadar model, to avoid the necessity of a stability function. For simplicity, variables such as the exchange coefficients and turbulence dissipation rate are parameterized based on dimensional arguments, and this is enough to lead the model to reproduce turbulent bursts.

It is important to notice that this is a simplified model, whose primary purpose is to understand how the interaction between the surface and the lowest layers of the atmosphere is capable of causing events of intermittent turbulence. Once this is achieved, the results allow one to describe the mechanisms by which these events are generated and propagated and the external conditions that favor their occurrence.

2. The model

a. Basic equations

To describe the flow in the nocturnal atmospheric boundary layer we consider the prognostic equations used by Blackadar (1979) and McNider et al. (1995), among others:

\[
\begin{align*}
\frac{\partial u}{\partial t} &= f(v - v_G) - \frac{\partial(u'w')}{\partial z}, \\
\frac{\partial v}{\partial t} &= f(u_G - u) - \frac{\partial(v'w')}{\partial z}, \\
\frac{\partial \theta}{\partial t} &= -\frac{\partial(w'\theta')}{\partial z},
\end{align*}
\]

(1-3)

and

\[
\frac{\partial \theta_s}{\partial t} = \frac{1}{C_g} \left( I_i - \sigma \theta_g^4 - H_0 \right) - k_m (\theta_g - \theta_m),
\]

(4)

where \(u, v\) are the east–west, north–south velocity components and \(\theta\) and \(\theta_g\) the potential temperature and ground temperature, respectively. Like McNider et al. (1995), we do not consider the radiative flux divergence above the surface, and the atmospheric radiative cooling is taken to be zero in Eq. (3).

The constants are the Coriolis parameter \(f\), \(u_G\) and \(v_G\), the zonal and meridional horizontal components of geostrophic wind at the boundary layer top; \(C_g\), the thermal capacity of a slab of unit area; \(k_m\), the heat transfer coefficient; and \(\theta_m\), the temperature of the substrate (Blackadar 1979). In Eq. (4), \(I_i\) is the long-wave back radiation from the atmosphere, which, following Staley and Jurica (1972), depends on cloud fraction \(Q_c\), specific humidity \(Q_a\) at a reference height, and the temperature \(\theta\) at the boundary layer top: \(I_i = \sigma [Q_c + 0.67(1 - Q_c)(1670Q_a)^{0.08}]\theta^4\). Also, \(H_0 = \rho C_p \frac{w\theta'}{g}\) is the surface sensible heat flux, whose parameterization is described at a later section. In these expressions, \(\sigma\) is the Stefan–Boltzmann constant, \(\rho\) is the air density, and \(C_p\) is the specific heat of air at constant pressure.

b. Ground parameterization

The ground parameterization is an important aspect to determine the nature and behavior of the planetary boundary layer (Zhang and Anthes 1982; van de Wiel et al. 2002a). In Eq. (4) we use a slab model to compute the ground potential temperature (Blackadar 1979; Zhang and Anthes 1982; McNider et al. 1995).

Here we consider that \(\theta_g\) is the same as the uniform temperature of the slab (Blackadar 1976). The heat capacity per unit area of the slab \((J K m^{-2})\) depends on the thermal conductivity \(\lambda\), earth angular frequency \(\omega\), and heat capacity per unit soil volume \(C_s = c_s \rho_s\), where \(c_s\) and \(\rho_s\) are respectively the soil specific heat and density (Blackadar 1976, 1979):

\[
C_g = 0.95 \left( \frac{\lambda C_s}{2\omega} \right)^{1/2}.
\]

For the substrate heat flux, the heat transfer coefficient is expressed as \(k_m = 1.18\omega\) (Blackadar 1976) and \(\theta_m\) is the mean temperature of the substrate, whose value is the mean surface air temperature during the previous 24 h.

c. Turbulence closure

The classical parameterization for the turbulent fluxes in Eqs. (1)–(3) is K theory (Blackadar 1976; Zhang and Anthes 1982; McNider et al. 1995; van de Wiel et al. 2002a, and many others) in which the fluxes are related to local gradients and an eddy diffusivity \(K\). Generally these exchange coefficients are parameterized with the use of arbitrarily chosen stability functions, which depend on the Richardson number. In our model, rather than using stability functions, we follow Duynkerke (1988) and evaluate the momentum exchange coefficient through a relationship between turbulent kinetic energy \(E\) and turbulent dissipation rate \(\epsilon\) as \(K_m = c_{\epsilon\mu}E^2/\epsilon\), where \(c_{\epsilon\mu}\) is a constant equal to 0.033. The exchange coefficients for heat and momentum are directly related by the turbulent Prandtl number \(Pr\) as \(K_m = Pr K_h\). TKE
is evaluated using a prognostic equation. TKE models are generally composed by prognostic equations for the mean variables [such as Eqs. (1)–(3)] and two more prognostic equations for TKE and $\epsilon$ (Duynkerke 1988; Cuxart et al. 2006). In the present study, we use a prognostic equation for TKE and the viscous dissipation of TKE is evaluated by an expression based on dimensional arguments. The TKE prognostic equation is

$$\frac{\partial E}{\partial t} = -u'w' \frac{\partial u}{\partial z} - v'w' \frac{\partial v}{\partial z} + \frac{g}{\Theta} w' \theta' - \frac{\partial}{\partial z} \left[ (w' E) + \frac{p' w'}{\rho_0} \right] - \epsilon, \quad (5)$$

where $g$ is the acceleration due to gravity, $\Theta$ the reference temperature, and $p$ pressure. The terms of Eq. (5) are mechanical production of turbulence (first and second rhs terms), thermal production or destruction of turbulence (third rhs term), vertical turbulent transport of TKE (fourth rhs term), and the viscous dissipation of TKE. On average, for stable conditions, TKE is directly proportional to the square of the friction velocity (Panofsky and Dutton 1984). According to Duynkerke (1988), the average proportionality coefficient from a list of estimates given by Panofsky and Dutton (1984) is 5.5, so we assume $E = 5.5u_0^2$. Here, it is important to notice that we are considering $u_0$ as the square root of the local momentum flux at any given height. We use the symbol $u_{\text{ref}}$ for its surface value.

The mechanical production terms can be parameterized as $Su_0^2$, where the wind shear $S = [(\partial u/\partial z)^2 + (\partial v/\partial z)^2]^{1/2}$. Assuming, as a simplification, that the Prandtl number equals unity, the thermal destruction term is written as $-\text{Re} Su_0^2$, where the local Richardson number $\text{Re} = (g/\Theta) (\partial \theta/\partial z)/S^2$. The transport term is modeled as (Duynkerke 1988)

$$- \frac{\partial}{\partial z} \left[ (w' E) + \frac{p' w'}{\rho_0} \right] = \frac{K_E}{\sigma_E} \frac{\partial E}{\partial z},$$

in which $\sigma_E$ is the turbulent Prandtl number for $E$. Duynkerke stated that $\sigma_E$ has no fixed value and used $\sigma_E = 1$. Weng and Taylor (2003) used the same value, and Baas et al. (2008) used one equivalent to $\sigma_E = 0.5$. Assuming that the first term in the expression above can be parameterized as $-\text{Re} Su_0^2$, the constant $\sigma_E$ must be smaller than unity, if both terms have the same sign, and larger than unity, otherwise. Evidence from laboratory (Deardorff and Willis 1985) and LES studies (Moeng and Wyngaard 1989; Moeng et al. 2004) show that the two terms tend to have opposite signs for the convective boundary layer. Under stable conditions, the amount of evidence is not as large, but Kosović and Curry (2000) indicate that, despite being small in magnitude, the two types of turbulent transport have opposite signs near the surface: the same conclusion was found by Puhales et al. (2010). Based on this evidence, we choose $\sigma_E = 2.5$—this is necessary for simulating intermittent turbulence. Using the value assumed by Duynkerke (1988), for instance, accelerates the turbulence diffusion, smoothing out the intermittent bursts.

The viscous dissipation term in Eq. (5) can be written as $\epsilon = c_2 (u_0^3/l)$, where the mixing length $l = \kappa z$, and $\kappa$ is the von Kármán constant. According to Cuxart et al. (2006), values ranging from 0.08 to 0.7 have been used for $c_2$. We assume $c_2 = 0.18$; tests indicated that the solution patterns are qualitatively similar for 0.08 $< c_2 < 0.3$, becoming progressively different as $c_2$ increases beyond 0.3, although intermittency always occurs for the entire range of values suggest by Cuxart et al. (2006). Accordingly, the exchange coefficients can be written as $K_m = K_E = K u_0 z$. The dependence of the exchange coefficients on stability is not explicitly considered but arises naturally as a consequence of the TKE budget. Thus, Eq. (5) can be rewritten as

$$\frac{\partial E}{\partial t} = Su_0^2 - \text{Re} Su_0^2 + \frac{\partial}{\partial z} \left[ \frac{K_E}{\sigma_E} \frac{\partial E}{\partial z} \right] - c_2 \frac{u_0^3}{l}. \quad (6)$$

The turbulent fluxes in Eqs. (1)–(3) are parameterized as $-(u' w') = u_0^2 \cos(\psi)$; $-(v' w') = u_0^2 \sin(\psi)$, where $\psi = \arctan(v/u)$; and $-(w' \theta') = u_0 \theta_s$, with $\theta_s = (K_D u_0)(\partial \theta/\partial z)$ so that $H_0 = -pc_p u_{\text{ref}} \theta_{\text{ref}}$.

**d. Boundary conditions and model discretization**

The sole prognostic variable that is not dependent on height is the temperature of the surface $\theta_s(t, i)$, which is evaluated at $z = 0$. All others are time and height dependent: $u(t, z), v(t, z), \theta(t, z)$ and $E(t, z)$.

In our model, the stable boundary layer (SBL) boundaries are the layer top $h$ and the ground surface ($z = 0$). Between these two limits, we consider $n$ levels with the first one fixed at $z = 5$ m and the others equally spaced between the first level and $h$. The prognostic equations for the wind components and potential temperature are calculated at these levels. However, for such evaluation, it is necessary to estimate the turbulent flux divergence, where each flux depends on TKE. Therefore, the prognostic equation for TKE is evaluated at intermediate levels $Z$ between the main levels $z$. These are defined as $Z = (z_i + z_{i-1})/2$ so that $z_1 = 5$ m and $Z_1 = 2.5$ m. It is important to notice that the index $i$ refers to any arbitrary level and that the corresponding intermediate level is below it, so that $Z_1$ is below $z_1$, and so on. The boundary
layer top is defined as \( z_{n+1} = h \) so that the highest intermediate level is located between \( i = n \) and \( i = n + 1 \): \( Z_{n+1} = (z_{n+1} + z_n)/2 \). At the boundary layer top, the variables are assumed as constants \( u(t, h) = u_G, v(t, h) = v_G, \theta(t, z) = \Theta \), where \( \Theta = 300 \) K.

At the first level, the initial condition for the zonal wind component is \( u(0, z_1) = 0.1 \text{ m s}^{-1} \), and a linear variation is assumed between this level and the top. For the meridional component \( v \), the initial condition at all levels is \( v(t, h) = v_G = 0 \). The initial value for ground temperature and potential temperature at all levels is the reference temperature: \( \theta_s(0) = \Theta \) and for TKE there is an initial value at all intermediate levels, which is also assumed to be the minimum possible TKE value \( E(0, Z_i) = 0.005 \text{ m}^2 \text{s}^{-2} \).

Therefore, the full set of prognostic equations used is

\[
\frac{\partial u}{\partial t} = f(u, v_G) + \frac{1}{Z_{i+1} - Z_i} \left[ u_{i+1}^2 \cos(\psi_{i+1}) - u_i^2 \cos(\psi_i) \right],
\]

(7)

\[
\frac{\partial v}{\partial t} = f(u_G - u) + \frac{1}{Z_{i+1} - Z_i} \left[ u_{i+1}^2 \sin(\psi_{i+1}) - u_i^2 \sin(\psi_i) \right],
\]

(8)

\[
\frac{\partial \theta}{\partial t} = \frac{1}{Z_{i+1} - Z_i} \left( u_{i+1} \theta_{i+1} - u_i \theta_i \right),
\]

(9)

\[
\frac{\partial \bar{g}}{\partial t} = \frac{1}{C_g} \left( I_1 - \sigma \bar{g}^4 - H_0 \right) - k_m (\theta_G - \theta_m),
\]

(10)

and

\[
\frac{\partial E_i}{\partial t} = S_i \bar{u}_{i+1}^2 - R_i S_i \bar{u}_i^2 + \frac{T_i - T_{i-1}}{z_i - z_{i-1}} - c_s u_i^2.
\]

(11)

The temperature of the substrate \( \theta_m \) is taken as \( \theta_m = 285 \text{ K} \). In Eq. (11), \( T_i \) is the vertical turbulent flux of TKE. It is defined at the main levels, so that it is parameterized as

\[
T_i = \frac{5.5}{\sigma_E} \left( \frac{K_{i+1} + K_i}{2} \right) \frac{(u_{i+1}^2 - u_i^2)}{Z_{i+1} - Z_i}.
\]

The model assumes that the SBL top is not a rigid wall, such as the surface. Therefore, considering that the friction velocity vanishes at \( h \), the upper boundary condition for \( T_i \) is

\[
T_i = \frac{5.5}{\sigma_E} \frac{K_i (-u_i^2)}{2} \frac{h - Z_i}{h}.
\]

Both \( Ri \) and \( S \) are evaluated at intermediate levels. Following McNider et al. (1995), the Richardson number at the lowest intermediate level is

\[
Ri_1 = \frac{g}{\Theta} \sqrt{\frac{z_i}{z_0}} \frac{(\theta_1 - \theta)}{\left( u_1^2 + u_1^2 \right)}. \]

At the other intermediate levels,

\[
Ri_i = \frac{g(z_i - z_{i-1})}{\Theta} \frac{(\theta_i - \theta_{i-1})}{(u_i - u_{i-1})^2 + (v_i - v_{i-1})^2}.
\]

Finally, the wind shear at the intermediate levels is

\[
S_i = \frac{(u_i - u_{i-1})^2 + (v_i - v_{i-1})^2}{z_i - z_{i-1}}^{1/2}.
\]

The model is integrated using a time step of 0.01 s. Tests performed with time steps of 0.001 s, and smaller, provided identical results, indicating numerical stability of simulations.

3. Results

The model presented here is based on the scheme proposed by Blackadar (1979) and also used by Revelle (1993), McNider et al. (1995), and van de Wiel et al. (2002a) for a single SBL. In all of these studies, the models provide appreciably different results between the two distinct states of connection or disconnection between the surface and upper boundary layer. In the connected state, the equilibrium values of variables such as mean wind and air temperature are distinctively larger than in the disconnected condition. Shi et al. (2005) demonstrated that the scheme behaves similarly when multiple layers are considered within the SBL. Our model with a single level within the SBL shows a qualitative resemblance to the results obtained by van de Wiel et al. (2002a) with periodic intermittent bursts that cause sudden fluctuations of both soil and air temperature, in which local soil temperature maxima are accompanied by air temperature minima (Fig. 2a). It is important, however, to notice that the intermittent bursts in our model do not depend on using a vegetated surface layer with a low thermal capacity, as in the study of van de Wiel et al. (2002a). When more SBL levels are considered, the solutions become increasingly more complex. With three levels (Fig. 2b), the temperature evolution is no longer periodic, and two main types of fluctuations are observed to occur. Small soil temperature peaks happen along with local minima of the lowest air temperature level. However, at higher SBL levels,
hardly any variations can be perceived. On the other hand, there are larger soil temperature peaks accompanied by increased air temperature at the lowest level and local minima at the higher SBL levels. With five or more levels a similar pattern can be perceived, but different modes of oscillation appear and the event amplitude decreases (Figs. 2c,d). Larger temperature fluctuations happen at the ground than in the air and, in general, the air temperature is more variable at the lowest levels. Nonperiodic bursts and larger temporal variability at lower levels are characteristics of observed intermittent turbulence (Sun et al. 2002; Banta et al. 2007). From this point on, the analysis will focus on the model with five SBL levels (Fig. 2c), chosen because the general behavior is qualitatively similar with five or seven SBL levels and five levels are enough to reproduce the unpredictability of the intermittent event, both in terms of their amplitude and frequency.

The distinction between the connected and disconnected states presented by McNider et al. (1995) is reproduced by the model. For a 50-m SBL, a 3 m s$^{-1}$ wind at its top is sufficient to keep the surface connected to the higher levels. For weaker winds, the solutions are intermittent, with small average temperature (Figs. 3a,c) and friction velocity (Figs. 3b,d). Increasing the wind from 1.3 to 2.4 m s$^{-1}$ increases the intermittency, causing larger and more frequent fluctuations. As a consequence, such increased wind causes a slight increase in the average temperature. However, if a 3 m s$^{-1}$ wind is assumed, an abrupt transition occurs, with a sudden warming of the surface, that becomes 4 K warmer than in the weaker wind case (Figs. 3a,e). The friction velocity, accordingly, no longer varies, reaching an equilibrium value near 0.2 m s$^{-1}$—indicating that the solution has reached the connected state. Revelle (1993) presented intermittent behavior over a range of geostrophic winds between 1.5 and 3.0 m s$^{-1}$, which is a similar maximum limit to that obtained for a 50-m SBL. In Fig. 3, the first 10 h of simulation are shown so that the transient solution can be observed to be similar to those in McNider et al. (1995) and van de Wiel et al. (2002a) models.

The boundary layer height plays an important role in the intermittent flow behavior. For the same wind magnitudes at the top, the surface remains connected to the upper levels if a 25-m SBL is considered (Fig. 4a) and intermittent behavior occurs with deeper boundary
layers (Figs. 4b–d). Additionally, the less frequent occurrence of very low friction velocity values with shallower boundary layers is indicative that the wind threshold above which connection occurs increases as the SBL height increases. This is an obvious consequence of the stronger wind shear for shallower boundary layers with the same wind at its top, causing increased mechanical turbulence production, represented by the first rhs term in Eq. (11). The geostrophic wind that causes connection between the surface and the upper levels for different SBL depths is shown in Fig. 5.

Van de Wiel et al. (2002a) concluded that intermittency is strongly related to the soil physical characteristics, specially its heat capacity and conductivity, stating that “… intermittency is not easily found above a homogeneous bare soil.” Although the present results do not support this conclusion, indicating that intermittency can indeed occur over different soil types, including bare soils, the solution are dependent on soil type. Three different soil types were considered, and the values of their physical properties are presented in the appendix. In general, lower heat capacity (dry peat soil) allows larger ground temperature fluctuations (Fig. 6). On the other hand, Fig. 6f shows that while connection has already occurred for dry peat with 3.5 m s\(^{-1}\) winds, the same forcing still leads to intermittency with dry sand and clay. Therefore, higher heat capacities cause reduced soil temperature variability but allow intermittent turbulence at larger geostrophic winds.

The air temperature and friction velocity variability are also higher with smaller heat capacity. However, the distinction among the different soil types is reduced for these variables (Fig. 7). The model can therefore reproduce unpredictable turbulence bursts even with a very high heat capacity, such as that of clay (Fig. 7d). These results show that intermittency dependence on soil type is a matter of degree, not of kind. Although Fig. 6 shows that the amplitude of soil temperature variability decreases appreciably as the soil heat capacity increases, Fig. 7 shows that the amplitudes of both air temperature and the turbulence intensity oscillations have a much smaller dependency on soil type. The analysis in the remainder of the paper will always consider the soil to be dry peat, although it is important to stress that intermittencies take place even on bare saturated soil.

The SBL behavior as it approaches the connected state is illustrated in Fig. 8. In this case, the air temperature presents nearly periodic oscillations of large amplitude. Slight increases of the mechanical forcing at
SBL top increasingly cause reduction in the oscillation amplitude until it reaches a constant value. It is important to notice that the whole range of behaviors shown in Fig. 8 happens for a very narrow range of geostrophic winds. This result indicates that the model has very high sensitivity to the initial conditions when it is near the limit between connection and disconnection, a result that was found by McNider et al. (1995).

4. Discussion

In this section, we analyze the outputs of the simplified model to identify some features of the solutions and whether it may improve the understanding of the physical processes leading to intermittency and the consequences of its occurrence.

We first compare the vertical profiles obtained with different geostrophic winds. The $u_G$ values considered vary from 1 m s$^{-1}$, when almost no turbulence exists (Fig. 7b), to 3.5 m s$^{-1}$, a value for which the surface is connected to the upper SBL (Fig. 6f).

The friction velocity profiles (Fig. 9a) give a clear idea of the vertical structure of the turbulence in the SBL. For the weakest forcing ($u_G = 1$ m s$^{-1}$), the turbulence bursts happen only at the lowest level, while at higher levels the friction velocity remains unchanged at minimum $u_\ast$.

As the geostrophic wind increases, the bursts start to progressively occur at more levels, causing the average $u_\ast$ at those levels to deviate from its minimum. For the intermittent regime, $u_\ast$ decreases from its surface value to the minimum close to the SBL top. Similar characteristics of friction velocity profiles in very stable boundary layers have been observed by Banta et al. (2007). A clearly different $u_\ast$ profile occurs when the surface...
becomes connected to the upper levels ($u_G = 3.5 \text{ m s}^{-1}$). Then appreciably larger winds are spread over the entire SBL so that $u_*$ decreases very slightly with height.

In terms of mean wind profiles (Fig. 9b), for the weakest forcings (1.0 and 2.0 m s$^{-1}$) a constant wind layer forms at the SBL top. It occurs at the same heights where the friction velocity is minimum and shows that, in these cases, the very low turbulence intensity restrains the surface drag to the lowest SBL levels. A shallow 10-m shear layer at extremely stable conditions with approximately constant winds above it has been reported by Mahrt and Vickers (2006) and Burns et al. (2010). For the intermittent cases, as the geostrophic winds increase, the flow approaches a logarithmic wind profile. In the connected state, a sharper gradient occurs at SBL top, and the wind profile becomes nearly linear, as the larger turbulent intensity allows the lower boundary condition to be effectively felt at higher heights. Furthermore, this result is in agreement with the observations that Steeneveld et al. (2006) compared to their model. Our setup, with constant wind at the top, may be analyzed in analogy with a Couette flow. In that sense, we refer to the study of van de Wiel et al. (2007), who provided a detailed analysis of the dynamical stability of such flow.

The temperature profiles also depend on mechanical forcing (Fig. 9c). As in the case of the other variables, the most noticeable difference occurs between disconnected and connected state. Although a stable layer occurs in all cases, a much warmer surface temperature causes a reduced thermal stratification in the connected case ($u_G = 3.5 \text{ m s}^{-1}$). For the disconnected condition, the profiles are generally similar. However, it is interesting to notice that, although the coldest surface air temperature occurs with weakest geostrophic winds, at upper levels the SBL gets warmer for the same conditions. This fact shows that the temperature profile is more strongly curved for weaker mechanical forcing and, therefore, stronger temperature gradients occur near the surface under these conditions, as has been observed by Mahrt and Vickers (2006).

The sharper gradients for lower geostrophic winds, along with small wind magnitudes, cause a very large Richardson number in these cases. In fact, Fig. 9d shows that the model critical Richardson number is 1 and, as the geostrophic wind increases, an increasingly larger number of levels have an average Richardson number near its critical value. For the connected state ($u_G = 3.5 \text{ m s}^{-1}$), it is below critical at all levels. It is important to stress that such a critical Richardson number is a natural result of the model, never being arbitrarily chosen, and that it does not represent a value above which turbulence disappears but, rather, a frontier between the connected and disconnected states. The value
of Ri_c = 1 is supported by Abarbanel et al. (1984), Galperin et al. (2007), and Zilitinkevich et al. (2008), among others, who claim that the critical Richardson number might be a value between 0.25 and 1.

The intermittent events drive different temperature tendencies, depending on event intensity (Fig. 10). At the lowest atmospheric level, the weakest events promote cooling, as can be seen in Figs. 3a and 3b. In such cases, the events are restrained to a shallow layer near the surface so that they mix the air toward the colder surface but not all the way toward the warmer boundary layer top. More intense events (on average above a 0.12 m s\(^{-1}\) \(u_g^*\) threshold), on the other hand, promote surface warming as they are able to mix the entire boundary layer, bringing warmer air down. A similar distinction as to the role of weak and strong bursts in respectively cooling and warming the surface has been reported by Acevedo and Fitzjarrald (2003) and by Acevedo et al. (2006). At upper levels, the opposite occurs, as only the most intense events are able to uplift the cold air from the surface, while the weak events cause local mixing, promoting warming. The most intense cooling caused by the intense bursts occurs at middle levels (20 and 35 m), a consequence of the fact that not all events are able to reach the upper SBL. Except for extremely weak events, the ground temperature tends to warm during the bursts.

Both observations and theory show that intermittency is favored under a range of weak mechanical forcing and clear skies (van de Wiel et al. 2002b, 2003). The response of the present model to such external forcings can be determined by varying the geostrophic wind in Eqs. (7) and (8) and the cloud cover in the \(I_1\) term of Eq. (10). The average air temperature depends on both cloud cover and mechanical forcing (Fig. 11a). The largest
differences occur between the connected and disconnected states, but the role of intermittency can be seen by the fact that the disconnected state average temperature is not a function of cloud cover alone, as was shown by Acevedo and Fitzjarrald (2001) for the two-layer model of McNider et al. (1995). The geostrophic wind also affects the average temperature, in the sense that the air near the surface warms as the mechanical forcing increases, causing more intense and frequent bursts. On the other hand, weak intermittent events cool the surface, as they promote mixing of a shallow layer next to the surface, causing the coldest average temperatures not to occur under the weakest geostrophic winds. In the connected state, on the other hand, the average temperature becomes approximately independent of the external forcings. Furthermore, as the cloudiness increases, weaker mechanical forcing is necessary for connection to occur.

The friction velocity also depends on both geostrophic wind and cloudiness (Fig. 11b), but in this case the relationship is simpler, as its average value increases with both variables. It shows that both cloudiness and mechanical forcing cause more intense and frequent turbulent bursts before connection occurs, leading to a more turbulent boundary layer after it becomes connected.

As a measure of intermittency, the temporal standard deviations of both air temperature (Fig. 11c) and friction velocity (Fig. 11d) are shown. For a 100-m SBL, the most intermittent conditions occur for clear skies and a geostrophic wind near 3 m s$^{-1}$, as these are the values that...
maximize both temperature and friction velocity standard deviations. For the connected state, no intermittency happens, as the variables reach a constant value. Similarly, there is not much variability with very weak turbulence. The intermittency diagrams represented by Figs. 11c and 11d confirm the result found by van de Wiel et al. (2002b). In that study, a similar contour plot is presented, suggesting that intermittency occurs over ranges of pressure gradient and net radiation that correspond to those of geostrophic wind and cloudiness shown in Fig. 11. Van de Wiel et al. (2003) concluded that observed data are in good agreement with that classification. Therefore, the present model provides a characterization of intermittency and its dependence on external forcing that is in good agreement with previous studies on the subject.

To understand the physical mechanisms causing turbulence intermittency, a specific event is analyzed in detail (Fig. 12). The burst originates near the surface and propagates upward. Initially, it is driven by the shear increase at the lowest levels. While the SBL is in the disconnected state, the shear tends to increase steadily, forced by the first rhs terms in Eqs. (7) and (8), causing the Richardson number to decrease. When it falls below unity, the TKE production term surpasses the buoyant destruction term, initiating the event. Once turbulence starts at the lowest level, it is transported upward by the turbulence transport term of the TKE equation. When the SBL is at a near critical state, which happens with stronger geostrophic winds, this mechanism transports turbulence all the way to the SBL top. At the same time, downward turbulence transport intensifies, even more, the friction velocity near the surface, causing the peaks to occur. For smaller geostrophic winds, the bursts are initiated by a similar process, near the surface. However, in that case, the upper levels are more stable so that the turbulent transfer is not enough to overcome the local buoyant destruction, and the events become restrained to a shallow layer.

5. Conclusions

The present study has shown that global intermittency is a natural characteristic of the nocturnal boundary layer. Under simplified conditions, with fixed winds at the boundary layer top, also constant during the simulations,
turbulence has an intermittent character for a given range of external parameters, characterized by low winds and clear skies. The intermittent events reproduced by the proposed model consist of fluctuations of air temperature that can reach as much as 3 K, while the friction velocity varies as much as 0.2 m s$^{-1}$. On average, the longest events may last as much as 2 h, but there is a whole range of variability in the modeled events. Short 10-min events may produce temperature fluctuations of 2 K, while much longer 5-h events occur when the SBL is about to connect, as shown in Fig. 8. Observed data show a similar large range of variability in turbulence fluctuation, both in terms of duration and intensity. In addition, the observed temperature fluctuations associated with the events is also largely variable. Therefore, the model proposed in this study qualitatively reproduces a large portion of the observed behavior during global intermittency. Furthermore, the fact that such occurrence is favored under the external conditions of low-wind and clear skies is also in agreement with observations.

Previous studies have associated global intermittency with mesoscale phenomena, such as density currents, gravity waves, low-level jets, and Kelvin–Helmholtz instabilities (Sun et al. 2002, 2004; Banta et al. 2007; Cuxart and Jiménez 2007). In these cases, the instabilities originate outside of the SBL, propagating downward. Such intermittent events tend to be larger and last longer than those reproduced by the present model. Although they cause very large scalar fluctuations near the surface, being very important when they happen, their occurrence is more uncommon than the smaller-scale events generated within the SBL that were analyzed here. The intermittent events simulated by van de Wiel et al. (2002a) are of a more similar type to those reported here, in the sense that they are also generated within the SBL near the ground surface. However, in that case it is suggested that the intermittency originates through the interaction between the surface layer and a vegetated layer of low heat capacity. Turbulence, in that case, is switched off by the rapid surface cooling and subsequently switched on by the wind speed acceleration. In the present model, on the other hand, the process is only initiated by the interactions at the surface, proceeding further by the vertical turbulence transport.

One important difference between the two processes is that in the one presented here, as in observed series, the turbulence bursts are not periodic. Besides, our results allow this surface-generated intermittency to occur over different types of surface, including homogeneous bare soil or even oceans.

When compared to other studies that used the Blackadar approach to represent the SBL and its interaction with the surface, such as those by Revelle (1993), McNider et al. (1995), and van de Wiel et al. (2002a), the reason why the present model is able to reproduce nonperiodic intermittency lies in the fact that it does not represent turbulent
intensity as a simple function of atmospheric stability through a stability function but, rather, uses a TKE prognostic equation. However, more complex models that also use a TKE budget equation, such as those by Duynkerke (1988) or Baas et al. (2008) do not report on reproducing intermittency. One reason is that such models usually consider small values for $s_E$, therefore increasing the turbulence diffusion and “smoothing” the intermittent events vertically. Increasing $s_E$ is a necessary condition for those models reproducing intermittency, but probably not a sufficient one, and a more detailed analysis is necessary. The proposed model must be regarded as an intermediate step toward a more physically based model.

In terms of applicability, the present study opens the possibility of coupling similar formulations to largescale meteorological models and improving their performance under stable conditions. It is not a trivial task, but it is potentially very promising in terms of applications. To achieve it, the model results need to be validated by a detailed comparison to observed data, which would give it a more physical character. Other features also need to be changed, such as the upper boundary condition, which needs to allow variable winds to simulate processes such as low-level jets. The vertical domain also must be expanded to include more than simply the SBL thickness. Being a simplified model, there are many other features that need to be included. Among these, we can mention radiative flux divergence and prognostic equations for other surface scalars such as humidity (including latent heat in the energy budget) or carbon dioxide. In the TKE budget equation, the dissipation rate can be determined by a prognostic equation (Duynkerke 1988; Cuxart et al. 2006).

The purpose of the present work is the reproduction of intermittent turbulence bursts using a simplified model of the surface–atmosphere exchange in the nocturnal

<table>
<thead>
<tr>
<th>u</th>
<th>Zonal component of wind</th>
<th>Prognostic variable</th>
</tr>
</thead>
<tbody>
<tr>
<td>v</td>
<td>Meridional component of wind</td>
<td>Prognostic variable</td>
</tr>
<tr>
<td>$\theta$</td>
<td>Air temperature</td>
<td>Prognostic variable</td>
</tr>
<tr>
<td>$\theta_g$</td>
<td>Ground temperature</td>
<td>Prognostic variable</td>
</tr>
<tr>
<td>$E$</td>
<td>Turbulent kinetic energy</td>
<td>Prognostic variable</td>
</tr>
<tr>
<td>$u_G$</td>
<td>Zonal geostrophic wind</td>
<td>External parameter</td>
</tr>
<tr>
<td>$Q_c$</td>
<td>Cloud cover</td>
<td>External parameter</td>
</tr>
<tr>
<td>h</td>
<td>Boundary layer height</td>
<td>External parameter</td>
</tr>
<tr>
<td>$Z$</td>
<td>Height of intermediate levels</td>
<td>External parameter</td>
</tr>
<tr>
<td>$\omega$</td>
<td>Earth angular velocity</td>
<td>7.27 × 10^{-5} rad s^{-1}</td>
</tr>
<tr>
<td>$k_m$</td>
<td>Soil heat transfer coefficient</td>
<td>1.18</td>
</tr>
<tr>
<td>$\kappa$</td>
<td>von Kármán constant</td>
<td>0.4</td>
</tr>
<tr>
<td>$Pr$</td>
<td>Turbulent Prandtl number</td>
<td>1</td>
</tr>
<tr>
<td>$\Theta$</td>
<td>Reference temperature</td>
<td>300 K</td>
</tr>
<tr>
<td>$\theta_m$</td>
<td>Temperature of the substrate</td>
<td>285 K</td>
</tr>
<tr>
<td>$\rho$</td>
<td>Air density</td>
<td>1.225 kg m^{-3}</td>
</tr>
<tr>
<td>$c_p$</td>
<td>Air specific heat at constant pressure</td>
<td>1005 J kg^{-1} K^{-1} (dry peat soil)</td>
</tr>
<tr>
<td>$\lambda$</td>
<td>Soil thermal conductivity</td>
<td>0.06 W m^{-1} K^{-1} (dry sand)</td>
</tr>
<tr>
<td>$c_s$</td>
<td>Soil specific heat</td>
<td>1.18 W m^{-1} K^{-1} (clay)</td>
</tr>
<tr>
<td>$\rho_s$</td>
<td>Soil density</td>
<td>1.92 × 10^3 J kg^{-1} K^{-1} (dry peat soil)</td>
</tr>
<tr>
<td>$e_a$</td>
<td>Specific humidity (at $z_a$)</td>
<td>0.80 × 10^3 J kg^{-1} K^{-1} (dry sand)</td>
</tr>
<tr>
<td>$\beta$</td>
<td>Coriolis parameter</td>
<td>1.25 × 10^3 J kg^{-1} K^{-1} (clay)</td>
</tr>
<tr>
<td>$Q_a$</td>
<td>Specific humidity (at $z_a$)</td>
<td>0.30 × 10^3 J kg^{-1} K^{-1} (dry peat soil)</td>
</tr>
<tr>
<td>$\sigma$</td>
<td>Stefan–Boltzmann constant</td>
<td>1.60 × 10^3 J kg^{-1} K^{-1} (dry sand)</td>
</tr>
<tr>
<td>$\sigma_E$</td>
<td>Turbulent Prandtl number for $E$</td>
<td>1.80 × 10^3 J kg^{-1} K^{-1} (clay)</td>
</tr>
<tr>
<td>$v_{Gz}$</td>
<td>Meridional geostrophic wind</td>
<td>1 × 10^{-4} s^{-1} (for $\phi = 45^\circ$)</td>
</tr>
<tr>
<td>$E_0$</td>
<td>Minimum value of $E$</td>
<td>0.003 g kg^{-1}</td>
</tr>
<tr>
<td>$u_{i0}$</td>
<td>Initial value for $u$ at the lowest level</td>
<td>5.669 × 10^{-8} W m^{-2} K^{-4}</td>
</tr>
<tr>
<td>$\tau_{e0}$</td>
<td>Minimum value of $E$</td>
<td>2.5</td>
</tr>
<tr>
<td>$C_{15}$</td>
<td>Constant</td>
<td>0.005 m^2 s^{-2}</td>
</tr>
<tr>
<td>$C_{10}$</td>
<td>Constant</td>
<td>0.1 m s^{-1}</td>
</tr>
<tr>
<td>$C_{18}$</td>
<td>Constant</td>
<td>0.033</td>
</tr>
<tr>
<td>$z_0$</td>
<td>Surface roughness</td>
<td>0.18</td>
</tr>
<tr>
<td>$z_0$</td>
<td>Surface roughness</td>
<td>0.1 m</td>
</tr>
</tbody>
</table>
boundary layer. The results obtained suggest that the modeled variables have a complex relationship among themselves. Studies such as Xin et al. (2001), Gallego et al. (2001), and Campanharo et al. (2008) have shown the occurrence of chaotic behavior for observed turbulent variables in the atmospheric boundary layer (ABL). However, no chaotic set of equations has been associated with the mean state of the ABL. The present model may provide such association, and this will be investigated in future work.

Acknowledgments. This work was supported by Brazilian Research Agency CNPq (Conselho Nacional de Desenvolvimento Científico e Tecnológico). FDC is supported by Comissão de Aperfeiçoamento de Pessoal de Nível Superior (CAPES). The authors are grateful to Universidade Federal de Santa Maria for covering the publication costs. Two reviewers provided valuable suggestions that truly improved the manuscript.

APPENDIX

List of Symbols and Constants

Table A1 provides a list of all symbols and constants used in the paper.

REFERENCES


