

Table 1. Micrometeorological variables within and above the forest measured on both the Tall and Instant (bold) towers. Adopted from González-Armas et al. (2025).

Variable	Instrument	Frequency	Height (m)
Air T and RH	Thermohygrometer (IAKM, Galltec, Germany)	1 min	25, 35, 50, 73, 81, 100, 127, 151, 223, 247, 274, 298
Air T	Thermometer (Hmt337, Vaisala, Finland)	1 min	5, 15
CO ₂ concentrations	Picarro (G2401, Picarro Inc. Santa Clara, Ca, USA)	30 min	4, 24, 38, 53, 79, 321
U and TKE	3D Ultrasonic Anemometer (CSAT-3b, Campbell Scientific instrument Inc., USA)	30 min	5, 25, 50, 81, 196
U and TKE	3D Ultrasonic Anemometer (Thies, Germany)	30 min	15, 35, 100, 127, 151, 172, 223, 247, 274, 298
LW _{out} and LW _{in}	Pyrgometer (CGR4, Kip and Zonen, the Netherlands)	10 min	75

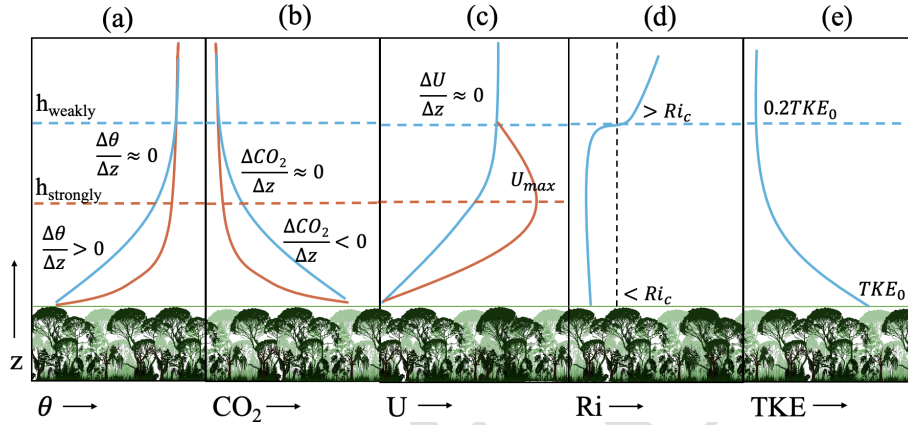


Figure 1. Five criteria for estimating the nocturnal boundary layer height (h_{NBL}): (a) potential temperature (θ), (b) CO₂, (c) horizontal wind speed (U), (d) Richardson number (Ri), and (e) turbulent kinetic energy (TKE). The blue lines represent the weakly stable regime and the red lines the strongly stable regime. The blue and red horizontal dashed lines indicate the estimated h_{NBL} of the weakly and strongly stable profiles, respectively. The vertical dashed black line in (d) represents the critical Richardson number (0.25).

2.3 Temporal evolution of stability within and above the canopy

To understand the importance of canopy-atmosphere interactions, we studied the temporal evolution of atmospheric stability, focusing on both the in-canopy atmosphere, the stable layer aloft and the interaction at the canopy-atmosphere interface. To compare the stability of the in-canopy and stable layer aloft, we introduce a bulk representation of Ri for both layers based on the work of Oliveira et al. (2013). For the stable layer above canopy, the Richardson number (Ri_{atm}) is defined by Eq. (1), and for the in-canopy layer including the canopy roughness layer, the Richardson number (Ri_{can}) is determined by Eq. (2). In the equations, the subscripts of θ and U , indicate the measurement heights of θ and U , g is the gravitational acceleration (9.81 m s^{-2}), Δz is the difference in height between the two heights indicated in the subscripts and θ_{ref} is a reference temperature, taken as the potential temperature at the highest measuring level on ATTO tall tower (298 m).

$$Ri_{\text{atm}} = \frac{g \Delta z}{\theta_{\text{ref}}} \frac{\theta_{81} - \theta_{35}}{(U_{81} - U_{35})^2} \quad (1)$$

$$Ri_{\text{can}} = \frac{g \Delta z}{\theta_{\text{ref}}} \frac{\theta_{35} - \theta_5}{(U_{35} - U_5)^2} \quad (2)$$

2.4 Quantifying canopy-atmosphere exchange of CO₂ and heat

To assess and quantify the coupling and exchange between the in-canopy atmosphere and the stable layer aloft, we evaluated the governing equations of TKE, heat and CO₂ for the selected strongly and weakly stable periods (Sect. 2.4.1).

To calculate the budgets using only our observations, we had to make the following two assumptions: (1) The atmosphere is horizontally homogeneous ($\frac{\partial \psi}{\partial x} = \frac{\partial \psi}{\partial y} = 0$) and (2) subsidence (vertically advective term) is negligible ($\bar{w} = 0$). In addition, we aligned the coordinate system with the mean wind such that only the u component of the wind remained. We evaluated the governing equations using 30 min averaged data (indicated by the overbars). Vertical derivatives were calculated using one level above and below the desired height, except for the vertical divergence of the CO₂ flux, for which, due to a lack of measurements at all levels, a first-order method was employed, which may affect the accuracy of the calculations. The residual (R), or non-closure of the budgets, represents the uncertainty due to neglected terms of

age of 134 ± 29.5 m (Fig. 2e). We observed that h_{Ri} is often significantly lower than h_θ and h_{CO_2} , during turbulent periods, raising questions as to whether this method accurately reflects the underlying dynamics. The h_{TKE} is calculated to be 193 ± 50.9 m (Fig. 5f), which is higher than average. The significant variation may be partly attributed to the limited number of estimations. Additionally, we observe that the TKE often did not reduce by 80 % within the vertical profile. It is possible that the frequent occurrence of very stable conditions complicated the application of this method during the CloudRoots-Amazon22 period, as previous studies at the ATTO site have yielded reliable results using similar turbulence-based methods (Mendonça et al., 2025; Souza et al., 2026).

The h_{NBL} above the Amazon rainforest has been investigated in a limited number of studies. All these studies reported h_{NBL} values ranging from 80 to 250 m (Oliveira et al., 2018; Carneiro and Fisch, 2020; Mendonça et al., 2025, 2026; Souza et al., 2026). We compare our results a bit more carefully to the three studies that also used ATTO data. Two estimated h_{NBL} utilising the vertical turbulence observations installed at ATTO (Mendonça et al., 2025; Souza et al., 2026) and one using the NLLJ nose height (Mendonça et al., 2026). The studies report h_{NBL} values ranging from 81 to 247 m, with Souza et al. (2026) specifically finding 120 ± 28 m for the 2022 dry season. While this range is consistent with our results, slight differences persist. Such variations were anticipated, given that the different criteria are governed by distinct physical processes.

Overall, we consider the range, derived from all criteria together, to yield a stable representation of h_{NBL} during variable turbulent regimes. With this finding, we also show that gradient methods give reliable estimations of h_{NBL} . These gradient methods can be useful in NBL studies, particularly for locations where no vertical turbulence profiles are available.

3.2 Temporal evolution of stability within and above the canopy

The temporal evolution of the vertical atmospheric structure differs within and above the canopy (Fig. 3). At the canopy – atmosphere interface, between 25 (24) and 81 (79) m, we observe an increase (decrease) in the θ (CO₂) gradient from 0.040 K m^{-1} (-0.09 ppm m^{-1}) to 0.046 K m^{-1} (-0.51 ppm m^{-1}) throughout the night. In contrast, below the canopy (between 5–25 m), the θ (CO₂) gradient decreased (increased) from 0.067 K m^{-1} (-1.77 ppm m^{-1}) to 0.015 K m^{-1} (-0.69 ppm m^{-1}).

Above the canopy, the atmospheric stratification strengthens during the night, while the within-canopy layer mixes over time. After sunset (18:00–20:00 LT), below the canopy, a shallow stable daytime layer was present within the canopy. The shallow stable layer has been observed to form during daytime by González-Armas et al. (2025). Throughout the

night, the atmospheric stratification within the canopy layer weakens, enabling air mixing and thereby leading to better mixed CO₂ profiles throughout the canopy layer.

We hypothesise that the mixing within the canopy layer originates from the cold layer forming in the dense canopy crown, roughly between 15–30 m (Gomes Alves et al., 2023). In this layer, the vegetation biomass cools down, creating a locally cold layer, which promotes sinking motions and enhances turbulence in the canopy layer. Enhanced turbulence within the canopy has also been reported in previous studies (Kruijt et al., 2000; Santos et al., 2016). Santos et al. (2016) found that a radiative imbalance of $\leq -40 \text{ W m}^{-2}$ is needed to make this sinking motion possible, this threshold was met during most nights during CloudRoots-Amazon22 campaign (not shown). In addition, we can confirm the downward motion of cold air by turbulence observations. When we inspect the selected hours used in Sect. 3.3 (Table 3) as an example, we observed a positive heat flux ($w'\theta'$) during S2, S3, S4, and W3 within the canopy layer, which can be associated to the downward movement of cold air ($w' < 0$ and $\theta' < 0$) (Appendix C).

In addition to the observed temporal evolution, we must acknowledge the large variability between the nights. The largest variation is observed just after sunset (18:00–22:00 LT) and before sunrise (04:00–06:00 LT). After sunset, the large vertical temperature variation, driven primarily by temperature changes within two-hour windows (black error bars), can be attributed to strong longwave radiative cooling of the canopy at the onset of the night (Stull, 1988). In contrast, the large variation in the CO₂ molar fraction at 4 m is mainly driven by inter-night variability (red error bars). This pronounced nocturnal CO₂ variability may result from three mechanisms: (1) the strength of the stable stratification, which is influenced by radiative cooling and the extent of CO₂ transport from the stable canopy layer into the atmosphere aloft (Moonen et al., 2025), (2) the absolute temperature and moisture conditions, influencing the respiration rate (Lloyd and Taylor, 1994; Zanchi et al., 2012) and amount of CO₂ stored in the stable daytime layer and (3) the preference of CO₂ to build up in low topographical points (de Araújo et al., 2010), where low-level horizontal advection can cause large differences in CO₂ molar fraction at the surface. Because of the non-homogeneous influence of the mechanisms over time, it remains hard to tell the dominant mechanism, this might well be variable between nights.

Just before sunrise, we observe large variation in CO₂ molar fractions throughout the vertical profile. The large variation can be explained by enhanced mixing during sunrise. In a previous study by Dupont et al. (2024), they found that the transport of CO₂ in the early morning differs between low and high wind speeds. The variable wind conditions during the CloudRoots-Amazon22 period (Appendix B), may explain the large contrast between nights.

$Ri_{\text{atm}} < Ri_c$, leading to $Ri_{\text{can}} > Ri_c$ by the end of the night. Applying this parameter to the profiles in Fig. 5e confirms that the strongly stable profile is decoupled, whereas the weakly stable profile remains coupled, permitting interaction between the two layers. This distinction between regimes controls the accumulation of CO₂ below the canopy and, consequently, the magnitude of the flux during sudden increases in turbulence (Acevedo et al., 2016).

3.3 Exchange of CO₂ and heat at the canopy–atmosphere interface

We evaluated the governing equations of TKE (Sect. 3.3.1), heat (Sect. 3.3.2) and CO₂ (Sect. 3.3.3) above the canopy, for the periods classified as strongly and weakly stable (Table 3). Vertical profiles for all six criteria differ distinctly between the two regimes (Fig. 5). The strongly stable vertical θ profile shows a larger temperature gradient between 25–50 m (2.08 K) than the weakly stable profile (1.29 K) (Fig. 5a). For CO₂ (Fig. 5b), a similar pattern is observed: the CO₂ gradient between 38–79 m is larger for the strongly stable regime (−21.8 ppm) than during the weakly stable regime (−5.5 ppm). These differences are attributed to a combination of the nocturnal temporal evolution of θ and CO₂, as well as the influence of variable turbulent activity during the regimes.

Horizontal wind speed clearly differs between the strongly and weakly stable regimes (Fig. 5c). Already at 25 m, the wind speed and gradient with height in the weakly stable profile are enhanced compared to the strongly stable profile. At 50 m, the threshold of 2.5 m s^{-1} is reached in the weakly stable profile, but not for the strongly stable profile. Additionally, in the weakly stable regime, TKE (Fig. 5d) and u_* (Fig. 5f) at 50 m exceed the thresholds (Table 2) and $Ri < Ri_c$ (Fig. 5e), in contrast to the strongly stable regime. The observed differences vanish above 200 m for TKE, Ri , and u_* , suggesting that the turbulent regime does not significantly influence the residual layer. The observed differences between the regimes are consistent with a previous study by Dias-Júnior et al. (2017).

3.3.1 Observational interpretation of the TKE budget

We evaluated the terms of the TKE budget (Eq. 3) above the canopy one by one, distinguishing between the strongly and weakly stable regimes (Fig. 6). In Fig. 6, the average logarithmic fitted profiles (least-squares polynomial) are plotted for the four strongly stable (S1, S2, S3, and S4) and four weakly stable (W1, W2, W3, and W4) profiles indicated in Table 3. The shaded area indicates the standard deviation between these four selected hours for both the strongly and weakly stable regimes.

For both the strongly and weakly stable regimes, the TKE tendency (Fig. 6a) is 10–100× smaller compared to the individual terms and shows no significant differ-

ence between the regimes. The buoyancy term (Fig. 6b) is most negative at the canopy top and increases with height. For the weakly stable regime, buoyancy is more negative ($-1.70 \times 10^{-3} \pm 0.94 \times 10^{-3} \text{ m}^2 \text{ s}^{-3}$) than for the strongly stable regime ($-0.30 \times 10^{-3} \pm 0.31 \times 10^{-3} \text{ m}^2 \text{ s}^{-3}$), indicating, that there is almost no vertical turbulent movement in the strongly stable regime. The shear term at the canopy top (Fig. 6c), is significantly larger for the weakly stable regime ($6.20 \times 10^{-3} \pm 2.05 \times 10^{-3} \text{ m}^2 \text{ s}^{-3}$) than for the strongly stable regime ($0.16 \times 10^{-3} \pm 0.52 \times 10^{-3} \text{ m}^2 \text{ s}^{-3}$). This is in accordance with Mironov and Sullivan (2016), who found that the magnitude of shear is about three times larger than buoyancy in the weakly stable regime.

The dissipation rate is highest at the canopy top and decreases logarithmically with height. In the weakly stable regime the dissipation is larger ($4.69 \times 10^{-3} \pm 0.25 \times 10^{-3} \text{ m}^2 \text{ s}^{-3}$) than for the strongly stable regime ($0.43 \times 10^{-3} \pm 0.44 \times 10^{-3} \text{ m}^2 \text{ s}^{-3}$). Our results for dissipation are in also in line with Mironov and Sullivan (2016), which find dissipation to have a magnitude of shear minus buoyancy.

The residual (Fig. 6e), which approaches zero with height, represents measurement and calculation errors, vertical and horizontal advection of TKE and the effect of pressure fluctuations. Previous studies have indicated that vertical and horizontal advection and pressure fluctuations can be important for the TKE balance above dense canopies (Dwyer et al., 1997; Vickers and Thomas, 2013; Chen and Chamecki, 2023). However, our results suggest that these terms are small relative to the buoyancy, shear, and dissipation terms under weakly stable conditions. Under strongly stable conditions, they may make an important contribution to the residual, but it remains hard to quantify the magnitude of the terms included in the residual. Further research incorporating high-frequency observations and large-eddy simulations with multilayer canopy representations could provide deeper insight into the magnitude of these terms in the residual under both stability regimes.

3.3.2 Observation interpretation of the heat budget

Secondly, we evaluated the governing equation of heat, which quantifies the heating of the atmosphere as a function of the turbulent transport of heat (driven by shear), the atmospheric longwave cooling rate and other processes represented by the residual (Fig. 7). The temperature tendency is negative for both the strongly ($-0.29 \pm 1.06 \text{ K h}^{-1}$ at the canopy top) and weakly ($-0.37 \pm 0.49 \text{ K h}^{-1}$ at the canopy top) stable regime (Fig. 7a). However, especially in the strongly stable regime, the temperature tendency varies strongly, caused by the variation between S1, S2, S3, and S4. Above the canopy, at 50 m, the temperature decreased during S2 (-1.7 K h^{-1}) and S4 (-0.03 K h^{-1}) but increased during S1 (0.3 K h^{-1}) and S3 (0.3 K h^{-1}). Two processes can cause the increase in temperature: (1) non-turbulent sub-mesoscale effects in strongly stable periods (Oliveira et al., 2018), and

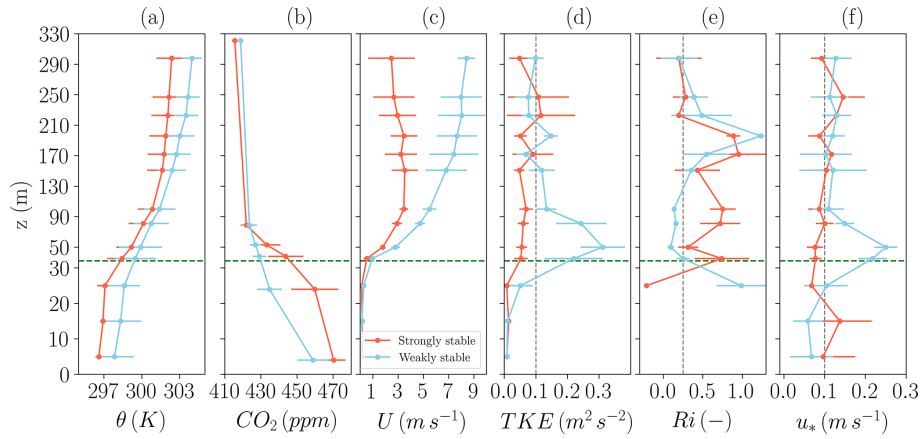


Figure 5. Vertical profiles above and below the canopy, the dashed green line indicates the aerodynamic top of the canopy (32 m). Profiles are shown for (a) potential temperature (θ), (b) CO₂, (c) wind speed (U), (d) turbulent kinetic energy (TKE), (e) Richardson number (Ri) and (f) friction velocity (u_*). In the figure, the red line shows the average of the strongly stable cases (S1, S2, S3, and S4), with the standard deviation in error bars. The blue lines show the same, but for the weakly stable cases (W1, W2, W3, and W4). The vertical lines in (c), (d), (e), and (f) represent the threshold values (at 50 m) used to select the strongly and weakly stable profiles, being 2.5 m s^{-1} , $0.1 \text{ m}^2 \text{ s}^{-2}$, 0.25 and 0.1 m s^{-1} respectively.

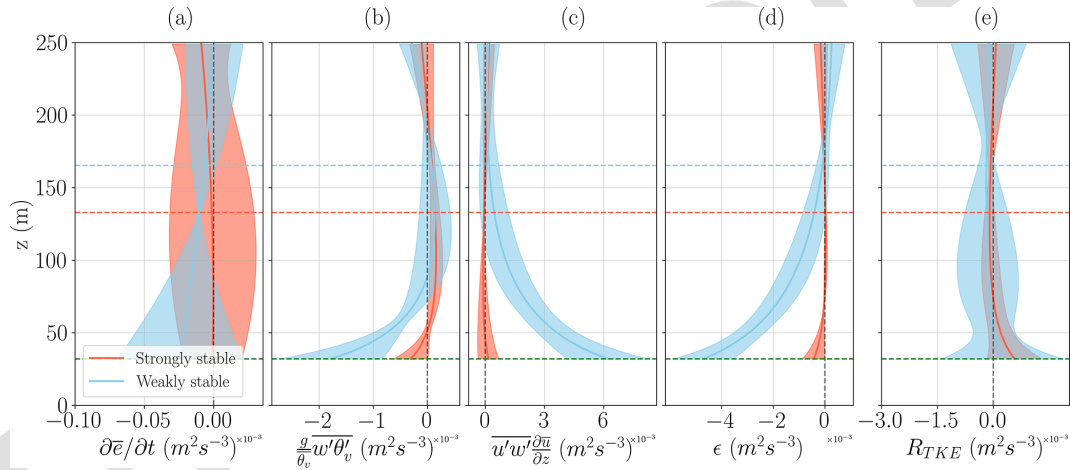


Figure 6. Conservation of turbulent kinetic energy (TKE), according to Eq. (3), based on the selected prototypes described in Table 3. Here, the red line represents the average of the four strongly stable profiles (S1, S2, S3, and S4), and the shaded area represents the standard deviation. Blue is the same, but for the four weakly stable profiles (W1, W2, W3, and W4). The graphs represent the terms of the TKE equation: (a) the TKE tendency, (b) the buoyant consumption, (c) the shear production, (d) dissipation, and (e) the residual. The horizontal lines indicate the average h_{NBL} , calculated for the strongly (red) and weakly (blue) stable regimes.

(2) the presence of the NLLJ during S3 and S4 which mixes air from above downward, increasing the temperature (Smedman, 1988; Mahrt, 1998; Karipot et al., 2006).

The second term, turbulent transport of heat, decreases logarithmically with height, and approaches zero above h_{NBL} (Fig. 7b). The weakly stable profile showed a more negative turbulent heat-flux divergence at the canopy top ($-3.62 \pm 1.54 \text{ K h}^{-1}$), than the strongly stable profile ($-1.23 \pm 0.79 \text{ K h}^{-1}$). This difference is consistent with the enhanced mechanical production of turbulent kinetic energy observed in the weakly stable regime (Fig. 6c), which promotes turbulent mixing between the cold canopy layer and the warmer

atmosphere aloft. Compared to a grassland surface, turbulent transport of heat above the Amazon canopy is up to 2 K h^{-1} higher (Stull, 1988). The difference between the two surface types can be explained by the higher surface roughness of the forest canopy, which enhances turbulent mixing.

The atmospheric longwave cooling rate decreases logarithmically with height above the canopy top (Fig. 7c). The strongly stable regime ($-0.35 \pm 0.02 \text{ K h}^{-1}$ at the canopy top) shows a higher atmospheric longwave cooling rate than the weakly stable regime ($-0.24 \pm 0.35 \text{ K h}^{-1}$ at the canopy top). Cloud cover likely reduced the longwave cooling rate during W1 and W2 (Appendix D), partly explaining the difference