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Temporal Scales of the Nocturnal Flow Within and Above a Forest Canopy in Amazonia

Daniel M. Santos^{1,5} · Otávio C. Acevedo² · Marcelo Chamecki³ · José D. Fuentes³ · Tobias Gerken³ · Paul C. Stoy⁴

Abstract Multiresolution decomposition is applied to 10 months of nocturnal turbulence observations taken at eight levels within and above a forest canopy in Central Amazonia. The aim is to identify the contributions of different temporal scales of the flow above and within the canopy. Results show that turbulence intensity in the lower canopy is mostly affected by the static stability in the upper canopy. Horizontal velocity fluctuations peak at time scales longer than 100 s within the canopy, which correspond to the scale of non-turbulent submeso motions above the canopy. In the vertical velocity spectrum near the surface, the peak occurs at time scales around 100 s, which are larger than the time scales of the turbulent flow above the canopy. Heat-flux cospectra within the canopy peak at the same temporal scales as the vertical velocity fluctuations at that level, suggesting the existence of buoyancy driven turbulence. Case studies are presented as evidence that low-frequency fluctuations propagate towards the canopy interior more easily than does turbulence.

Keywords Amazonia · Canopy–atmosphere interaction · Intermittency · Stable boundary layer · Temporal scales

1 Introduction

The Amazon rainforest has received significant attention in studies of canopy–atmosphere interactions because of its relevance to the global climate (Nobre et al. 1991; Werth and

Avissar 2002). Early studies (Shuttleworth et al. 1985; Fitzjarrald et al. 1990; Fitzjarrald and Moore 1990; Viswanadham et al. 1990) investigated the flow characteristics within the canopy and its interaction with the atmosphere above the forest. More recently, Kruijt et al. (2000) used multi-level observations to investigate how atmospheric stability modulated the micrometeorological conditions within and above two Amazonian forested sites. Since then, the focus of micrometeorological studies in the region has been directed to the estimation of scalar fluxes, mainly carbon dioxide (CO_2), between the forest and the atmosphere (Malhi et al. 1998; Araújo et al. 2002; Saleska et al. 2003; Miller et al. 2004; Ometto et al. 2005, among others). Other studies have focused on specific aspects of the flow within and above the canopy, such as the role of coherent structures (Bolzan et al. 2002; Ramos et al. 2004), the occurrence of gravity waves (Zeri and Sá 2011), and subcanopy drainage flows (Tota et al. 2008, 2012).

Understanding the flow characteristics within and above forest canopies is important in quantifying the exchange of momentum, water and scalars between the forest and the atmosphere. The flow of air within the canopy is highly complex because obstacles such as leaves and trunks reduce turbulent intensity and lead to different stability regimes across the vertical extension of the forest. This complexity is enhanced during the night as the reduced levels of turbulence limit mixing within the canopy (Baldocchi and Meyers 1988) and flow structures become more localized in space and time (Thomas 2011).

Decomposition of atmospheric variables within the canopy into temporal scales is usually employed to investigate how the forest canopy interacts with the diverse flow scales that may coexist. Spectral analysis has been a useful tool in this context. Baldocchi and Meyers 1988 found that the most energetic wavenumber of all wind components within a foliated deciduous forest can be an order of magnitude larger than that above it. Observations in Amazonia also showed that the most energetic temporal and spatial scales of the flow are longer within the canopy than above it (Fitzjarrald and Moore 1990; Kruijt et al. 2000). A similar result was obtained for an open pine forest (Vickers and Thomas 2013) and a deciduous orchard forest with and without leaves (Dupont and Patton 2012). In the latter study, however, the difference between the most energetic frequencies above and within the canopy only reached one order of magnitude in the stable case and during the foliated period. Villani et al. (2003) found a shift of the spectral peak towards lower frequencies within the canopy, but they also remarked that the significance of this result is questionable because of uncertainties regarding the determination of the time scale of the spectral peak. Oliveira et al. (2013) found that the temporal scales of the maximum of the vertical velocity spectra, as well as those of heat-flux and CO_2 -flux cospectra, were larger within a pine forest than above it, and that the contrasts were largely enhanced for increasing levels of static stability. On the other hand, studies such as those of Blanken et al. (1998), Launiainen et al. (2007) and Vickers and Thomas (2014) did not find significant differences between dominant temporal scales above and within the canopy.

It is not completely clear what causes the increase in the characteristic time scales of transport within the forest and why this phenomenon is present in some canopies but not in others. Baldocchi and Meyers (1988) and Fitzjarrald and Moore (1990) suggested that the observation of larger time scales of flow within the canopy confirms the idea proposed by Shaw and Seginer (1985) according to which leaves remove part of the turbulent fluctuations, such that longer scales, which are less affected, become more dominant. Consequently, the spectra of within-canopy turbulent wind components decrease more rapidly with wavenumber than that proposed by Kolmogorov's law for fully-developed turbulence. However, although this mechanism explains the energy reduction within the canopy at short time scales, it does not completely clarify why the spectral peak within the canopy shifts to longer temporal scales

than those observed above it. Along that line of reasoning, [Villani et al. \(2003\)](#) speculated that such scale dependence might "... represent the survival of primarily larger scale eddies with increasing penetration into the canopy". In general, therefore, there is a body of evidence supporting the idea that fluctuations associated with longer time scales are not significantly dampened by the canopy.

In recent years it has been recognized that nocturnal boundary-layer flow may include an appreciable contribution of non-turbulent modes, which coexist with turbulence ([Mahrt 2009](#)). These modes are termed "submeso" because they have larger spatial and temporal scales than turbulence, but smaller than what is usually referred to as "mesoscale". The suppression of turbulence under stable conditions favours an increased role of submeso flow in these situations, such that non-turbulent contributions may have a dominant contribution to nocturnal variables at time scales of tens of seconds ([Acevedo et al. 2014](#)). These variables include the variances of the horizontal wind components and the vertical fluxes of scalars. It is not yet known the extent to which these longer non-turbulent fluctuations that occur above the canopy in stable conditions affect the flow within the forest. [Thomas \(2011\)](#) provided some evidence in that direction and found that submeso flow has a significant contribution to the observed variability of temperature and wind speed near the surface within a closed Douglas-fir canopy. [Andreae et al. \(2015\)](#) showed that events related to large CO₂ flux within an Amazonian canopy are triggered when the submeso variability favours increased wind shear above the canopy.

The purpose of the present study is to identify how the temporal scales of nocturnal flow within a forest canopy in Amazonia depend on flow characteristics above the forest, and how this relationship is affected by atmospheric stability. Particular attention is given to the contrasting behaviour between turbulent and submeso wind fluctuations and how they are differently affected by the canopy. This is done using a 10-month dataset of turbulence observations conducted at 10 heights within and above a mature tropical rainforest in the central Amazon. The mechanisms that induce the submeso flow are not presently investigated, although this is an important line of inquiry. Previous studies have suggested that submeso flow may be associated with terrain characteristics ([Monti et al. 2002](#)) or vegetation contrasts ([Vickers and Mahrt 2007](#)). In the humid Amazon environment, it is also possible that cloudiness plays a role in the origin of submeso flows.

2 Data and Methods

2.1 Study Site Characteristics and Data

The dataset was collected during the GoAmazon 2014–15 project, from March 2014 to January 2015. Observations were carried out in northern Brazil, at the Cuieiras Biological Reserve (2°36'32"S, 60°12'33"W), located 60 km north of the city of Manaus. The site is managed by Instituto Nacional de Pesquisas da Amazonia (INPA), and has been used for micrometeorological and ecological research for over two decades ([Malhi et al. 1998](#); [Kruijt et al. 2000](#); [Araújo et al. 2002](#); [Tota et al. 2012](#), among others). Vegetation at the site consists of primary tropical rain forest, with an average height of the highest trees (h) of 35 m. Estimates of the leaf area index range from 5 to 7.3 m² m⁻² ([McWilliam et al. 1993](#); [Marques Filho et al. 2005](#); [Tota et al. 2012](#)) and the leaf area density profile has local maxima near the surface and at 0.75 h ([Kruijt et al. 2000](#)). The tower is situated on a plateau that is about 60 m higher than the minimum elevation of the valleys, located 2 km away. [Araújo](#)

[et al. \(2002\)](#) detailed the main characteristics of the site regarding weather, climate, terrain, soil and vegetation.

The data were collected on a 50-m micrometeorological tower (the K34 tower, labeled TK0 for the GoAmazon 2014-15 project). Nine sonic anemometers (model CSAT3, Campbell Scientific, Inc., Logan, Utah) were installed and, of these, six were situated within the canopy at heights of 7, 13.5, 18.4, 22.1, 24.5 and 31.6 m above the surface. One was located at 34.9 m, which is used as a reference for conditions at canopy top. Above the canopy, there were two levels of observations at 40.4 and 48.2 m. An additional sonic anemometer was installed at 1.5 m above the forest floor on a second tower, deployed approximately 15 m away from the main tower. All sonic anemometers acquired data at 20 Hz. Net radiation measurements, used in Sect. 4.3, have been taken at 39.0 m with a temporal resolution of 1 min.

2.2 Data Analysis

In the present analysis, flow structures are simultaneously compared at the different measurement heights, therefore only data that had continuous observations, without spikes at all levels, are considered. Most of the analysis is performed using sonic anemometer data from eight levels: 7, 13.5, 18.4, 22.1, 24.5, 34.9, 40.4 and 48.2 m. The 31.6-m level was removed because it frequently presented gaps, so that its inclusion would reduce the entire dataset by more than half of its size. Another constraint imposed on the data was that only nights with continuous data during the entire period from 2100 to 0600 local standard time (LST) were used. The reason for such a requirement is that, given the intermittent nature of turbulence above the canopy at the site, the use of smaller portions of any given night could favour one turbulent regime over another and could thus introduce biases on the statistical analysis. Although sunset at the site occurs within a few minutes of 1800 LST throughout the year, the evening period between sunset and 2100 LST is not considered in the analysis to avoid sampling intense events that are common during that period and that may be associated with its transitional characteristics. Eighty-eight nights satisfied these constraints and were used in the study.

Two different time windows are used for the analysis of the turbulent series: 5 and 54.6 min. [Campos et al. \(2009\)](#) showed that at this site the contribution of turbulence to the nocturnal fluxes of sensible heat, latent heat and carbon dioxide above the canopy is restricted to time scales smaller than 200 s. For this reason statistical moments of turbulent quantities are determined for 5-min periods, in agreement with the procedure of [Dupont and Patton \(2012\)](#). Therefore, in most cases the 5-min statistics are dominated by contributions from turbulent processes. This is important because the use of longer time windows would lead to statistical moments affected by unknown relative contributions of turbulent and non-turbulent processes. To determine those contributions, the temporal decomposition of flow variables such as wind speeds and temperature are determined using multiresolution decomposition ([Mallat 1989](#); [Howell and Mahrt 1997](#)), which is applied to time series of 2^{16} points (corresponding to 54.6-min periods). The advantage of such a method over other more commonly used spectral decompositions, such as Fourier transformation, is that it decomposes a time series into its characteristic temporal scales without the constraint of periodicity. Therefore, a multiresolution value for a given time scale implies the occurrence of physical processes with that temporal scale, while the result of a Fourier transform must be interpreted in terms of periodic repetition on that time scale. Given that nocturnal boundary-layer processes (including those associated with canopy flows) are seldom periodic, the multiresolution decomposition better suits our purpose. In particular, low-frequency events are better represented by this method, as in this case a single occurrence is sufficient for their identification. In the present

study, the method is used both for the analysis of specific cases (see Sect. 5) and for understanding average behaviour (see Sect. 4). The decomposition was applied sequentially to the nocturnal time series starting at 2100 LST with no data overlap between subsequent blocks of data. This approach resulted in a total of nine decompositions from each night ending at 0511 LST, and totaling 792 decomposed series for each vertical level.

Quantities determined for 5-min periods include average horizontal wind speed (V), standard deviations of the wind components (σ_u , σ_v and σ_w), vertical kinematic momentum ($\overline{u'w'}$ and $\overline{v'w'}$) and heat ($F_H = \overline{w'\theta'}$) fluxes and the third- and fourth-order moments of the turbulent fluctuations of the wind components. From these, derived quantities are determined, such as the standard deviation of the horizontal wind fluctuations given by $\sigma_V = (\sigma_u^2 + \sigma_v^2)^{1/2}$, the turbulent velocity scale, defined as $V_{TKE} = [0.5 (\sigma_u^2 + \sigma_v^2 + \sigma_w^2)]^{1/2}$, the friction velocity $u_* = (\overline{u'w'^2} + \overline{v'w'^2})^{1/4}$, the velocity aspect ratio (Vickers and Thomas 2013), given by $VAR = 2^{1/2} \sigma_w (\sigma_u^2 + \sigma_v^2)^{-1/2}$, the skewness ($Sk_i = \overline{u_i'^3} \sigma_i^{-3}$) and kurtosis ($K_i = \overline{u_i'^4} \sigma_i^{-4}$) of the wind-velocity components, where $i = u, v, w$.

Multiresolution decompositions were performed for 54.6-min blocks and produced spectra of the three velocity components, represented by $S_u(\tau)$, $S_v(\tau)$ and $S_w(\tau)$, from which derived quantities $S_V(\tau) = S_u(\tau) + S_v(\tau)$ and $S_{TKE}(\tau) = 0.5 [S_u(\tau) + S_v(\tau) + S_w(\tau)]$ were determined. In Sect. 4.2, spectra are integrated to produce the standard deviation of the wind velocities over the entire 54.6-min time windows. In this case, they are referred to as σ_V^{1h} and σ_w^{1h} , and include both turbulent and non-turbulent contributions.

Two bulk Richardson numbers are considered for quantifying atmospheric stability. These are a “within-canopy Richardson number” and an “above-canopy Richardson number”, defined respectively as Ri_{can} and Ri_{top} :

$$Ri_{can} \equiv \frac{g}{\theta} \Delta z \frac{\theta_{34.9} - \theta_{24.5}}{(V_{34.9} - V_{24.5})^2} \quad (1a)$$

and

$$Ri_{top} \equiv \frac{g}{\theta} \Delta z \frac{\theta_{48.2} - \theta_{34.9}}{(V_{48.2} - V_{34.9})^2} \quad (1b)$$

where g is the acceleration due to gravity, θ is mean potential temperature, $\bar{\theta}$ is the average potential temperature in the air layer, the indices refer to the height in m at which the observations are being taken, and Δz is the height difference between the two levels of interest. Bosveld et al. (1999), Mammarella et al. (2007) and Oliveira et al. (2013) used a “canopy Richardson number”, with definitions generally similar to the one given above. A comparison (not shown) between Ri_{can} and Ri_{top} shows that they are poorly correlated. On the other hand, the choice of the lower level used in the definition Ri_{can} is not very important, as choosing a different lower level leads to values highly correlated to Ri_{can} as defined above. The choice of the bulk Richardson number instead of other dimensionless quantities such as the flux Richardson number Ri or the stability parameter $\zeta = (z - d) L^{-1}$ (where d is the displacement height and L is the Obukhov length) follows recent practice in micrometeorological research (Burns et al. 2010; Sun et al. 2012; Mahrt et al. 2013). It is based on the fact that the use of Ri and ζ may introduce self-correlation (Hicks 1978; Klipp and Mahrt 2004), because they are defined in terms of some of the same variables whose dependency on stability is currently being investigated. Furthermore, the turbulent fluxes used in the definitions of Ri and ζ are subject to low-frequency fluctuations that often make the values highly dependent on the temporal scale used for their evaluation.

3 Mean Profiles

3.1 Overall Averages

Some important characteristics of the turbulence structure can be inferred from the mean vertical profiles for the entire nocturnal period. The average nocturnal wind-speed profile (Fig. 1a) is similar to the mean profile under stable conditions observed at the same site by [Kruijt et al. \(2000\)](#). Within the canopy, a wind-speed maximum exists at $0.2h$, above most sub-canopy vegetation. Minimum wind speeds are observed, on average, near the surface and at the heights with larger leaf area density (from $0.5h$ to $0.7h$) as also noted by [Kruijt et al. \(2000\)](#). The average profiles of both horizontal (Fig. 1b) and vertical (Fig. 1c) components of

Fig. 1 Average nocturnal profiles of: **a** mean wind speed normalized by its value at canopy top; **b** σ_V normalized by friction velocity at canopy top; **c** σ_w normalized by friction velocity at canopy top; **d** velocity aspect ratio; **e** sensible heat flux, normalized by its absolute value at canopy top; **f** skewness of the longitudinal wind component; **g** skewness of the vertical wind component; **h** kurtosis of the longitudinal wind component; **i** skewness of the vertical wind component. In all panels, horizontal bars indicate the standard deviation of the observations at each level

TKE also present a local maximum at trunk space height, and a local minimum slightly below the regions of maximum leaf area density. Both horizontal and vertical velocity variances peak at canopy top, decreasing above it. This fact suggests that most turbulence near the canopy top is produced by shear and there is strong suppression of surface-layer turbulence by static stability. The vertical profiles of VAR (Fig. 1d) have values, on average, < 1 , showing that turbulence is predominantly horizontal at all heights. The contribution of the vertical velocity component to the TKE is largest at the heights of maximum foliar density, where VAR approaches 0.9. This is in agreement with the suggestion of [Vickers and Thomas \(2014\)](#) that, for turbulence, “...the canopy inhibits horizontal fluctuations more than vertical ones”. The kinematic sensible heat fluxes F_H are negative above the canopy and in the upper canopy, switching signs at lower levels (Fig. 1e). [Dupont and Patton \(2012\)](#) also observed this pattern for an orchard canopy with leaves. Such a flux profile suggests decreasing potential temperature with height from the surface to $0.7h$ with a stable layer above. It also implies a sensible heat-flux convergence at $0.7h$, which produces local turbulent warming at that height. The minimum temperature at $0.7h$ must be maintained by large radiative flux divergence driven by longwave emission from the leaves, whose density peaks at that height.

The vertical profiles of higher order statistics are quite similar to those observed under stable conditions by [Dupont and Patton \(2012\)](#) both with and without leaves, and have many of the features previously observed by [Kruijt et al. \(2000\)](#) at the same site. Within the upper canopy, the positive skewness of the streamwise velocity component (Fig. 1f) and negative skewness of the vertical velocity (Fig. 1g) are typical of canopy flows and indicate that the most intense events correspond to downward sweeping motion. As found by [Kruijt et al. \(2000\)](#), the u component is, on average, not skewed at the lower canopy. However, contrary to their finding, w is positively skewed at those levels indicating that the most intense events at those levels consist of upward motion. [Dupont and Patton \(2012\)](#) found the same and attributed it to the association with the positive heat flux at those levels (also observed in the present study). The skewness values found are much smaller than those typical for canopy turbulence during neutral and daytime conditions ([Dupont and Patton 2012](#); [Chamecki 2013](#)) and are, therefore, not characteristic of the strong sweep and ejection cycles produced by the wind shear at the top of the canopy. Kurtosis of the u component is < 3 at most heights, showing that the streamwise velocity component is not intermittent, on average. The opposite occurs for the w component, whose kurtosis exceeds 3 at all heights, therefore indicating intermittency of the vertical motion. The highest values of K_w occur above the canopy.

3.2 Dependence on Stability

It is clear that, although some aspects of the flow structures can be inferred from the average profiles, there is variability in all statistics, as indicated in Fig. 1. It is important, therefore, to understand the processes controlling such large variability. In most figures, dimensional values are presented without any normalization; the reason is that in very stable conditions, all scaling quantities become very small producing normalized values much larger in this stability regime than in the others. Contour lines increase logarithmically in all figures of this subsection, unless otherwise noted.

In Fig. 2, vertical profiles of V_{TKE} are classified in terms of V_{TKE} at canopy top (Fig. 2a), mean wind speed at canopy top (Fig. 2b), Ri_{can} (Fig. 2c) and Ri_{top} (Fig. 2d). The vertical minimum in V_{TKE} at $0.5h$ disappears when turbulence at canopy top is sufficient; namely, when V_{TKE} approaches 0.35 m s^{-1} at canopy top, it becomes nearly constant with height from $0.2h$ to $0.5h$ (Fig. 2a), a possible consequence of enhanced turbulent mixing towards lower levels within the canopy. Such a vertical homogenization of V_{TKE} in the lower canopy

Fig. 2 Contours of the average turbulence velocity scale, according to the logarithmic colour scale, as a function of height and: **a** turbulence velocity scale at canopy to; **b** mean wind speed at canopy top; **c** Ri_{can} and **d** Ri_{top}

is also evident when the data are classified in terms of Ri_{can} (Fig. 2c), but less so when Ri_{top} is used (Fig. 2d), indicating that the upper canopy stability opposes the downward transport of turbulence more so than the stability above the canopy. On average, the suppression of the V_{TKE} minimum at $z = 0.5h$ occurs when $Ri_{can} < 0.2$. As high wind speeds tend to occur with intense turbulence and less stable conditions, use of the wind speed at canopy top to classify the data also allows the identification of many of these patterns (Fig. 2b). These results differ from those obtained by [Krujijt et al. \(2000\)](#), where a local maximum of σ_u at $0.25h$ was found for all stability classes, from very stable to very unstable. In agreement with the present findings, [Launiainen et al. \(2007\)](#) found a local maximum in both σ_u and σ_w at $0.1h$ for stable conditions within a pine forest, which disappeared in neutral conditions. [Dupont and Patton \(2012\)](#) obtained similar results for σ_u , but not for σ_w . Given that the vertical profile of V_{TKE} shows a systematic dependence on Ri_{can} , but not on Ri_{top} , the former is used as the stability indicator in the forthcoming analysis.

The greatest wind shear occurs in weak stability from $0.7h$ to $1.0h$ ($Ri_{can} < 0.2$) (Fig. 3a). Within the canopy, the value of the minimum wind speed that occurs between $0.5h$ and $0.7h$ becomes even smaller in more stable conditions. Above the canopy, σ_w is an order of magnitude smaller during the most stable cases than in nearly neutral conditions (Fig. 3b).

Fig. 3 Contours of the average: **a** mean wind speed; **b** σ_w ; **c** velocity aspect ratio; **d** sensible heat flux, as a function of height and Ri_{can} and according to colour scale in each panel. All colour scales are logarithmic, except in **d**, where it is linear

The σ_w maximum at $Ri_{can} < 0.2$ is restricted to heights above $0.5h$, indicating that the intense stability of the upper canopy damps the turbulent fluctuations, preventing their propagation towards the surface. As stability becomes stronger, σ_w becomes smaller than 0.1 m s^{-1} above the canopy, and even smaller within it. However, σ_w still presents a small secondary (or local) maximum at $0.2h$, which is intensified under more stable conditions. The vertical profile of VAR is also highly dependent on stability (Fig. 3c). Its maximum near $0.7h$ exceeds unity in near-neutral conditions, indicating that, in this case, vertical velocity fluctuations are larger than the horizontal ones. Therefore, if such a maximum implies that the canopy inhibits horizontal turbulence fluctuations more than vertical fluctuations as suggested by [Vickers and Thomas \(2014\)](#), this inhibition is not as effective in very stable conditions.

The height at which F_H changes sign also depends on stability (Fig. 3d). Under very stable conditions ($Ri_{can} > 1$), the layer with upward heat flux extends from the surface to $0.8h$, while in neutral condition it only reaches $0.6h$. This result shows that stronger stratification in the upper canopy reduces the penetration of turbulence into the forest, reducing levels of vertical mixing towards the surface and increasing the layer dominated by the positive heat flux from the ground.

The vertical profiles of higher order statistical moments of turbulence also exhibit significant stability dependence (Fig. 4). The upper canopy positive maximum of Sk_u is enhanced

Fig. 4 Contours of the average: **a** skewness of the longitudinal wind speed; **b** skewness of the vertical wind speed; **c** kurtosis of the longitudinal wind speed; **d** kurtosis of the vertical wind speed, as a function of height and Ri_{can} and according to colour scale in each panel. All colour scales are linear. In all panels, the yellow line represents the contour for zero heat flux, shown in Fig. 3d

at $Ri_{can} < 0.2$, both in terms of magnitude and vertical extension (Fig. 4a). While positive Sk_u extends from the surface to $1.2h$ in near-neutral conditions, it is confined in the lower canopy in very stable conditions. Sk_w is generally positive in the lower canopy, regardless of stability (Fig. 4b). At higher levels, however, a maximum of Sk_w above the canopy occurs in weakly stable conditions, while negative values occur when it is very stable. An opposite pattern is observed in the upper canopy, between $0.6h$ and h , where Sk_w is negative for $Ri_{can} < 0.5$, and positive when it is more stable. The contour line for zero heat flux (from Fig. 3d) is also shown in Fig. 4. The typical canopy signature on the skewness, with negative Sk_w and positive Sk_u exists only above that line, at levels where the heat fluxes are negative. In the region of positive heat flux, located below that line, there is small Sk_u and positive Sk_w , consistent with the occurrence of thermal convection. These patterns indicate that the zero heat-flux level can be used to separate the region where turbulence has the characteristics of canopy turbulence from that where it has characteristics of thermal convection.

For both components, maximum kurtosis occurs at $0.7h$, for $Ri_{can} < 0.2$, showing that highly intermittent turbulence occurs at that height, but only for weakly stable conditions.

Above the canopy, the maximum in K_w at $0.2 < Ri_{\text{can}} < 1$ (with a corresponding K_u maximum, although smaller in magnitude) indicates that turbulence is more intermittent under intermediate stability regimes.

These findings are in agreement with stable boundary-layer theory, which indicates continuous turbulence at the neutral limit, and almost total turbulence suppression at the very stable limit, with intermittency being favoured in between ([Mahrt 1999](#); [Van de Wiel et al. 2002](#)). Such intermittency may originate above the surface, possibly due to a low-level jet ([Sun et al. 2004](#); [Cuxart and Jimenez 2007](#)) propagating downward or by interactions between the mean flow, turbulence and stability at the lower boundary, in which case the intermittency would propagate upwards ([Van de Wiel et al. 2002](#); [Costa et al. 2011](#)). This is a possible mechanism at this site because nocturnal low-level jets have been observed above the Amazonian canopy by [Greco et al. \(1992\)](#), [Oliveira and Fitzjarrald \(1993\)](#) and by [Zeri and Sá \(2011\)](#), and have been numerically simulated by [Oliveira and Fitzjarrald \(1994\)](#). These previous studies reported wind-speed maxima up to 18 m s^{-1} , typically at heights between 100 and 500 m. The present results also indicate that upper canopy intermittency, which may be driven by inflection point instability ([Raupach et al. 1996](#)), is a different process than the intermittency observed above the canopy, occurring at different ranges of stability. Figure 4 shows that the most intense turbulent events, characterized by intense gusts of downward motion, only occur in the upper canopy when stability is weak. These events rarely reach heights below $0.5h$ at any stability and they also rarely occur when $Ri_{\text{can}} > 0.5$ at any height within the canopy.

4 Temporal Scales

4.1 Overall Averages

Above the canopy, the average multiresolution spectra of the horizontal wind components present a well-defined spectral gap at temporal scales around 100 s (Fig. 5). This gap separates the smaller scales of turbulent motion, whose spectral peak occurs at 20 s, from the larger-scale processes, which are typically non-turbulent and predominantly horizontal. Based on this observation, the expressions “non-turbulent” and “submeso” are used here to refer to processes that occur at time scales longer than the spectral gap. The bimodal horizontal spectra prevailing above the forest affect the turbulence in the upper canopy, where two peaks are visible above $0.75h$. Below this height, a single spectral peak exists from $0.2h$ to $0.6h$ at time scales larger than 100 s. This is similar to the scale of the non-turbulent maximum observed above the canopy, suggesting that it could be caused by the direct transference of low-frequency processes from above the forest downward. At those heights the horizontal spectra increase sharply between time scales of 5 and 100 s, suggesting that the canopy efficiently suppresses horizontal wind fluctuations at smaller scales, as observed by [Kruijt et al. \(2000\)](#). At a height $0.2h$, a secondary peak is visible at small time scales, around 1 s. Such a high-frequency maximum was previously observed ([Kruijt et al. 2000](#)) and attributed to wake turbulence production by the foliage motion ([Cava and Katul 2008](#)).

For the vertical direction, the average velocity spectrum above the canopy peaks between time scales of 5 and 10 s (Fig. 6) without a spectral gap, confirming the horizontal character of the long-time scale maximum observed in Fig. 5. Within the canopy, the scale of the vertical velocity spectral peak progressively increases as one moves towards the surface, reaching 100 s at $0.2h$. A sharper spectral peak exists between $0.4h$ and $0.6h$, at time scales between

Fig. 5 Contours of the average multiresolution spectra of the horizontal wind components, S_V ($10^{-3} \text{ m}^2 \text{ s}^{-2}$), as a function of height and time scale, according to the logarithmic colour scale at the right side. Blue dotted line indicates the canopy top

Fig. 6 The same as in Fig. 5, but for the spectra of the vertical velocity component, S_w ($10^{-3} \text{ m}^2 \text{ s}^{-2}$)

10 and 100 s. At the long limit of scales considered here, approaching 1000 s, there is a significant reduction in the vertical fluctuations at all vertical levels, contrasting with that observed for the horizontal components in Fig. 5. The upward heat flux observed within the

Fig. 7 The same as in Fig. 5, but for the sensible heat flux, C_{wT} (10^{-4} K m s $^{-1}$). The colour scale is linear

canopy also occurs at long time scales, peaking between 100 and 300 s (Fig. 7). Although the longer-than-100 s time-scale maximum of the horizontal turbulent components within the canopy (Fig. 5) may be attributed to the downward propagation of the non-turbulent motion observed at those scales above the canopy, such a mechanism does not explain the vertical component (Fig. 6), which lacks energy at these long time scales above the canopy. It is possible, therefore, to speculate that the horizontal energy at scales longer than 100 s initiates vertical velocity fluctuations. The most likely mechanism to produce such a transfer is pressure redistribution through the return-to-isotropy terms of the equations for the variances of the wind components, though determination of this term is not possible with the observations available at the site. The enhanced vertical variances drive the upward heat-flux maximum at those long time scales (Fig. 7), and, as such, a positive heat flux constitutes further production of vertical velocity fluctuations, creating a positive feedback mechanism. In this hypothesis, the process, with its long time scales, is initiated by the downward propagation of non-turbulent modes from above the canopy, but is enhanced and maintained by the organized convective motion that occurs in the unstable environment of the lower canopy. This mechanism explains why, in several previous studies, the temporal scale of the turbulent flow varies largely between the top of the canopy and its interior, while in others cases this difference is severely reduced. As an example, Dupont and Patton (2012) found that the frequency of the vertical velocity spectral peak above the canopy was an order of magnitude larger than near the surface during the foliated period, when the dense canopy caused positive heat flux at its lower levels. In contrast, when leaves were absent, the upward heat flux was likewise absent in the lower canopy and the time scales of the spectral peaks were much more similar across the vertical levels. Oliveira et al. (2013) observed an increase of an order of magnitude of the time scales of CO $_2$ fluxes and latent heat fluxes within the canopy with respect to those observed above it, at a site where the heat flux near the surface was also positive. These results demonstrate that scalar fluxes may also be affected by this process.

4.2 Dependence on Stability

To address how the scales of the flow depend on stability, TKE spectra have been classified in terms of Ri_{can} , in four classes, each containing an equal number of cases (Fig. 8). The most significant differences are observed above the canopy, where it is clear that stability has a strong influence on the magnitude of the turbulence-related peak (at time scales between 10 and 30 s). In fact, only in the two least stable classes (Fig. 8a, b) is such a turbulent peak distinguishable in the average spectra. On the other hand, the low-frequency peak at the long time-scale limit of the analysis is independent of stability, having similar magnitude and scales for all stability classes considered. The turbulent fluctuations observed above the canopy in the two least stable classes (Fig. 8a, b) penetrate down the upper canopy to $0.6h$, where the maximum associated with the peak above the canopy is still visible. Closer to the surface (at approximately $0.4h$), stability affects the spectral distribution of the velocity fluctuations more than their peak magnitude. In the least stable class (Fig. 8a), a broad maximum occurs at time scales larger than 10 s, while in the most stable class (Fig. 8d) the magnitude of the maximum is similar, but only exists at time scales larger than 50 s.

Fig. 8 Contours of the average multiresolution spectra of turbulent kinetic energy, as a function of height and time scale, according to the logarithmic colour scale at the middle, for four classes of stability, as given by Ri_{can} .

Fig. 9 Average dependence of σ_V on: **a** Ri_{can} ; **b** mean wind speed at canopy top; **c** kinetic energy at canopy top for time scales from 0.05 to 25.6 s; **d** kinetic energy at canopy top for time scales from 409.6 to 1638.4 s. In all panels, each colour refers to a vertical level, as specified in legend

To understand the processes controlling the turbulence intensity within the canopy, and to address the extent to which the turbulence and low-frequency fluctuations that occur above the canopy are transferred to lower levels, horizontal and vertical velocity variances are compared to four different quantities: Ri_{can} , to address the role of upper canopy stability (Figs. 9a, 10a); mean wind speed at canopy top (Figs. 9b, 10b); turbulent kinetic energy at canopy top, given by the integral of the V_{TKE} spectra from scale of 0.05 to 25.6 s (Figs. 9c, 10c) and low-frequency kinetic energy at canopy top, given by the V_{TKE} spectrum integrated between time scales of 409.6 to 1638.4 s (Figs. 9d, 10d). In this analysis, the horizontal and vertical velocity variances are determined from the integration of the respective multiresolution spectra over all time scales, so that they include fluctuations as long as the temporal window used for the decomposition (54.6 min) and for this reason are referred to as σ_V^{1h} and σ_w^{1h} , respectively. This is done to include the low-frequency contribution to the kinetic energy and identify its effect on the activity observed within the canopy.

Horizontal turbulence, given by σ_V^{1h} (Fig. 9) is highly dependent on stability, wind speed and TKE at higher levels within the canopy. This is shown by the fact that, at 0.7h, the average

Fig. 10 The same as in Fig. 9, but for σ_w

σ_V^{1h} is 50 % larger for $Ri_{can} < 0.2$ than it is for $Ri_{can} > 5$ (Fig. 9a). Similar distinctions are observed between higher and lower wind-speed conditions (Fig. 9b) and conditions of greater and lesser turbulence (Fig. 9c). At the same height, the horizontal fluctuations are less dependent on the low-frequency kinetic energy (Fig. 9d). This pattern changes as one moves towards the surface, such that at 0.2h the average σ_V^{1h} is not significantly different when $Ri_{can} < 0.2$ than it is when $Ri_{can} > 5$. The factor that most consistently controls σ_V^{1h} near the surface is low-frequency kinetic energy (Fig. 9d), and at 0.2h, σ_V^{1h} increases steadily with low-frequency kinetic energy. These results suggest that low-frequency, submeso motions above the canopy induce an important fraction of the horizontal velocity fluctuations in the lower canopy. It also shows that the increase of σ_V with stability in the lower levels is caused by the increased relevance of low-frequency fluctuations above the canopy in very stable conditions. The present analysis is not capable of identifying the physical mechanism through which such a relationship between low-frequency motions above the canopy affect horizontal velocity fluctuations in the lower canopy. Thomas (2011) identified a similar relationship and suggested microfronts as a possible low-frequency mechanism causing wind fluctuations near the surface, during a study of horizontal coherence of wind and temperature

Fig. 11 **a** Average multiresolution spectra of the horizontal wind components at the vertical levels of 7.0 and 13.5 m for three classes of Ri_{can} , according to legend; **b** same as in (a), but for three classes of kinetic energy at canopy top in time scales from 409.6 to 1638.4 s, according to legend; **c** same as in (a), but for the spectra of the vertical wind component; **d** same as in (b), but for the spectra of the vertical wind component

fluctuations. Vertical turbulence is more dependent on stability in the upper canopy than are the horizontal components. This fact is illustrated by the fact that, at $0.7h$, average σ_w^{1h} more than doubles between $Ri_{\text{can}} < 0.2$ and $Ri_{\text{can}} > 5$ (Fig. 10a). At lower levels, such a dependency is largely reduced, so that at $0.2h$ σ_w^{1h} shows almost no difference between those two classes, although it shows a minimum (as does σ_v^{1h} , Fig. 9a) at intermediate stabilities. In contrast to the horizontal components of turbulence, σ_w^{1h} is independent on low-frequency kinetic energy at all heights (Fig. 9d).

The average spectra of horizontal and vertical velocity fluctuations from $0.2h$ to $0.4h$ (two lowest levels) are shown in Fig. 11 for three classes of Ri_{can} and of low-frequency kinetic energy. Upper canopy stability, given by Ri_{can} , affects both horizontal (Fig. 11a) and vertical turbulence (Fig. 11c) at all temporal scales. Most of the turbulent increase in less stable conditions occurs at time scales smaller than 100 s, while in more stable conditions there is an enhancement of both horizontal and vertical velocity spectra at time scales longer than 100 s. On the other hand, the effects of low-frequency kinetic energy are most important for the horizontal components (Fig. 11b) and at time scales longer than 50 s, providing evidence that low-frequency fluctuations of the horizontal wind components are transferred from above the canopy to lower levels near the surface.

Fig. 12 **a** Contours of the average mean vertical velocity as a function of height and net radiation at 39.0 m, according to *colour scale*; **b** same as in (a), but for contours of σ_w ; **c** average multiresolution spectra of the horizontal wind components at the vertical levels of 34.9, 40.4 and 48.2 m for three classes of net radiation, according to *legend*; **d** same as in (c), but for the spectra of the vertical wind component

4.3 Dependence on Net Radiation

During clear sky nights, the elevated radiative loss at canopy top produces negative values of net radiation. In this subsection, we investigate whether this radiative cooling process is sufficient to drive downward motion of cold air from the canopy top to its interior. The temporal scales of such movements are also investigated.

When the mean vertical motion is shown as a function of net radiation and height (Fig. 12a), the most intense downward motion at canopy top occurs when there is large net radiative loss at a height 39 m. This result supports the idea that radiative cooling at canopy top locally cools the air to an extent that is sufficient to drive mean downward motion. Vertical turbulence is also affected, as σ_w shows two peaks at canopy top (Fig. 12b). The first, at near-zero net radiation, is expected since it occurs along with the near-neutral conditions that dominate under these circumstances. The second peak, however, occurs at the most negative limit of net radiation, and is unexpected from a simple reasoning based on stability, since very stable conditions and, therefore, reduced turbulence intensity, are expected to prevail at canopy top when there is large radiative loss.

When multiresolution spectra of horizontal and vertical components of the wind are classified in terms of net radiation (Fig. 12c and d, respectively), it is clear that such downward

Table 1 Average values of Ri_{can} , mean wind speed and turbulence velocity scale at 34.9 m for all nights

Night	Ri_{can}			V at 34.9 m (m s^{-1})			V_{TKE} at 34.9 m (m s^{-1})		
	Min	Max	Mean	Min	Max	Mean	Min	Max	Mean
25 July	0.031	9761	152.9	0.036	1.28	0.56	0.069	0.33	0.18
4 September	0.16	25.8	1.45	0.34	1.95	1.07	0.097	0.57	0.32
27 November	0.011	0.24	0.070	0.57	1.90	1.20	0.34	1.19	0.69

motion induced by large radiative loss at the canopy top affects the turbulent scales of motion more than its low-frequency scales. In both cases, the class with large radiative loss is more energetic than the intermediate class on time scales smaller than 50 s. Longer time scales are unaffected by this process, as shown by the spectra of the horizontal wind components, which is least energetic for the class with large radiative loss for scales longer than 100 s.

The net negative vertical velocity and the peak in σ_w at the negative limit of net radiation are consistent with turbulent motions driven by negative buoyancy near the canopy top.

5 Case Studies

To illustrate how the average characteristics described in the previous sections evolve in time, turbulence data from three nights with distinct characteristics are analyzed in detail. The first, 25 July 2014, had the weakest turbulent activity (given by the average V_{TKE} for the entire period) among all nights considered. The second, 27 November 2014, was the most turbulent. A third night, 4 September 2014, with intermittent turbulence, was chosen to show how the different temporal scales of the turbulent flow above the canopy affect the activity within the canopy. Maximum, minimum and average values of Ri_{can} , wind speed and V_{TKE} at canopy top are presented at Table 1 for each study night. It is important to stress that these are examples that depend on the particular evolution of turbulence and stability at the nights under consideration, so that they should not be considered as typical conditions.

5.1 Calm and Turbulent Nights

Above the canopy, there is a large contrast between the weakly turbulent night of 25 July (Fig. 13a) and the strongly turbulent case of 27 November (Fig. 13b). At canopy top, average σ_V is 0.12 m s^{-1} on 25 July, while on 27 November it was four times larger, reaching 0.49 m s^{-1} . The contrast is even larger for σ_w at the same height, a 5.5-fold increase, from 0.065 m s^{-1} to 0.36 m s^{-1} between both nights. On the other hand, the average turbulent activity is similar between the two nights in the lower canopy: at $0.2h$ σ_V increased only by 42% from the calm to the turbulent night, while σ_w increased by 13%. The vertical profiles of these two variables are shown in Fig. 12c, d. The difference in turbulent activity above the forest has a much stronger effect on turbulence in the upper half of the canopy than in the lower levels (Fig. 13c, d). Despite the much larger fluctuations of the horizontal components of wind velocity on 27 November than on 25 July, above the canopy both components present a well-defined prevailing direction on 27 November, but not on 25 July. On the turbulent night above the canopy (Fig. 13b, black lines), the mean u component is always positive, while the mean v component is always negative, although it is also possible to identify longer-period fluctuations on both components. Conversely, during the calm night (Figure 13a, black lines),

Fig. 13 **a** Time series of the three velocity components along the night of 25 July 2014, at 34.9 m (black lines) and at 7 m (blue); **b** same as in (a), but for the night of 27 November 2014; **c** average vertical profiles of σ_v for the nights of 25 July 2014 (red lines) and 27 November 2014 (purple); **d** same as in (c), but for σ_w

both u and v components at canopy top switch signs many times throughout the night, with no clear periodicity. It is possible to identify different periods with prevailing flow from all quadrants during the calm night. Wind direction is highly variable at 0.2h (Fig. 13a, b, blue lines) during both nights. These differences illustrate the relative role of turbulent and low-frequency processes, which is seen more clearly by the average multiresolution spectra of the horizontal (Fig. 14a, d) and vertical (Fig. 14b, e) velocity components. The absence of intense turbulent fluctuations above the canopy on 25 July is shown by the lack of a turbulent peak in S_v (Fig. 14a), which presents only a low-frequency maximum at time scales near 1000 s. This peak corresponds to the long-term fluctuations responsible for wind direction shifts (Fig. 13a). In the vertical direction, a maximum associated with turbulent motion is present above the canopy on 25 July (Fig. 14b), with a temporal scale around 10 s, but its magnitude is $<10^{-3} \text{ m}^2 \text{ s}^{-2}$. In contrast, on 27 November, both S_v (Fig. 14d) and S_w (Fig. 14d) exhibit large maxima above the canopy and are associated with turbulent motions at time scales between 10 and 50 s. In the horizontal directions, a low-frequency maximum

Fig. 14 **a** Average multiresolution spectra of the horizontal wind component as a function of height and time scale, for the night of 25 July 2014; **b** same as in **(a)**, but for the vertical wind component; **c** same as in **(a)**, but for the sensible heat-flux cospectra; **d** same as in **(a)**, but for 27 November 2014; **e** same as in **(b)**, but for 27 November 2014; **f** same as in **(c)**, but for 27 November 2014. Values for panels **a**, **b**, **d** and **e** are given by logarithmic colour scale in their middle. Values for panels **c** and **f** are given by linear colour scale in between them. The first colour in this scale covers a longer range than the others for better visualization

above the forest also occurs on 27 November, at time scales around 1000 s. The low-frequency maximum has similar magnitudes in both cases.

On 27 November, the turbulent maximum observed above the canopy in S_V and S_w is only visible within the canopy at heights above $0.6h$. At lower levels, the time scale of the most energetic eddies shifts to longer scales, in the same manner observed in the average spectra shown in Figs. 5 and 6. Near the surface, the magnitude and time scales associated to the spectral maxima are similar between the two nights for horizontal and vertical velocities. At those heights, the most important difference is that S_V exhibits a broader peak on 27 November than on 25 July, and it causes the small difference in σ_V at lower levels between the two nights (Fig. 13c). On 25 July, the time scale of the S_V maximum near the surface is the same observed above the forest (Fig. 14a), once more suggesting that these processes propagate towards the lower canopy more easily than turbulence does. In the vertical direction (Fig. 14b), the S_w peak at lower levels not only occurs at longer time scales but also exceeds in magnitude the corresponding peak above the canopy. This suggests that part of the energy is being transferred from other TKE components or being locally produced by shear or buoyancy. The upward heat flux within the canopy also peaks at time scales near 100 s (Fig. 14c). It is possible, therefore, that a fraction of the vertical velocity fluctuations in the lower canopy is produced by buoyancy, but their occurrence at large time scales suggests that

they are likely initiated by the energy of the horizontal components, through the mechanism proposed at the end of Sect. 4.1.

5.2 Intermittent Night

Intermittent turbulence above the canopy is a common occurrence at the site, following that typically observed at other forests (Hollinger and Richardson 2005; Van Gorsel et al. 2011; Oliveira et al. 2013; Cava et al. 2014). To illustrate the intermittence of turbulence during the nighttime, 5 h of data from 4 September 2014 are analyzed in detail (Fig. 15a). In that period, three intermittent events occurred, during which turbulence activity increased and was later suppressed. These events can be identified by large fluctuations of all velocity components, but especially in w , occurring approximately at 2330, 0040 and 0200 LST. Their non-periodic nature is evidenced by their variable magnitude and duration. The comparison focuses on the two latter events (labeled “a” and “b” in Fig. 15a). The period chosen in both cases corresponds to the time window of the multi-resolution decomposition that best captures each of the events. From the vertical velocity time series, it is clear that above the canopy event b was much more turbulent than event a . This is confirmed by the spectra of both horizontal (Fig. 15b) and vertical velocity components (Fig. 15c). Event b (black dashed lines) was more energetic than event a (black solid lines) for all time scales shorter than 20 s. On longer time scales, however, event a was more energetic than event b , especially in the horizontal directions. This increase in low-frequency energy in event a is associated with wavelike fluctuations in both horizontal velocities (Fig. 15a). Such oscillatory behaviour is evident during period a , but almost entirely absent in period b , which is dominated by turbulence. At lower levels, on the other hand, both horizontal and vertical velocity spectra

Fig. 15 **a** Time series of the three wind components along the night of 4 September 2014, at heights of 34.9 m (black lines) and at 7 m (blue); **b** multi-resolution spectrum of the horizontal wind components for events “a” (solid lines) and “b” (dashed) at 34.9 m (black) and at 7 m (blue); **c** same as in (b), but for the spectrum of the vertical wind component

are similar between the two nights for temporal scales as long as 20 s, while for longer time scales, there is much more energy in event *a* than in event *b*. This contrast is more evident in S_V , but can also be clearly noticed in S_w .

This case shows that wave activity is one of the physical mechanisms through which processes with longer temporal scales propagate downward from above the canopy to its interior. In contrast, both on 25 July and 27 November (Figs. 13, 14), the motion in the lower canopy is dominated by processes with long temporal scales, without the occurrence of waves in either case. Therefore, the present analysis indicates that waves are one possible mechanism for energy propagation towards the surface through the canopy, but likely not the only one. Further investigation is needed to clarify this matter.

6 Conclusion

This study provides a detailed analysis of the structure of nocturnal turbulence above and within a forest in Amazonia. In comparison to previous studies ([Fitzjarrald and Moore 1990](#); [Kruijt et al. 2000](#)), the present dataset represents a much longer time period and with higher vertical resolution. As a consequence, important characteristics of the interactions between the air above the canopy and in its interior have been identified. In particular, it is established that the temporal scale of the flow near the surface is longer than above the forest. The results presented indicate that the nighttime flow inside the Amazon forest is a combination of quasi two-dimensional low-frequency motions that propagate from the canopy top and vertical turbulence locally produced by buoyancy in the unstably stratified region.

Important evidence to support this conclusion comes from the spectral analysis, which shows that the temporal scale of the horizontal velocity fluctuations near the surface coincides with a maximum observed above the canopy at time scales longer than 100 s. This spectral maximum has been recently associated with non-turbulent flow structures that become relevant in conditions of enhanced static stability, and which have been referred recently as submeso flow ([Mahrt 2009](#); [Acevedo et al. 2014](#)). However, it has not been established as to whether the flow within and above the canopy is actually correlated at those longer temporal scales. It is essential to establish that such longer modes propagate from the top of the canopy towards its interior.

This finding may be particularly relevant to the quantification of the total nocturnal exchange of quantities, such as energy and trace gases between the forest and the atmosphere. This is because the present results indicate a substantial difference in the character of the exchange between weakly stable and very stable nights. In the former case, fully-developed turbulence prevails, so that most of the exchange tends to happen during cycles of sweeps and ejections, as found previously ([Gao et al. 1989](#); [Collineau and Brunet 1993](#); [Lu and Fitzjarrald 1994](#), among others). In the latter case, on the other hand, the static stability above the canopy suppresses the turbulent mixing, so that non-turbulent fluctuations may become dominant at time scales as small as 10 s ([Acevedo et al. 2014](#)). Here, it has been shown that the propagation of this energy towards the interior of the canopy may be fairly efficient, driving relevant scalar fluxes near the surface on scales longer than 100 s. The present results allow speculation that, at the lower canopy, these fluxes may be at least as large as those that occur on weakly stable nights. It is important, therefore, to extend the present analysis to turbulent fluxes of scalars, such as carbon dioxide. It has also been shown that the motion within the canopy can be initiated by the radiative cooling at canopy top. However, it is still important to address how this process affects the nocturnal energy budget.

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