The Effects of Critical Layers on Residual Layer Turbulence

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ABSTRACT

The authors report results of a study of finescale turbulence structure in the portion of the nocturnal boundary layer known as the residual layer (RL). The study covers two nights during the Cooperative Atmosphere–Surface Exchange Study 1999 (CASES-99) field experiment that exhibit significant differences in turbulence, as indicated by the observed turbulence dissipation rates in the RL. The RL turbulence sometimes reaches intensities comparable to those in the underlying stable boundary layer.

The commonly accepted concept of turbulence generation below critical values of the gradient Richardson number (Ri$_g$) is scale dependent: Ri values typically decrease with decreasing vertical scale size, so that critical Ri$_g$ values (~0.25) occur at vertical scales of only a few tens of meters. The very small scale for the occurrence of subcritical Ri poses problems for incorporating experimentally determined Ri$_g$-based methods in model closures in models with poor resolution.

There appear to be two distinct turbulence “regimes” in the RL: a very weak but ever-present background turbulence level with minimal temporal and spatial structure and a more intense intermittent regime during which turbulent intensity can approach near-surface nighttime turbulent intensities. It is hypothesized that the locally produced RL turbulence can be related to upward-propagating atmospheric gravity waves generated by flow over the low-relief terrain. The presence of critical layers in the RL, caused by wind turning with height, results in the generation of intermittent turbulence.

1. Introduction

Turbulence in stably stratified atmospheric flows remains an ongoing focus of boundary layer research. In the stably stratified boundary layer, turbulence develops during nighttime conditions when cooling by thermal radiation from the earth’s surface produces a steep positive temperature gradient near the surface. This gradient partially isolates the overlying atmosphere from the surface, and the depth of this stably stratified layer increases with time, producing a three-tiered nocturnal boundary layer (NBL): the stable boundary layer (SBL), the residual layer (RL), and a capping inversion that separates the entire NBL from the free atmosphere (Glickman 2000; Stull 1993; Jacobson 1999).

Stably stratified turbulence can be difficult to observe because the turbulence is weak and sometimes highly intermittent (e.g., Mahrt 2007; Nakamura and Mahrt 2006). Classical methods of aggregating data from high-rate instruments over longer periods become problematic when the turbulence is not stationary and when turbulence intensities temporarily are below the sensitivity threshold of the instruments. Modeling of
the stable boundary layer turbulence is also problematic. Experiments show that modeled turbulent fluxes during stable conditions are typically larger than observed (e.g., Tjernström et al. 2005). This is often connected to a too-deep PBL, at least in weather forecast and climate models (Cuxart et al. 2006; Svensson and Holtslag 2007). Relaxing the stability functions in operational models to obtain more realistic mixing, as compared with observations, usually degrades model forecast scores in terms of too-long-lived synoptic-scale systems. The too-deep SBL and the residual layer structure in models also have consequences for the low-level jet structure. Such model deficiencies can alter the transition between the nighttime and daytime conditions and thereby the diurnal cycle in near-surface wind and temperature (Svensson and Holtslag 2007).

The base of the RL, residing just above the SBL, typically lies at heights between several tens of meters and hundreds of meters. The RL has been less studied than the SBL primarily because of its relative inaccessibility. Most high-resolution research-grade observations are performed on instrumented masts that are only sometimes sufficiently high to reach into the RL (see, e.g., Finningan et al. 1984; Einaudi and Finningan 1993; De Baas and Driedonks 1985). Remote sensing (e.g., sodar, radar, or lidar; Coulter and Kallistratova 2004; Rotach et al. 2005; Strauch et al. 1976; Eaton et al. 1995; Fochesatto et al. 2001; Banta et al. 2002, 2003; Rotach et al. 2005) has been successfully employed in addition to tethered soundings (Holden et al. 2000; Siebert et al. 2002; Balsley et al. 2003, 2006, 2008; Muschinski et al. 2001; Rotach et al. 2005) and research aircraft (e.g., Poulos et al. 2002). The classical view of the RL is that of a quiescent remnant of the previous day’s well-mixed convective boundary layer, where the turbulence has since decayed. This view is still reflected in current models, where the RL is often seen to be more or less unchanging during the night and affected only by radiation and large-scale processes, such as horizontal advection and subsidence. There is a growing realization that this may be an oversimplification, and it has been shown that active turbulence exists in the RL (see, e.g., Mahrt and Vickers 2002 and Balsley et al. 2008). However, the origins of this turbulence are unclear. One possible source is turbulence generation by gravity waves. For example, it is well known that gravity waves can temporarily modify relatively large local $R_i$ values, sometimes reducing them to values at which production of turbulence can start (Fua et al. 1982; Nappo 2002, Nappo et al. 2004; Meillier 2008). Breaking gravity waves might also be a source of turbulence.

In this paper, we analyze observation taken by the University of Colorado’s Cooperative Institute for Research in the Environmental Sciences (CIRES) Tethered Lifting System (TLS), during the Cooperative Atmosphere–Surface Exchange Study 1999 (CASES-99) field experiment in Kansas (Poulos et al. 2002). We use data obtained by the TLS from both the SBL and the RL on two separate nights. We focus here primarily on the turbulence structure of the RL. We also explore the possible role of terrain-generated gravity waves in the observed turbulence structure of the RL. The paper is organized with a description of the data and methods in section 2, a presentation of main results in section 3, and a discussion of these results in section 4. The main conclusions are presented in section 5.

2. Method

a. The data

We use data obtained during CASES-99 by the TLS on the nights of 13–14 October (TLS Flight #7) and 18–19 October (TLS Flight #8), hereafter referred to as F7 and F8, respectively. Specifically, we use the sawtooth-like profile portions of these flights, each portion consisting of a series of slow and more or less continuous ascents and descents. This procedure provides eight profiles during F7 and six profiles during F8, in both cases over a few hours. More information on the background conditions during these two nights can be found in Fritts et al. (2003) and Balsley et al. (2008).

The TLS technique involves lofting either a kite or an aerodynamic balloon (kites for moderate-to-strong wind conditions; balloons for low-wind conditions) to carry a suite of lightweight, state-of-the-art turbulence packages from the ground through the first 1–2 km of the atmosphere. Details of the technology and a variety of results have been covered in earlier papers (e.g., Balsley et al. 1998, 2003, 2008; Muschinski et al. 2001; Frehlich et al. 2003, 2004) and will not be repeated here. It is sufficient to point out the unique advantages of the TLS with its ability to provide relatively inexpensive, point-by-point, high-resolution measurements of winds, temperatures, and turbulence throughout the lower parts of the atmosphere.

On both flights analyzed in this paper, three identical but independent instrument packages were suspended from the same tether, separated by 12 m on F7 and by 6 m on F8. Data from all three turbulence packages were treated the same way. Incorporating data from all three packages increases the amount of independent data samples by a factor of 3, thereby increasing the significance of the statistical analysis [means, probability distribution functions (PDFs), etc.]. Thus, the average value of a specific variable over a height interval consists...
of data from three independent instruments that, over a short time period, pass through that height interval.

Each package, in addition to carrying instruments for low-frequency temperature and wind speed and direction, carried hot- and cold-wire instruments for high-rate (200 Hz) observations of wind speed and temperature, respectively. Note that the wind speed is essentially the along-wind component of the horizontal wind because the sensors are directed into the wind by a wind-vane system. The vertical wind speed component was not measured.

Once-per-second power spectra of wind speed and temperature are generated from the archived data after carefully calibrating the hot- and cold-wire data using the low-frequency wind speed and temperature sensors (Frehlich et al. 2003). Estimates of the dissipation rate of turbulent kinetic energy (TKE) \( \varepsilon \) and the temperature structure function \( C_f^2 \) are derived from these spectra, using inertial subrange turbulence theory (e.g., Frehlich et al. 2003). This procedure provides a 1-Hz resolution estimate of small-scale turbulence. Taking into account the typical \( \pm 0.45 \) m s\(^{-1}\) vertical velocity of the turbulence packages during ascents and descents, the 1-Hz temporal resolution corresponds to a vertical resolution of 0.45 m.

The calibration and quality-assurance procedure for these estimations are covered extensively in Frehlich et al. (2003) and will not be repeated here. The accuracy of the individual estimates of \( \varepsilon \) is \( \sim \pm 15\% \), whereas the detection limit is \( \varepsilon \sim 10^{-7} \) m\(^2\) s\(^{-3}\) (Frehlich et al. 2003). Additionally, random errors are minimized using averages over large numbers of independent individual estimates; this is facilitated by using several instrument packages on the same tether. For F7, there are thus in total more than 24 000 individual estimates of \( \varepsilon \) in the RL and over 43 000 estimates in the SBL; the corresponding numbers for F8 are 39 000 and 6000, respectively.

Throughout the rest of this paper, we use the term "turbulence" to refer specifically to \( \varepsilon \) as a proxy for TKE itself. The use of high–temporal resolution \( \varepsilon \) is particularly useful as a proxy for TKE under stable conditions because it is relatively insensitive to the problems of stationarity and locality (common problems when observing turbulence in stably stratified flows; Vickers and Mahrt 2003; Basu et al. 2006). In turbulence theory, \( \varepsilon \) is often related to TKE through a time or a length scale (e.g., Stull 1993) and these scales are often related to the frequency or wavenumber of the peak of the power spectra of wind speed. Such length (or time) scales may be difficult to estimate in stably stratified turbulence because Fourier analysis requires continuous turbulence for a well-defined spectral peak. However, although \( \varepsilon \) is not uniquely related to TKE, it is reasonable to assume here that high values of \( \varepsilon \) imply larger TKE, and vice versa, during reasonably stationary background conditions and over limited time periods.

b. Analysis of profiles

Although it is not directly critical to the results in this paper, we have defined the top of the SBL, and thus the base of the RL, as the location of the wind speed maximum in the low-level jet. There is an ongoing discussion as to how best to define the top of the SBL: minimum shear, maximum stratification, or the level at which some measure of turbulence decreases below a fraction of its surface value (see Balsley et al. 2006 and references therein). Realizing that SBLs may exist also when there is no low-level jet, we have chosen this definition here as a matter of convenience.

To relate the turbulence estimates to the background profiles, we use the gradient Richardson number \( R_i \):

\[
R_i = \frac{g}{\Theta} \left( \frac{\partial U}{\partial z} \right)^2,
\]

where \( \Theta \) is the potential temperature, \( U \) is the along-wind wind speed component, and \( g \) is the acceleration of gravity. Following the procedures developed in Balsley et al. (2008), we estimate \( R_i \) for a range of vertical scales as follows: For each point observation by an instrument package in a profile, we first establish the height interval around it corresponding to a certain scale and then make a linear fit of the wind and temperature profiles over that interval. The gradients are calculated from these linear fits, and in turn \( R_i \) is calculated. Finally, the resulting profile of \( R_i \) is smoothed over the same height-scale interval. The procedure is repeated for several different vertical scales. Because of the continuous ascents and descents with the instrument package, it is thus possible (unlike in similar observations using fixed instruments on masts) to calculate \( R_i \) as a continuous function of the vertical scale down to \( < 1 \) m. This provides a means of determining the effect of replacing the true \( R_i \) with its corresponding bulk value \( (R_{ib}) \), as is always the case when estimating \( R_i \) from mast or tower observations. Moreover, because the TLS profiles extend over hundreds of meters vertically, such studies can extend from the surface through the SBL and well into the RL. Because the same

\[1 \text{ In } R_i \approx R_{ib} = g/\Theta \partial \Delta z/(\Delta U)^2, \text{ where the } \Delta \text{ corresponds to a finite difference of the respective variable.} \]
instrument package is used for all estimates, bias problems associated with gradient calculations from a set of individual sensors at different heights are avoided.

Using $Ri_g$ to diagnose turbulence is a classical approach, and the importance of $Ri_g$ as an indicator for turbulence is a subject that is covered in most textbooks on boundary layer turbulence (e.g., Stull 1993; Garratt 1994). Note that $Ri_g$ can be interpreted as the ratio of the buoyancy to the shear turbulence generation terms in the TKE equation.\(^2\) Whereas a negative $Ri_g$ is an indication of buoyancy-generated turbulence, positive values are associated with stably stratified turbulence, as is the case here. Recently, the existence of a critical value of $Ri_g = Ri_{cr}$ beyond which turbulence in the atmosphere cannot be sustained, has been questioned. Many studies indicate a presence of turbulence at values of $Ri_g$ that appear supercritical (Gossard et al. 1985; Rohr et al. 1988; Banta et al. 2003; Mauritsen and Svensson 2007). There are also suggestions of a hysteresis, where $Ri_g$ in a laminar flow must drop below $Ri_g \approx 0.25$ to become turbulent; but turbulent flow, once initiated, can exist up to $Ri_g \approx 1.0$ before again becoming laminar (e.g., Stull 1993). Other studies imply that some other processes, such as those related to gravity waves, can generate and maintain turbulence at supercritical values of $Ri_g$ (e.g., Meillier 2004, 2008, and references cited therein). Recall, however, that the classical value of $Ri_c = 0.25$ was originally derived from linear instability analysis (Miles 1961) or energy considerations (e.g., Chandrasekhar 1961), and thus says very little about nonlinear instabilities and therefore nothing about turbulence. Balsley et al. (2008) also show that a proper determination of $Ri_g$ is in fact strongly scale dependent.

c. Wave analysis

Bretherton (1969) showed that the stress generated by gravity waves launched by three-dimensional terrain over Wales might be as large as the frictional stress at the ground surface. Using a method developed by Nappo and Svensson (2008), we here analyze the possible effect of gravity waves on the turbulence in the RL over the more gentle terrain at the CASES-99 site. The method, based on linear theory, is an extension of that outlined in Shutts (1995). The main addition is a parameterization of the effects of nonlinear wave breaking below critical levels.

To evaluate the wave-stress vertical profile, the Taylor–Goldstein equation is solved. Over nonuniform terrain, the integration must be made for all the horizontal projection of the mean wind because waves can propagate in all directions (Bretherton 1969; Hines 1988). Thus, we integrate over the direction $\phi$ for $\theta_0 - \pi/2 \leq \phi \leq \theta_0 + \pi/2$, where $\theta_0$ is the direction of the surface wind vector (note that surface-generated waves in a direction opposite to the surface wind will immediately reach a critical layer). A critical level is reached where the wind in the direction of the wave propagation equals the phase speed of the wave (Booker and Bretherton 1967). For the case of terrain-generated gravity waves, which are stationary relative to the ground surface, a critical level exists where the wind speed in the direction of wave vector at the ground surface vanishes. Shutts (1995) demonstrated that when the wind turns with height, a critical level is present. The effect of wave breaking below the critical level is here parameterized using the so-called terrain-height adjustment wave-saturation technique (Nappo 2002; Nappo et al. 2004). Thus, in a stably stratified layer with directional shear, the area-averaged stress will diverge with height. A consequence of this is that the net wave drag may be in other directions than the drag exerted by the shear-generated turbulence. Based on the physics described above, we suggest that in a stably stratified RL with directional shear over nonuniform terrain, wave breaking will occur and can be observed as small-scale motions (i.e., turbulence).

3. Results

The two cases discussed in this paper are illustrated in Fig. 1, which shows mean profiles of potential temperature, wind speed, and wind direction. Rather than showing one single-mean profile of each variable for each night, we show the range between the 5th and 95th percentiles for the entire time periods, evaluated over 20-m intervals in height. Also shown are the median profiles (solid) and the single profiles used for the wave calculation discussed later in this paper (dashed).

The two nights (F7 and F8) shown in Fig. 1 exhibit both distinct similarities and striking dissimilarities. Although both cases feature significant nocturnal low-level jets, the jet strength and height and the wind shear in the RL differ. The first night (F7) has a strong jet with peak winds reaching about 14 m s\(^{-1}\) at the top of a relatively deep (~180 m) SBL (Fig. 1a). In contrast, the jet during F8 is weaker, with a peak wind of ~7 m s\(^{-1}\) at the top of a much shallower (~60 m) SBL (Fig. 1b). Both cases show significant stability, but F7 has an almost constant potential temperature gradient throughout the entire depth of the TLS soundings, whereas F8
has a stronger temperature gradient in the SBL than the corresponding RL. F7 exhibits a larger variability in the wind direction, as illustrated by the wider band between the low and high percentiles. Note, however, that a significant wind direction shear is apparent only in the SBL of F7, whereas F8 exhibits a significant wind-direction shear aloft (i.e., above ~270 m).

The different conditions during F7 and F8 are clearly reflected in the turbulence structure, as illustrated in Fig. 2 by the probability distribution functions of $\varepsilon$ as a function of height. In F7, the largest values of $\varepsilon$ in the upper half of the SBL occur just below the jet maximum; the location of the wind speed maximum, and hence the SBL height, is indicated by the horizontal gray band. Significantly lower values of $\varepsilon$ occur at the lower heights, consistent with the notion of an “upside-down” SBL (Mahrt 1999) where SBL turbulence production is dominated by the shear below the jet rather than by the near-surface shear [see Balsley et al. (2006) for a discussion of this case]. In contrast, F8 exhibits the largest values of $\varepsilon$ close to the surface, with decreasing values with increasing height up to the height of the jet maximum around 50–70 m, also indicated with a gray band. Most of the SBL turbulence is thus generated at the surface, by surface friction and the associated wind shear.

The most intriguing differences between these two nights, however, lie above the jet in the RL. In F7, $\varepsilon$ drops dramatically to low values across the jet and continues to decrease slowly with increasing height in the RL. In contrast, although $\varepsilon$ in F8 also decreases markedly with height across the jet, mean values of $\varepsilon$ remain low up to ~150 m and then increases significantly with height, approaching values comparable to those found in the upper SBL. Comparison of the widths of the $\varepsilon$ PDFs in the RL (Figs. 2a,b) shows a much narrower structure for F7 than for F8. The reason for the broader distributions of $\varepsilon$ in the F8 RL becomes more obvious in Fig. 3, which illustrates the temporal variations of the small-scale turbulence structure as a function of height during the two nights. Although an $\varepsilon$ maximum near 100 m in F7 (somewhat below the jet maximum) is obvious, consistently lower and reasonably constant $\varepsilon$ values are apparent throughout the RL on this night (Fig. 3a). In contrast, F8 shows intermittent high $\varepsilon$ values at all heights above 200 m (Fig. 3b). The mean increase of $\varepsilon$ with height above 200 m in the PDF for F8 (Fig. 2b) is seen to be clearly composed of highly time-variable $\varepsilon$ values at essentially all heights above 150 m. It is also worth mentioning that throughout the RL even the lowest values of $\varepsilon$ in both F7 and F8 are well above the threshold of the instrument system.

That the $\varepsilon$ values in the SBL in Figs. 2–3 are associated with local velocity shears is reinforced by results shown in Fig. 4. The relationship between $\varepsilon$ and the corresponding Ri$_g$ for the F7 SBL is presented in Fig. 4a for Ri$_g$ values at six selected scale sizes. The intervals shown in this figure were determined using a double-sided Student’s $t$ test at the 95% confidence level; the medians, in the center of each interval, are omitted for clarity. Because it is not possible to draw a line of constant $\varepsilon$ staying within the intervals, the null hypothesis (i.e., that $\varepsilon$ is not a function of Ri$_g$) can be discarded. It is thus clear that, over a wide range of vertical scales, larger values of $\varepsilon$ are associated with Ri$_g$ < 0.1 and that $\varepsilon$ decreases by roughly two orders of magnitudes with increasing Ri$_g$ for 0.1 < Ri$_g$ < 1 and then remains low.
FIG. 2. Plots of TKE dissipation rate \( \varepsilon \) (m\(^2\) s\(^{-3}\)) from (a) F7 and (b) F8, showing PDFs (in shading) as a function of height for the entire flights. The gray band across each panel indicates the height interval at which the maximum wind speed in the low-level jet is found during the time of observation.
for even larger values of $Ri_g$. A similar analysis for the RL of F7, shown in Fig. 4b, reveals no such relationship at any of the scales, although the $Ri_g$ values extend over about the same range. In this example, RL $\varepsilon$ values appear to be consistently small and independent of $Ri_g$. A comparable study for the F8 SBL was, unfortunately, not possible because the SBL was too shallow to statistically examine such relationships. Results from the F8 RL, on the other hand, are shown in Fig. 4c. These results are surprisingly similar to those shown for the SBL in F8. They are, moreover, strikingly different from the F7 RL. It is significant that the subcritical (with
respect to $Ri_g$ values in F8 (Fig. 4c) are at least an order of magnitude above comparable $\varepsilon$ values for F7 (Fig. 4b). The striking differences in these two figures suggests a causal relation between the relatively small-scale RL structures apparent in Fig. 3b and the enhanced turbulence at small values of $Ri_g$ shown in Fig. 4c. The results in Figs. 4a,c also conform to the picture of stably stratified turbulence suggested by Mahrt (1998), with weakly stable turbulence for $Ri_g < \sim 0.1$, strongly stable turbulence for $Ri_g > \sim 1$, and a transition in between. This relationship was shown to hold for a large number of different independent datasets with stable boundary layer conditions in Mauritsen and Svensson (2007).

The scale dependence of $Ri_g$ in both the SBL and RL is shown separately for both F7 and F8 in Fig. 5 (see also Balsley et al. 2008), using the peak value of the PDF of $Ri_g$, evaluated separately over the entire SBL, and the portion of the RL available from the TLS data. Two things are clear from this plot: (i) $Ri_g$ increases with increasing vertical scale size in both the SBL and the RL of both F7 and F8, and (ii) in F8, $Ri_g$ values in the RL are smaller than those in the SBL for all scale sizes, whereas the opposite is true in F7. It follows that the RL is dynamically much less stable in F8 ($0.05 \leq Ri_g \leq 0.5$) than in F7 ($0.2 \leq Ri_g \leq 1.1$). Conversely, the SBL is dynamically somewhat more stable in F8 ($0.05 \leq Ri_g \leq 0.6$) than in F7 ($0.0 \leq Ri_g \leq 0.35$). It is worth noting in this context that models typically calculate $Ri_g$ over significantly larger scales than those used here. Thus, it appears possible that turbulence closures in models become systematically biased because the vertical resolution in the models is much poorer than that in the data (e.g., from towers) on which closures are based.

4. Discussion

a. Wave–turbulence interaction

We have shown in the above examples two different nights during CASES-99 having a reasonably similar mean RL structure but exhibiting strikingly different detailed turbulent structure. The obvious issue that needs to be resolved is the cause for these differences. In this section we examine the possibility that they are caused by terrain-generated gravity waves.

Nappo and Chimonas (1992) demonstrated that low-relief topography could launch gravity waves in the SBL. The vertical transport of momentum by these waves is constant unless wave dissipation occurs. At a critical level, where the mean flow in the direction of wave propagation vanishes, the wave is effectively absorbed.

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**Fig. 4.** Plots showing the variation of TKE dissipation rate $\varepsilon$ (m$^2$ s$^{-3}$) from the (a) F7 SBL, (b) F7 RL, and (c) F8 RL, as a function of $Ri_g$, evaluated over different scales. The thin solid vertical line outlines $Ri_g = 0.25$. In (a) and (c), shaded areas indicate the 95% significance interval around the median, which is in the center of each interval and is omitted for clarity. In (b), only the median is shown and different lengths of the lines reflect the span of $Ri_g$ found over the different scales.
and its stress acts as a drag against the mean flow (Booker and Bretherton 1967). As a wave approaches such a critical level, its amplitude increases and wave breaking (i.e., wave dissipation) occurs before the critical level is reached. Under the assumption of linear wave-saturation theory (Fritts 1984), wave amplitudes are virtually reduced such that the wave field remains convectively stable. Thus, between the height of the onset of wave breaking and the critical level, a continuous wave stress divergence exists and it is assumed that the stress divergence takes the form of a turbulence drag.

Note that the wave saturation parameterization applied here only considers convective stability. If a criterion for dynamic instability were included, the wave adjustment could occur at even lower heights.

Figure 6 shows area-averaged wave-drag profiles for gravity waves generated by the low-relief terrain at the CASES-99 site for the two nights under study. The profiles in Fig. 6 were calculated from select TLS mean profiles (see the dashed lines in Fig. 1) using the method described in section 2c. In Fig. 6, it is seen that during F7 the area-averaged wave-stress profile shows no significant divergence except close to the surface in the SBL, indicating that waves pass through the RL without depositing momentum. In contrast, there is a significant wave-drag divergence in the F8 RL at heights above 200–250 m. This divergence is associated with critical layers present because of the wind directional shear shown in Fig. 1b. The results in Fig. 6 is consistent with the results in Figs. 2 and 3 that show an increase in turbulence intensity above 150 m (Fig. 2b), with the intermittently largest values occurring at heights above 250 m (Fig. 3b). The change in wind direction starts above 250 m and becomes stronger above 350 m. Note that according to the linear wave-saturation theory discussed above, deposition of wave momentum begins as the upward-propagating waves approach critical levels (i.e., below the height where the wind directions start turning). In summary, it thus appears that the primary difference between these two nights lies in the wave-drag deposition. This arises from the significant wind-direction shear in the F8 RL, which is not observed in the F7 RL of F7.

Based on the above results, we hypothesize that breaking gravity waves result in enhanced TKE levels affecting the local mean gradients of temperature and wind, thereby bringing Ri to smaller values. However,
even before the waves become convectively unstable, they could modify the mean profiles in such a way that the Ri$_g$ is reduced below the critical value, which in turn could lead to turbulence production (i.e., dynamic instability). As waves propagate through the RL, these processes might occur at different heights and times, explaining the seemingly stochastic behavior in the turbulence above ~250 m in F8 seen in Fig. 3b.

This hypothesis is supported by results presented in Fig. 7, which examines the relationship between the square of wind shear ($S^2$) and the Brunt–Väisälä frequency ($N^2$) at different vertical scales. These curves show that for vertical scales of 10 m and larger, the mean value of $N^2$ lines up along the line 0.25 times $S^2$ (the dotted line in Fig. 6). This result suggests that for large values of shear, mean Ri$_g$ values tend toward Ri$_g$ = Ri$_e$ = 0.25. This is a reasonable result because the wave-drag momentum deposition enables generation of TKE by extracting momentum from the wave field only at a rate necessary to keep the waves convectively stable, in accordance with wave saturation theory. This relationship does not appear to hold for scales smaller than about 5 m (not shown); for these scales, $N^2$ does not seem to be a function of $S^2$, a fact that could possibly be related to the vertical scale of the waves themselves.

b. The existence of two turbulent regimes

We have demonstrated above that, given favorable conditions, intermittently intensified turbulence in the RL can be generated by gravity wave–turbulence interaction, where the upward-propagating waves are generated by stably stratified flow over low-relief terrain. In the region below these enhanced values of turbulence in F8 but still above the wind jet (at ~70–150 m), median turbulence values are found to be consistently well below the enhanced levels (i.e., $\varepsilon \sim 3 \times 10^{-6} \text{ m}^2 \text{s}^{-3}$, with maximum values reaching $\sim 10^{-4} \text{ m}^2 \text{s}^{-3}$). Throughout the RL of F7 (which shows little if any suggestion of local turbulence generation), typical median $\varepsilon$ values extend between $\varepsilon \sim 3 \times 10^{-6} \text{ m}^2 \text{s}^{-3}$ and $10^{-5} \text{ m}^2 \text{s}^{-3}$. Although relatively small, these values are more than an order of magnitude larger than the instrumental threshold at $\varepsilon \sim 10^{-7} \text{ m}^2 \text{s}^{-3}$ and are thus significant. Indeed, turbulence levels rarely if ever fall below the TLS threshold (see, e.g., Figs. 2a,b).

Based on the above comments, we suggest that, at least in the profiles presented here, turbulence is always present at a low but nonzero level, even for quite large values of Ri$_g$. This suggestion corroborates conclusions reached earlier (e.g., Balsley et al. 2008). Given the proper conditions, such as those in the upper RL of F8, this weak turbulence can grow to much larger values. We therefore postulate the existence of essentially two independent regimes of turbulence: (i) a weak background turbulence regime that appears to be to some degree always present and (ii) an active, dynamic turbulence regime present only under certain conditions. Thus, whereas SBL turbulence is kept dynamically active by wind shear overcoming buoyancy and dissipation in both of the cases shown here, it appears likely that the shear is too weak in the RL to sustain turbulence generation even when Ri$_g$ < 0.1, perhaps because these small Ri$_g$ values only occur over shallow layers of ~10 m or less (Fig. 4). However, when vertically propagating waves deposit momentum while approaching a critical level, turbulence is maintained without the presence of any additional instability mechanism.

Further evidence for this idea is provided by the results shown in Fig. 8. This figure presents a series of $\varepsilon$ PDFs obtained from thin vertical “slabs,” roughly between ~100 and 230 m during F7. Recall from Fig. 1a that the height of the jet in F7 was approximately 180 m. The PDFs in Fig. 8 were obtained over narrow intervals of normalized altitude, using the height to the jet peak to normalize the height axis. Each single profile was analyzed separately, thus taking the slow variation in jet height into account. The results for all the profiles during F7 were then averaged for each normalized height interval (or slab) shown in Fig. 8.

Examination of this figure shows that the PDF from the lowest layer (corresponding to the upper SBL, roughly 100–145 m) is unimodal, with a single dominant peak at $\varepsilon \sim 3 \times 10^{-3} \text{ m}^2 \text{s}^{-3}$. The shape of this distribution appears to be approximately lognormal,

**Fig. 7.** Plots of the median $N^2$ ($\text{s}^{-2}$) as a function of $S^2$ ($\text{s}^{-2}$) calculated from intervals in the latter at different vertical scales from the RL in F8. The dashed line is $N^2 = 0.25 S^2$. 
indicating a height range with active turbulence generated locally by the wind shear below the jet. In the uppermost layer (corresponding to the lower RL, roughly 200–230 m) the PDF also appears to be lognormal and unimodal, although the peak level is reduced to $\varepsilon \sim 10^{-5} \text{ m}^2 \text{ s}^{-3}$. This upper region of weaker turbulence suggests the presence of the weak background turbulence discussed above. Approaching the jet from above or below, the PDFs appear to remain approximately lognormal and unimodal, with slightly shifted peaks in agreement with the PDF profile in Fig. 2a. The two PDFs from the $\sim 10$-m-thick layers either just above or below the jet ($z = 0.9–0.95$ and $z = 1.0–1.05$) deviate only slightly from lognormal, showing “tails” toward higher and lower values, above and below the jet, respectively. The remaining PDF obtained from an approximately 10-m-thick layer centered just below the wind speed maximum ($0.95 < z_i^{-1} < 1$), is clearly different from the other PDFs. It is reasonable to expect that this PDF consists of samples from both above and below the separation between the two turbulence regimes because the SBL–RL interface changes slightly in height during the time that the sensor passes through the region, even during an individual single ascent or descent. Recall that with the 0.45 m s$^{-1}$ vertical speed of the instrument package, it takes about 4.5 s to traverse the 10-m interval. Taking into account the jet wind speed, this corresponds to a horizontal distance of $\sim 60.0$ m. It is quite reasonable that the height of the jet peak varied in height by 10 m over such horizontal distances. These results illustrate the very sharp transition from the actively turbulent SBL to the weakly turbulent RL in this case.

Weak but ubiquitous turbulence levels at some small but nonzero level throughout the residual layer were thus observed in both cases studied in this paper, as well as in our earlier studies (Balsley et al. 2008). It is our experience that the presence of weak background turbulence is the rule rather than the exception. This statement is based on all the TLS flights during CASES-99, as well as results from essentially all other campaigns during which TLS observations have been made. We suggest that there is no reason to assume that this would not also be the case in the free troposphere, although we have shown no results above the residual layer here. It follows directly from this statement that the concept of a laminar free troposphere above a turbulent boundary layer needs to be reconsidered at a fundamental level.

5. Conclusions

We have analyzed two nights of TLS observations from the CASES-99 field experiment using both the mean profiles and the 1-Hz information of TKE dissipation rate $\varepsilon$ deduced from the hot-wire measurements from several sawtooth-like profiles with the instrument system through the lowest several hundred meters of the atmosphere. Our main conclusions are as follows:

- Turbulence is always present in the residual layer, at least at some minimal level, even when the $R_i$ is large.
- When gravity waves propagate through a stably stratified residual layer where one or more critical levels are present, the waves might contribute to local turbulence.
- We hypothesize that waves, such as those launched by the modest CASES-99 terrain, are able to enhance the RL turbulence provided background conditions are right. The observed level of RL turbulence in such conditions can be of the same magnitude as that found in the SBL.
- The estimation of $R_i$ is strongly scale-dependent and is clearly different for different vertical scales and for different conditions.

These conclusions touch on some central assumptions for the modeling of effects of atmospheric turbulence on the larger-scale flow. Moreover, formulations of turbulence closures in numerical models of the atmosphere ultimately rest on empirical evidence. The fact that $R_i$ generally decreases with decreasing vertical scale imposes a problem when applying high–vertical resolution measurements to model closures in models with significantly coarser resolution. This fact may provide one explanation for why models with $R_i$-based closures
usually need a so-called “long-tail” formulation, whereby turbulence is allowed at higher $R_i$ than is typically considered supercritical (e.g., Steeneveld et al. 2007).

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