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**INSTITUTO NACIONAL DE PESQUISAS DA AMAZÔNIA – INPA
UNIVERSIDADE DO ESTADO DO AMAZONAS - UEA
PROGRAMA DE PÓS-GRADUAÇÃO EM CLIMA E AMBIENTE**

**Regimes de turbulência noturna e a relação com a
altura da camada limite noturna, jatos de baixos
níveis e mortalidade de árvores na Amazônia.**

ANNE CRISTINY SANTOS DE MENDONÇA

**Manaus, Amazonas
Fevereiro, 2026**

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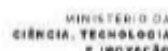
Orientador: Dr. Cléo Quaresma Dias Júnior

Coorientador: Dr. Daniel Magnabosco Marra

Tese apresentada ao Instituto Nacional de Pesquisas da Amazônia e Universidade do Estado do Amazonas, como parte dos requisitos para obtenção do título de Doutora em Clima e Ambiente.

Manaus, Amazonas

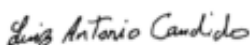
Fevereiro, 2026



**ATA DE DEFESA PÚBLICA DE TESE DE DOUTORADO DO
PROGRAMA DE PÓS-GRADUAÇÃO EM CLIMA E AMBIENTE
DO INSTITUTO NACIONAL DE PESQUISAS DA AMAZÔNIA E
UNIVERSIDADE DO ESTADO DO AMAZONAS.**

Aos 23 dias do mês de fevereiro do ano de 2026, às 13:30 horas (Manaus), por videoconferência, reuniu-se a Comissão Julgadora de Defesa de Tese, composta pelos seguintes membros: Prof. Dr. Cleo Quaresma Dias Júnior (Presidente), Profa. Dra. Rita Valeria Andreoli de Souza (UEA), Prof. Dr. Luiz Antonio Candido (INPA), Profa. Dra. Livia Souza Freire Grion (USP) e a Profa. Dra. Rachel Ifanger Albrecht (USP), tendo como suplente a Profa. Dra. Julia Clarinda Paiva Cohen (UFPA), sob a presidência do primeiro, a fim de proceder a arguição do trabalho de Tese de **Anne Cristiny Santos de Mendonça**, intitulado “Regimes de turbulência noturna e a relação com a altura da camada limite noturna, jatos de baixos níveis e mortalidade de árvores na Amazônia”, orientada pelo Prof. Dr. Cleo Quaresma Dias Júnior (INPA) e coorientada pelo Prof. Dr. Daniel Magnabosco Marra.

O Presidente da Comissão Julgadora deu início à seção e informou os procedimentos do exame. O aluno fez a exposição do seu trabalho que foi avaliado pelos membros da Comissão Julgadora de Defesa de Tese. A Comissão Julgadora, então, se reuniu e os membros emitiram os seguintes pareceres:

Nome	Parecer	Assinatura
Dr. Cleo Quaresma Dias Júnior	(X) Aprovada () Reprovada	
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Nada mais havendo a relatar, foi lavrada a presente ata que, após lida e aprovada, foi assinada pelos membros da Comissão Julgadora.

(*) Membros que participaram por vídeo conferência, com parecer reconhecido pelo presidente da banca.

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Sinopse:

Estudou-se a organização da turbulência noturna acima do dossel florestal na Amazônia central e sua relação com a variabilidade da altura da camada limite noturna. Avaliou-se o papel dos jatos de baixos níveis e de *downdrafts* na dinâmica da turbulência, bem como a ocorrência de ventos extremos e sua relação com eventos de mortalidade de árvores.

Palavras-chave: Floresta tropical, camada limite noturna, turbulência atmosférica, jatos de baixos níveis, ventos extremos, distúrbio florestal.

Dedico este trabalho a minha família que
muito me apoiou e me incentivou a realizá-lo.

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Muito obrigada!

*"An equation means nothing to me
unless it expresses a thought of God."
(Srinivasa Ramanujan)*

RESUMO

A turbulência desempenha um papel fundamental nas trocas de energia e massa entre a floresta e a atmosfera. Entretanto, durante a noite, seu enfraquecimento e o aumento da estabilidade atmosférica podem promover o desacoplamento entre o dossel e a atmosfera, tornando a estrutura da Camada Limite Noturna (CLN) particularmente complexa. Na Amazônia, ainda é limitada a compreensão de como a organização da turbulência determina a altura da CLN e de que forma fenômenos de maior escala, como Jatos de Baixos Níveis (JBNS) e sistemas convectivos, interferem nessa dinâmica, em parte devido à escassez de medições turbulentas com alta resolução vertical. Nesta tese, investigou-se a organização da turbulência noturna acima do dossel florestal na Amazônia Central, sua relação com a variabilidade da altura da CLN e com os JBNS, bem como a influência de *downdrafts* e ventos extremos associados à mortalidade de árvores. Para isso, foram utilizadas medições de alta frequência realizadas em diferentes alturas acima do dossel, incluindo aquelas obtidas ao longo da torre de 325 m do Observatório da Torre Alta da Amazônia (ATTO), cuja configuração observacional possibilitou investigar diretamente a estrutura vertical da turbulência ao longo da CLN, até então predominantemente estudada por meio de abordagens indiretas. Os resultados mostraram a existência de dois regimes turbulentos distintos, um muito estável e outro fracamente estável. A transição entre esses regimes foi modulada pela proximidade do dossel, pela sazonalidade e pela ocorrência de *downdrafts*, que promoveram o acoplamento entre o interior da floresta e a atmosfera. A altura da CLN (h_N), obtida a partir de perfis turbulentos, variou com a estabilidade atmosférica e rugosidade da superfície. Além disso, h_N foi menor que estimativas previamente reportadas com base na inversão térmica, indicando que métodos indiretos podem superestimar h_N . Em noites com JBNS, verificou-se que a altura do nariz do jato não representa universalmente h_N . Sob condições fracamente estáveis, a altura do nariz do jato foi consistente com h_N , enquanto, em condições muito estáveis, apresentou valores superiores, mostrando que a utilização dos JBNS como indicadores de h_N depende do estado de acoplamento entre o dossel e a atmosfera. A influência de *downdrafts* associados a sistemas convectivos foi também investigada com dados de reanálise ERA5. A análise de um evento associado a um *blowdown* previamente documentado mostrou que a reanálise reproduziu os principais padrões térmicos e dinâmicos associados à passagem da tempestade, embora sua resolução tenha limitado a representação das magnitudes locais das rajadas. A análise dos ventos extremos revelou intensificação entre 2005 e 2020, especialmente na estação chuvosa. Em conjunto, os resultados mostram que a estrutura da atmosfera noturna sobre a floresta Amazônica resulta da interação entre processos turbulentos locais e forçantes atmosféricas de maior escala. Esses resultados ampliam a compreensão dos mecanismos físicos que controlam o acoplamento floresta-atmosfera e estabelecem uma conexão entre a dinâmica da camada limite e os processos atmosféricos associados à mortalidade de árvores na Amazônia.

Palavras Chaves: *Floresta Amazônica; Camada Limite Noturna; Regimes de Turbulência; Jatos de Baixos Níveis; Ventos extremos; Mortalidade de Árvores.*

ABSTRACT

Turbulence plays a fundamental role in energy and mass exchanges between the forest and the atmosphere. However, during nighttime, reduced turbulence and increased atmospheric stability can lead to decoupling between the canopy and the atmosphere, making the structure of the Nocturnal Boundary Layer (NBL) particularly complex. In the Amazon, our understanding of how turbulence organization determines NBL height and how larger-scale phenomena, such as Low-Level Jets (LLJs) and convective systems, influence these dynamics remains limited, in part due to the scarcity of turbulence measurements with high vertical resolution. In this work, we investigated the organization of nocturnal turbulence above the forest canopy in Central Amazonia, its relationship with NBL height variability and LLJs, as well as the influence of downdrafts and extreme winds associated with tree mortality. We used high-frequency measurements collected at different heights above the canopy, including those obtained along the 325-m tower at the Amazon Tall Tower Observatory (ATTO), this unique observational setup allowed us to directly investigate the vertical structure of turbulence throughout the NBL, which had previously been studied using indirect approaches. The results revealed two distinct turbulence regimes, a very stable regime and a weakly stable regime. Transitions between these regimes were modulated by proximity to the canopy, seasonality, and the occurrence of downdrafts, which promoted coupling between the forest and the atmosphere. NBL height (h_N), determined from vertical turbulence profiles, varied with atmospheric stability and surface roughness. In addition, h_N was lower than previous estimates based on thermal inversion, indicating that indirect methods may overestimate h_N . During nights with LLJs, the jet nose height did not universally represent h_N . Under weakly stable conditions, it was consistent with h_N , whereas under very stable conditions it was located above h_N , showing that the use of LLJs as indicators of h_N depends on the coupling state between the canopy and the atmosphere. The influence of downdrafts associated with convective systems was also investigated using ERA5 reanalysis data. The analysis of an event associated with a previously documented blowdown showed that the reanalysis reproduced the main thermal and dynamical patterns associated with the passage of the storm, although its resolution limited the representation of local gust magnitudes. The analysis revealed an intensification of extreme winds between 2005 and 2020, particularly during the wet season. Overall, the results show that the structure of the nocturnal atmosphere over the Amazon forest results from the interaction between local turbulent processes and larger-scale atmospheric forcing. These findings advance the understanding of the physical mechanisms controlling forest–atmosphere coupling and establish a connection between boundary-layer dynamics and atmospheric processes associated with tree mortality in the Amazon.

Keywords: Amazon Forest; Nocturnal Boundary Layer; Turbulence Regimes; Low-Level Jets; Extreme Winds; Tree Mortality.

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LISTA DE ABREVIATURAS E SIGLAS

ATTO: *Amazon Tall Tower Observatory* / Observatório da Torre Alta da Amazônia.

NBL (CLN): *Nocturnal boundary layer* (Camada Limite Noturna).

CWS: *Critical wind-speed* / Velocidade Crítica do Vento.

ECMWF: *European Centre for Medium-Range Weather Forecasts*.

EEST / ZF2: Estação Experimental de Silvicultura Tropical / ZF2.

ERA5: *ECMWF Reanalysis v5* / Reanálise atmosférica de quinta geração do ECMWF.

GOES-13: *Geostationary Operational Environmental Satellite-13*.

GoAmazon: *Green Ocean Amazon Experiment*.

GPM: *Global Precipitation Measurement*.

IMERG: *Integrated Multi-satellite Retrievals for GPM*.

INPA: Instituto Nacional de Pesquisas da Amazônia.

ITCZ (ZCIT): *Intertropical Convergence Zone* (Zona de Convergência Intertropical).

LLJ (JBN): *Low-Level Jet* (Jato de Baixos Níveis).

NN: *Near-neutral regime* (Regime Quase-Neutro).

RSL: *Roughness Sublayer* / Subcamada Rugosa.

SRTM: *Shuttle Radar Topography Mission*.

TKE: *Turbulent Kinetic Energy* / Energia Cinética Turbulenta.

VS: *Very Stable regime* / Regime muito estável.

WD: *Wind Direction* / Direção do vento.

WD_{nose}: *Wind direction at the jet nose* / Direção do vento na altura do nariz do jato.

WS: *Weakly Stable regime* / Regime fracamente estável.

LISTA DE SÍMBOLOS

C_p : Calor específico à pressão constante do ar.

d : *zero-plane displacement height* / Altura de deslocamento do plano zero.

$d\theta/dz$: Gradiente vertical da temperatura potencial.

dV/dz : Cisalhamento vertical da velocidade horizontal do vento.

g : *gravitational acceleration* / Aceleração da gravidade.

H : *Sensible heat flux* / Fluxo de calor sensível.

h_N : *Nocturnal boundary layer height* / Altura da camada limite noturna.

h_{N_flux} : Altura da CLN estimada a partir do perfil de fluxo turbulento.

h_{N_jet} : Altura da CLN estimada a partir da altura do nariz do JBN.

h_{nose} : *Jet nose height* / Altura do nariz do jato.

$h_{maxstab}$: Altura do máximo gradiente vertical de temperatura potencial.

L : Comprimento de *Monin-Obukhov*.

O_3 : Concentração de ozônio.

R_N : *Net radiation* / Saldo de radiação.

R_i : *Richardson number* / Número de Richardson local.

R_{i_bulk} : *bulk Richardson number* / Número de Richardson global.

T : Temperatura do ar.

T_2 : Temperatura do ar a 2 metros.

T_h : Temperatura medida pelo termo-higrômetro.

u : Componente zonal do vento.

u' : Flutuação da componente zonal do vento.

u_* : *Friction velocity* / Velocidade de fricção.

$\overline{w'u'}$: *Kinematic momentum flux* / Fluxo cinemático de momento.

v : Componente meridional do vento.

v' : Flutuação da componente meridional do vento.

V : Velocidade horizontal do vento.

V_{10} : Velocidade horizontal do vento a 10 metros.

V_{100} : Velocidade horizontal do vento a 100 metros.

V_L : *Wind speed threshold* / Velocidade limiar do vento.

V_T : *Wind speed crossover threshold* / Limiar de cruzamento da velocidade do vento.

V_{TKE} : Escala de velocidade turbulenta.

w : Componente vertical da velocidade do vento.

w' : Flutuação da componente vertical do vento.

$\overline{w'T'}$: *Kinematic sensible heat flux* / Fluxo cinemático de calor sensível.

z_0 : *roughness length* / Comprimento de rugosidade aerodinâmica.

Z_e : *Equivalent radar reflectivity factor* / Fator de refletividade equivalente do radar.

κ : Constante de von Kármán.

ρ : *Air density* / Densidade do ar.

σ : *Standard deviation* / Desvio padrão.

θ : *Potential temperature* / Temperatura potencial.

INTRODUÇÃO GERAL

A capacidade da floresta Amazônica de trocar calor, vapor d'água, dióxido de carbono e uma variedade de gases e partículas com a atmosfera depende fundamentalmente da turbulência. Essas trocas influenciam a circulação atmosférica, o ciclo hidrológico e o balanço de energia na superfície (Malhi et al., 2002; Von Randow et al., 2004; Thomas e Foken, 2007; Knohl e Baldocchi, 2008; Alves et al., 2016; Gerken et al., 2016). Durante a noite, entretanto, o enfraquecimento da turbulência torna essa interação complexa, podendo reduzir a eficiência das trocas e promover o desacoplamento entre o dossel e a atmosfera acima da floresta (Acevedo et al., 2009; Thomas et al., 2013; Cava et al., 2022). Compreender como e até que altura a turbulência mantém essa conexão é, portanto, essencial para explicar como a floresta interage com a atmosfera e como essa interação é modulada por diferentes condições atmosféricas, incluindo eventos extremos.

As trocas turbulentas entre a superfície e a atmosfera governam a estrutura e a evolução temporal da Camada Limite Atmosférica (CLA), que responde às forçantes da superfície em escalas de tempo da ordem de uma hora ou menos (Stull, 1988). Por corresponder ao volume de ar diretamente influenciado pela superfície, a altura da CLA determina a extensão vertical dessa influência, sendo um parâmetro fundamental para compreender a dispersão de gases, partículas e poluentes e para aplicações em previsão do tempo, modelagem climática, qualidade do ar e química atmosférica (Kotthaus e Grimmond, 2018; Levi et al., 2020). O resfriamento radiativo da superfície favorece a formação da Camada Limite Noturna (CLN), caracterizada por turbulência fraca ou intermitente e sensível à atuação de diferentes forçantes atmosféricas. Sobre a floresta Amazônica, essa complexidade é ampliada pela presença do dossel, cuja elevada rugosidade modifica o escoamento atmosférico e dá origem a uma subcamada rugosa (*roughness sublayer*) (Raupach et al., 1996; Finnigan, 2000). Assim, a caracterização das trocas turbulentas sobre florestas tropicais constitui um desafio maior do que sobre superfícies homogêneas.

Nas últimas décadas, avanços importantes na compreensão da turbulência atmosférica noturna revelaram a existência de diferentes regimes turbulentos. Em latitudes médias, Mahrt (1998) evidenciou que o comportamento dos fluxos turbulentos em relação ao parâmetro de estabilidade de *Monin-Obukhov* permite distinguir dois regimes: “fracamente estável” e “muito estável”. Posteriormente, Sun

et al. (2012) identificaram um valor limiar de velocidade do vento que caracteriza a mudança entre esses dois regimes. Abaixo desse limiar, a turbulência é fraca e dominada pelo cisalhamento local (condição muito estável), enquanto acima dele a intensidade da turbulência aumenta com a velocidade do vento (condição fracamente estável). Aprofundando essa interpretação, Acevedo et al. (2016) identificaram um limiar crítico de vento que organiza a CLN em dois estados distintos: um estado acoplado, com turbulência contínua e conexão vertical eficiente entre as camadas da atmosfera, e outro desacoplado, caracterizado por turbulência fraca ou intermitente. Esses resultados demonstram que a turbulência não controla apenas a intensidade dos fluxos, mas também o acoplamento vertical entre as camadas da CLN.

Embora esses conceitos tenham sido inicialmente desenvolvidos a partir de experimentos conduzidos em latitudes médias, evidências recentes indicam que esses diferentes regimes turbulentos também ocorrem sobre florestas tropicais. Estudos realizados na Amazônia mostraram que os regimes muito estável e fracamente estável são caracterizados por padrões distintos de concentração de CO₂ acima do dossel e também por diferentes alturas do ponto de inflexão do perfil de velocidade do vento (Mafra et al., 2016; Dias-Júnior et al., 2017b), indicando que esses regimes influenciam as trocas de massa e energia na interface floresta-atmosfera. Adicionalmente, foi evidenciado que a mudança de um regime para outro pode estar associada à passagem de sistemas convectivos de mesoescala, como as linhas de instabilidade (Bezerra et al., 2021). Dessa forma, a organização da turbulência em diferentes regimes resulta da interação entre processos locais de produção e dissipação da turbulência (microescala) e forçantes atmosféricas de maior escala (mesoescala).

Devido à altura da CLA corresponder à máxima extensão vertical da mistura turbulenta que mantém a atmosfera diretamente acoplada à superfície (Lenschow et al., 1988; Stull, 1988), compreender como a turbulência se organiza é fundamental para explicar os mecanismos físicos que controlam a variabilidade da sua altura. Entretanto, ao contrário da camada limite diurna, cuja intensa mistura turbulenta resulta em um topo relativamente bem definido (Hennemuth e Lammert, 2006), na CLN a turbulência fraca e intermitente, associada à forte estabilidade atmosférica e à atuação simultânea de processos em diferentes escalas, resulta em uma estrutura vertical menos definida, dificultando a identificação de seu topo.

Como medições de variáveis turbulentas com alta resolução vertical são, em geral, difíceis de obter, diferentes parâmetros têm sido amplamente empregados para estimar a altura da CLN. Entre eles, destaca-se a análise de perfis médios de temperatura, umidade, velocidade do vento e concentração de aerossóis, obtidos por radiossondas, balões cativos e instrumentos de sensoriamento remoto de superfície, como ceilômetros, sodar e radiômetros (Garstang et al., 1990; Acevedo et al., 2004; Eresmaa et al., 2006; Carneiro e Fisch, 2020; Dias-Júnior et al., 2022). Cada uma dessas abordagens apresenta vantagens e limitações, mas todas são consideradas metodologias indiretas, uma vez que utilizam variáveis que atuam como aproximações (proxy) da intensidade da turbulência (Mahrt et al., 1979; Vickers e Mahrt, 2004).

Entre os parâmetros utilizados como aproximação da altura da CLN, destaca-se a altura do Jato de Baixos Níveis (JBN), cuja utilização baseia-se na influência que esse fenômeno exerce sobre a estrutura da turbulência. Os JBNs são caracterizados pela ocorrência de máximos de velocidade do vento (nariz do jato) próximos à superfície, localizados entre algumas dezenas e poucas centenas de metros de altura. Eles desempenham um papel importante na dinâmica da CLN ao modificar o transporte de umidade, CO₂, ozônio e poluentes atmosféricos (Banta, 2008; Karipot et al., 2008; Corrêa et al., 2021). Além disso, intensificam o cisalhamento vertical do vento abaixo do nariz do jato, constituindo uma importante fonte de produção mecânica de turbulência, capaz de influenciar as trocas entre a superfície e a atmosfera (Prabha et al., 2008; Santana et al., 2015; Dias-Júnior et al., 2017a). Assim, embora os JBNs sejam considerados um fenômeno de mesoescala, seus efeitos influenciam diretamente a turbulência local. Por essa razão, diversos estudos passaram a utilizar a altura do nariz do JBN como uma aproximação da altura da CLN, assumindo que o máximo de velocidade do vento coincide com o limite superior da mistura turbulenta.

A ocorrência de JBNs tem sido amplamente documentada em diversas regiões do mundo, sendo atribuída a diferentes mecanismos que podem atuar de forma isolada ou combinada (Blackadar, 1957; Shapiro et al., 2022). Na Amazônia, a formação do JBN tem sido associada à estratificação térmica gerada pelo resfriamento radiativo do dossel florestal. Esse processo favorece o desacoplamento turbulento, reduzindo o arrasto acima da inversão térmica (Greco et al., 1992). Estudos posteriores mostraram que os JBNs também podem ser modulados por

circulações de brisa (floresta-rio), correntes de densidade e sistemas convectivos (de Oliveira e Fitzjarrald, 1993, 1994; Dias-Júnior et al., 2017a; Anselmo et al., 2020). Apesar dessas evidências, a hipótese de que a altura do nariz do jato representa o topo da CLN ainda carece de validação baseada em medições diretas de turbulência com alta resolução vertical.

A infraestrutura do *Amazon Tall Tower Observatory* (ATTO) constitui uma oportunidade singular para investigar a dinâmica atmosférica na Amazônia, ao disponibilizar medições turbulentas em diversos níveis acima do dossel. Com uma torre de 325 m instrumentada com sensores de alta frequência, o ATTO permite o monitoramento ao longo de praticamente toda a extensão da CLN, superando as limitações observacionais que historicamente restringiram o estudo desses processos em florestas tropicais (Andreae et al., 2015; Dias-Júnior et al., 2019). Ao final de 2021, foi concluída a instalação de 11 níveis de medição acima do dossel, uma configuração extremamente rara em estudos da CLN. Essa configuração torna possível analisar diretamente a organização vertical da turbulência, sua relação com a profundidade da CLN e a influência de fenômenos como os JBNs.

Além dos JBNs, sistemas convectivos de mesoescala constituem outro importante mecanismo capaz de modificar a organização da turbulência. Na Amazônia, a elevada disponibilidade de vapor d'água e o intenso aquecimento da superfície favorecem o desenvolvimento frequente desses sistemas (Betts et al., 2002, 2003; Machado et al., 2002, 2004). Durante sua evolução, correntes descendentes de ar (*downdrafts*) transportam em direção à superfície ar resfriado pela evaporação da precipitação (Fujita, 1985; Nelson, 1994; Garstang et al., 1998; Negrón-Juárez et al., 2010, 2017, 2018). Ao atingir a superfície, esse ar mais frio e denso forma uma piscina fria (*cold pool*), cuja propagação horizontal dá origem às frentes de rajada (*gust fronts*). Ao longo desta tese, o termo *downdrafts* será utilizado para representar esse conjunto de processos associados a um mesmo evento convectivo. Esses fenômenos modificam os fluxos turbulentos, promovem aumentos abruptos da velocidade do vento e alterações simultâneas na temperatura e na pressão, capazes de reorganizar a dinâmica da camada limite em escalas de poucos minutos (Garstang e Fitzjarrald, 1999; Dias-Júnior et al., 2017a; Oliveira et al., 2020). Essas alterações podem inclusive promover transições rápidas entre os diferentes regimes turbulentos (Bezerra et al., 2021), evidenciando como processos convectivos interferem na organização da turbulência observada acima do dossel.

Além de modificar as propriedades termodinâmicas e químicas da massa de ar próxima à superfície, as rajadas de vento constituem um dos principais agentes naturais de distúrbio da floresta Amazônica. Quando suficientemente intensas, essas rajadas causam a quebra de troncos ou o desenraizamento de árvores, resultando em eventos de mortalidade de árvores em larga escala, conhecidos como *blowdowns* (Urquiza Muñoz et al., 2021, 2024). A frequência e severidade desses distúrbios e o tempo de recuperação da floresta influenciam diretamente o estoque e o balanço de carbono, bem como a composição funcional dos ecossistemas (Marra et al., 2014, 2018; Negrón-Juárez et al., 2018). Além disso, a decomposição da biomassa morta promove emissões de CO₂ (Chambers et al., 2004) e, potencialmente, de metano (CH₄), contribuindo para o aumento da concentração de gases de efeito estufa na atmosfera. Assim, tempestades convectivas desempenham papel central tanto na modulação da turbulência quanto na ocorrência de distúrbios florestais de grande escala, estabelecendo uma conexão entre processos atmosféricos e ecológicos na Amazônia.

Dessa forma, a compreensão da atmosfera noturna sobre a Amazônia requer uma abordagem capaz de considerar processos que atuam desde a escala da turbulência até os fenômenos convectivos associados à mortalidade de árvores. Embora avanços importantes tenham sido obtidos na caracterização desses mecanismos, ainda é limitado o conhecimento sobre como diferentes processos físicos contribuem para a organização da CLN, para as trocas entre a floresta e a atmosfera e para a ocorrência de ventos extremos na Amazônia.

OBJETIVO GERAL

Identificar e avaliar os regimes de turbulência acima do dossel florestal, explorando sua influência sobre variabilidade da altura da camada limite noturna e sua relação com jatos de baixos níveis, bem como sua associação com *downdrafts* e suas implicações para a mortalidade de árvores na Amazônia.

Objetivos Específicos:

- 1) Identificar e caracterizar os regimes de turbulência e sua variabilidade sazonal, e analisar a relação dos *downdrafts* com esses regimes e com o grau de acoplamento entre o dossel florestal e a atmosfera no período noturno.
- 2) Estimar a variabilidade da altura da camada limite noturna sob diferentes regimes de turbulência e avaliar o efeito da rugosidade da superfície nessas estimativas.
- 3) Avaliar sob quais condições a altura do jato de baixos níveis fornece uma boa aproximação para o topo da camada limite noturna.
- 4) Identificar padrões atmosféricos associados a *downdrafts* que causam *blowdowns* na Amazônia Central e examinar tendências de vento extremo usando ERA5.

JUSTIFICATIVA

A compreensão da turbulência noturna é um dos principais desafios da micrometeorologia em razão da complexidade dos processos físicos que atuam em diferentes escalas espaciais e temporais. Embora avanços importantes tenham sido obtidos, permanece limitada a compreensão de como diferentes processos físicos contribuem para a organização da atmosfera noturna sobre a floresta Amazônica. Essa limitação dificulta uma interpretação física mais consistente da estrutura da atmosfera acima da floresta e de sua resposta à atuação de fenômenos atmosféricos de maior escala.

Parte dessa limitação decorre das dificuldades observacionais inerentes ao estudo da atmosfera noturna. Grande parte do conhecimento disponível sobre a estrutura da CLN foi construída a partir de perfis atmosféricos obtidos em horários

específicos ou por metodologias que utilizam variáveis proxy da turbulência. Embora essas abordagens tenham produzido avanços importantes, elas oferecem uma representação pontual ou indireta da camada limite, limitando a investigação de sua variabilidade e de sua relação com os diferentes estados turbulentos.

A realização desta pesquisa foi viabilizada pelos avanços recentes da infraestrutura observacional disponível na Amazônia Central. Em especial, os dados coletados no sítio ATTO, uma infraestrutura científica de grande porte resultante da cooperação científica entre o Instituto Nacional de Pesquisas da Amazônia (INPA), no Brasil, e a Sociedade Max Planck, na Alemanha. A disponibilidade de medições turbulentas distribuídas ao longo de uma torre de 325 m tornou possível investigar diretamente processos que, até recentemente, eram analisados predominantemente por meio de aproximações indiretas. Essa capacidade observacional representa um avanço significativo em relação às abordagens tradicionalmente empregadas.

Neste contexto, esta tese busca uma investigação integrada dos diferentes aspectos da atmosfera noturna amazônica, permitindo construir uma interpretação física mais abrangente de sua organização. Essa abordagem conecta os regimes turbulentos observados acima do dossel às variações da altura da CLN e aos JBNs e permite ainda investigar a influência de sistemas convectivos sobre a dinâmica da camada limite e sua relação com a ocorrência de ventos extremos associados a *blowdowns*. O conhecimento produzido poderá contribuir para uma representação mais realista do transporte turbulento e da camada limite em modelos atmosféricos, para a interpretação das trocas floresta-atmosfera e para a compreensão da vulnerabilidade da floresta Amazônica frente a eventos atmosféricos extremos.

ORGANIZAÇÃO DA TESE

A presente tese está estruturada em quatro capítulos na forma de artigos científicos, cada capítulo é apresentado de forma independente, contendo todos os materiais necessários, desde a introdução até as referências bibliográficas. Os Capítulos 1, 2 e 3 foram publicados na revista *Agricultural and Forest Meteorology*, uma das principais revistas internacionais da área.

O Capítulo 1, intitulado "*Turbulence regimes in the nocturnal roughness sublayer: Interaction with deep convection and tree mortality in the Amazon*" (Mendonça et al., 2023), estabelece a base conceitual da tese ao investigar a

organização dos regimes turbulentos acima do dossel florestal amazônico. O estudo utiliza dados de turbulência obtidos durante o experimento GOAmazon (*Green Ocean Amazon Experiment*), realizado em 2014, com medições em três níveis acima do dossel. Embora essa configuração observacional não permitisse avaliar diretamente toda a estrutura vertical da CLN, os resultados permitiram identificar e caracterizar diferentes regimes turbulentos, sua variabilidade sazonal e sua relação com o acoplamento entre a floresta e a atmosfera. Além disso, o estudo demonstra que a ocorrência de *downdrafts* pode induzir transições rápidas entre esses regimes.

Partindo da caracterização dos regimes turbulentos, o Capítulo 2, intitulado "*Estimation of the nocturnal boundary layer height over the Central Amazon forest using turbulence measurements*" (Mendonça et al., 2025), examina como esses diferentes regimes influenciam a altura da CLN. Utilizando dados da torre ATTO, o capítulo apresenta uma estimativa direta dessa altura, permitindo avaliar sua variabilidade sob diferentes regimes turbulentos. A disponibilidade de medições turbulentas distribuídas ao longo de praticamente toda a CLN permitiu superar a limitação das abordagens tradicionalmente empregadas. Dessa forma, o estudo amplia a compreensão da estrutura vertical da atmosfera noturna e estabelece a ligação entre os regimes turbulentos e a profundidade da CLN.

O Capítulo 3, intitulado "*Is Low-Level Jet height a good approximation for the top of the nocturnal boundary-layer?*" (Mendonça et al., 2026), aprofunda essa discussão ao investigar a relação entre a altura da CLN e os JBNs. O capítulo explora em quais condições a altura do nariz do jato pode ser utilizada como uma aproximação para o topo da CLN. Os resultados demonstram que essa aproximação depende do regime turbulento predominante, oferecendo uma nova interpretação para a utilização dos JBNs como indicadores da estrutura da atmosfera noturna.

Por fim, o Capítulo 4 apresenta resultados preliminares sobre a atuação de *downdrafts* e sua relação com eventos de vento extremo responsáveis pela ocorrência de *blowdowns* na Amazônia Central. Nesse capítulo são identificados os principais padrões atmosféricos associados a esses eventos e analisada a variabilidade temporal dos ventos extremos utilizando dados da reanálise ERA5. Dessa forma, o estudo amplia a escala de investigação, conectando a organização da turbulência observada acima do dossel aos processos atmosféricos responsáveis pelos principais distúrbios naturais da floresta Amazônica.

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Turbulence regimes in the nocturnal roughness sublayer: Interaction with deep convection and tree mortality in the amazon

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Abstract

We investigated the influence of seasonality and proximity to the forest canopy on nocturnal turbulence regimes in the roughness sublayer of a Central Amazon forest. Since convective systems of different scales are common in this region, we also analyzed the effect of extreme wind gusts (propagated from convective downdrafts) on the organization of the turbulence regimes, and their potential to cause the mortality of canopy trees. Our data include high-frequency winds, temperature and ozone concentration at different heights during the dry and wet seasons of 2014. In addition, we used critical wind-speed data derived from a tree-winch experiment and a modeling study conducted in the same study site. Two different turbulence regimes were identified at three heights above the canopy: a weakly stable (WS) and a very stable regime (VS). The threshold wind speeds that mark the transition between turbulence regimes were larger during the dry season and increased as a function of the height above the canopy. The turbulent fluxes of sensible heat and momentum during the WS accounted for 88% of the entire nighttime flux. Downdrafts occurred only in the WS and favored a fully coupled state of wind flow along the canopy profile. The destructive potential of winds was four times higher than on nights without downdrafts.

Keywords: Downdrafts, extreme wind speed, seasonality, tropical forest, turbulence regimes, wind disturbance

1.1 Introduction

A growing number of observational and modeling studies show that the world's forests play a key role in global climate (Malhi et al., 2002; Bonan, 2008; Alves et al., 2016) and emphasize its importance to the water, radiation and surface-energy balance. All these are affected by exchanges of mass (water vapor, carbon dioxide, methane, ozone, and a variety of gases and particles), heat (sensible and latent) and momentum between the forest and the atmosphere (Von Randow et al., 2004; Thomas and Foken, 2007; Knohl and Baldocchi, 2008; Gerken et al., 2016, among others). This near-surface boundary layer that is dynamically influenced by vegetation is called the roughness sublayer (RSL). Particularly at night, the transport efficiency decreases above the forest, which inhibits its penetration into the canopy (Cava et al., 2022) and leads to a decoupling between the sub- and upper canopy flows (Thomas et al., 2013; Freundorfer et al., 2019; Cava et al., 2022).

Assessments of turbulent flow are challenging in the nocturnal RSL, particularly during the occurrence of intense downdrafts that typically originate within a convective system (e.g. squall lines, super-cell and single-cell thunderstorms). Intense downdrafts can develop when cooled air from evaporating precipitation sinks rapidly towards the ground. Such events transport relatively cooler and drier air from the troposphere to the lower levels leading to changes in the thermodynamic and kinematic properties of the near-surface air mass (Wakimoto, 2015). These downdrafts can extend to the surface in the form of sudden variations in the values of the wind, pressure, turbulent fluxes and a drop in equivalent potential temperature (Garstang and Fitzjarrald, 1999; Betts et al., 2002; Dias-Júnior et al., 2017a; Melo et al., 2019; D'Oliveira et al., 2022). Such phenomena can gain an interesting physical interpretation when the turbulence is classified into distinct turbulent nocturnal regimes. In mid-latitude regions for example, Mahrt (1998) categorized turbulence as “weakly stable” and “very stable” to differentiate the conditions of sustained and non-sustained turbulence, respectively. Sun et al. (2012) classified the turbulence regimes as either weak or strong, according to the relationship between turbulent kinetic energy (TKE) and wind speed. Subsequently, such terms have been widely used to refer to the regimes.

The weakly stable regime follows Monin-Obukhov similarity theory, which is used in most atmospheric models for expressing the surface layer. On the other hand, the same is not true in the very stable regime (Mahrt, 1998, 2014). In general,

numerical schemes such as those used in numerical weather prediction and atmospheric models reproduce the occurrence of both nocturnal turbulence regimes, but conditions that mark the transition between them depend on the parameterizations used to represent the eddy diffusivity (Costa et al., 2020). The transition between regimes is not universal, varying from one site to another. Such differences may be associated with local factors, for example by increased cloudiness (Acevedo and Fitzjarrald, 2003), surface roughness (Mahrt et al., 2013) and net nocturnal radiative loss (Sun et al., 2020; Acevedo et al., 2021).

Recent studies (Santos et al., 2016; Cava et al., 2022) have evidenced that the nature of the nocturnal-scalar exchange between the canopy and the overlying atmosphere is contrastingly dependent on the turbulence regime. While the weakly stable regime is dominated by turbulence, under very stable conditions a relatively larger fraction of the exchange is produced by non-turbulence motion with long time scales. This fact is particularly important for quantifying these fluxes since under very stable conditions the typical eddy-covariance methods might not properly capture the total exchange. Therefore, when assessing exchange patterns it is important to consider the following: Is the exchange seasonally dependent? If so, what seasonal processes affect the regimes? Is the nocturnal turbulence regime influenced by the frequency of convective downdrafts, which can affect the total canopy-atmosphere exchange of scalars? Therefore, experimental results obtained at different sites, such as in dense tropical forests, are essential to better understand the biosphere-atmosphere interaction and to improve existing knowledge on the nocturnal boundary layer.

From another perspective, in regions often covered by intense convective systems and convective cloud clusters, such as the Amazon region (Betts et al., 2002, 2003; Machado et al., 2002, 2004), the downdrafts may propagate strong downward wind gusts with high velocities (Fujita, 1985; Nelson, 1994; Garstang et al., 1998; Negrón-Juárez et al., 2010, 2017, 2018). When they reach the surface, strong wind outflow with horizontal dimensions is propagated (Fujita, 1985; Garstang et al., 1998). These wind gusts can cause widespread tree damage and mortality through snapping and/or uprooting. These modes of tree mortality occur when the mechanical load caused by wind exceeds the resistance of the trunk or the root-soil anchorage (Nelson, 1994; Marra et al., 2014; Ribeiro et al., 2016).

The severity of damage/mortality caused by extreme wind events influences biomass balance and the functional composition of terra-firme forests in Central and Northwestern Amazon (Urquiza-Muñoz et al., 2021; Negrón-Juárez et al., 2018; Marra et al., 2018, 2014). While some of the carbon contained in dead or damaged trees is decomposed and incorporated into the soil rather than directly emitted into the atmosphere (dos Santos et al., 2016), dead trees represent future CO₂ emissions (Chambers et al., 2004) and potentially methane (CH₄), which contributes to global warming. Using field data (Ribeiro et al., 2016) combined with static and dynamic biomechanical models, Peterson et al. (2019) estimated that wind speeds > 10.75 ms⁻¹ are critical and can snap or uproot trees in terra-firme forests of Central Amazon.

Current knowledge about the effects of wind disturbances in the Amazon forest derives from studies that have quantified the spatial extent of these events using satellite imagery, in particular Landsat (Nelson, 1994; Negrón-Juárez et al., 2010; Chambers et al., 2013; Espírito-Santo et al., 2014), or from post-perturbation forest inventories to assess associated tree damage/mortality and their effects on forest structure and diversity in subsequent years to decades (Marra et al., 2014, 2018; Schwartz et al., 2017; Silvério et al., 2019; Urquiza-Muñoz et al., 2021). The estimated time to recover 90% of pre-disturbance biomass is up to 40 years (Marra et al., 2018) in the Central Amazon and 30 years in the northwestern Amazon (Urquiza-Muñoz et al., 2021).

The goal of our study was to identify the existence of different nocturnal turbulence regimes above a terra-firme forest in the Central Amazon and their possible association with seasonality and proximity to the forest canopy. Furthermore, to assess the effect of extreme winds associated with convective downdrafts on the organization of turbulence regimes in the nocturnal RSL and the potential of observed winds to cause damage/mortality of canopy trees. We recognize that damaging winds and downdrafts in this region are also associated with convective storms and with fronts that can pass during the day, but that is beyond the scope of the current analysis. To our knowledge, this is the first assessment of the influence of extreme winds on turbulence regimes and of the potential of associated wind to promote tree damage/mortality. Apart from a better understanding of the processes of turbulent exchange between the biosphere and the atmosphere, especially in rainforests with dense canopies, our study provides insights into the vulnerability of Amazon forests to

ongoing shifts on the intensity and frequency of convective storms resulting from climate change (Feng et al., 2023).

1.2 Material and methods

1.2.1 Study site and data acquisition

Our data were collected at the *Estação Experimental de Silvicultura Tropical* (EEST, also known as ZF2), Central Amazon, Brazil (Figure 1-1). The EEST is located about 60 km northwest of Manaus (2°36' S, 60°12' W, 130 m a.s.l.) (Araújo et al., 2002) and is predominantly covered by old-growth terra-firme forest with closed canopy (e.g. 593 ± 28 trees ha^{-1}), dense understory and high tree-species diversity (Braga, 1979; Marra et al., 2014). Trees are relatively tall and slender (Oliveira et al., 2008; Ribeiro et al., 2016). The proportion of trees reaching the canopy increased with diameter at breast height (DBH): from 21% for trees 10-20 cm DBH, and 57% for 20-30 cm DBH, up to 100% for trees above 70 cm DBH (Araujo et al., 2020).

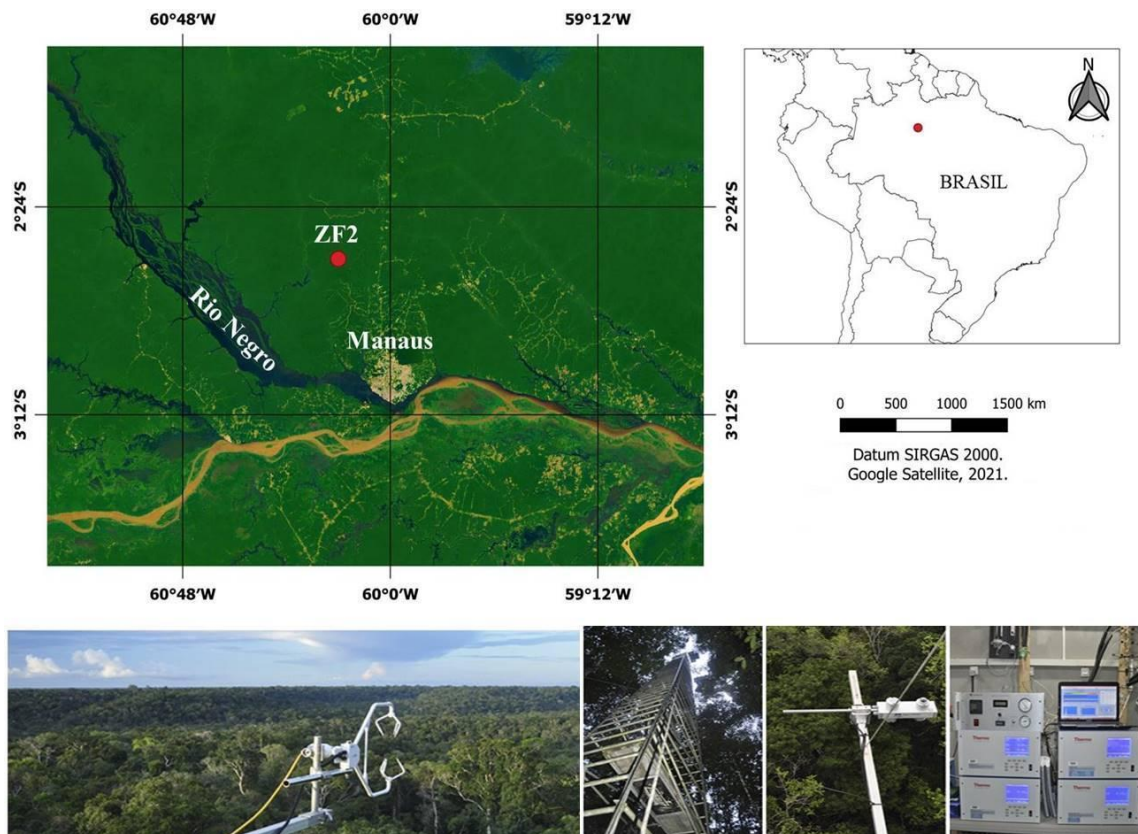


Figure 1-1. Study area at the Estação Experimental de Silvicultura Tropical (EEST) near Manaus, Brazil, and micrometeorological tower and sensors. (Top panel) Map of Brazil showing the study region. (Bottom left) A panoramic view of the forest and the

employed 3D-sonic anemometer. (Bottom center left) Micrometeorological tower. (Bottom center right) Net radiometer. (Bottom right) Ozone analyzers.

The topography at the study area comprises a mosaic of plateaus, slopes and valleys with elevation differences of about 50 m. The vegetation cover on the plateau and slope areas is composed by forest with height varying between 30 to 40 m, maximum surface area density of $0.35 \text{ m}^2\text{m}^{-3}$. In valley areas, the vegetation is smaller with heights from 15 to 25 m, but with significant surface area density more than the $0.35 \text{ m}^2\text{m}^{-3}$ (Tota et al., 2012). The average value of leaf area index for the local vegetation was $6.1 \text{ m}^2\text{m}^{-2}$ (Marques Filho et al., 2005).

Oxisols rich in kaolinite clay dominate plateaus and the upper portions of slopes. The soils on slope bottoms and valleys are sandy and mixed with organic matter (Spodosol) (Telles et al., 2003). The mean annual temperature is 27°C and mean annual rainfall was about 2365 mm for the period between 1971 and 2000, with a distinct dry season between July and September (rainfall $< 100 \text{ mm month}^{-1}$) (Negrón-Juárez et al., 2017).

Our data were collected using sensors installed on a micrometeorological tower of 54 m height. We used the sonic virtual temperature (T) and wind components (u, v and w) measured at 10 different heights (i.e. 1.5, 7.0, 13.5, 18.4, 22.1, 24.5, 31.6, 34.9, 40.4 and 48.2 m above ground) by 3D-sonic anemometers (model CSAT3, Campbell Scientific Inc., Logan, UT, USA) at a rate of 20 Hz. For analyzing the vertical temperature gradient, we used thermo-hygrometer data from sensors (HMP45AC, Vaisala) installed at 35.5 m and 51 m height. These data were available at a 30 min resolution. Net radiation (R_N) was measured with a net-radiometer (model CNR1, Kipp & Zonen, Delft, Netherlands) installed at 39 m height. We used data from April 2014 to January 2015 collected in the course of the Green Ocean Amazon Experiment (GoAmazon). For more details about instruments and respective measurements, see Fuentes et al. (2016).

Ozone concentration (O_3) was measured continuously at 40 m height at a frequency of 1 Hz using an ultraviolet light absorbed gas analyzer (model 49i, Thermo Fisher Scientific Inc., Waltham, MA). For more details on ozone measurements see Gerken et al. (2016). Convective cloud coverage over the study site was assessed with GOES-13 (Geostationary Operational Environmental Satellites) satellite imagery

data available at the Brazilian National Institute for Space Research (INPE) database (<http://satelite.cptec.inpe.br/acervo/>, accessed on 30 May 2021).

Our data were analyzed for two time periods: (i) 55 days in the wet season (from April to May) and (ii) 36 days during the dry season (from July to September). To avoid transitional periods, we used data from 2000 to 0500 local standard time (LST). The dry and wet seasons were characterized based on the seasonality of precipitation using the Global Precipitation Mission (GPM) level-3 product (mm h^{-1}). Estimates were carried out using data downloaded from the website of National Aeronautics and Space Administration (NASA) (<https://giovanni.gsfc.nasa.gov>, accessed on 28 October 2020) for the period between 1st April and 31st December 2014. Overall, we used a product of the Integrated Multi-satellitE Retrievals for GPM (IMERG) at a spatial resolution of $0.1^\circ \times 0.1^\circ$ and at a temporal resolution of 1 month (Final Run IMERG v06). The IMERG algorithm combines information from the GPM satellite constellation and precipitation gauges. The system “final run” uses monthly gauge data to create research-level products.

Local estimates of critical wind-speed (CWS) were compiled from Peterson et al. (2019). According to these authors, the CWS is the wind speed necessary to generate a turning moment at the base of the trunk greater than or equal to the turning moment that results in tree failure. CWS were modeled based on critical turning moments measured locally in a winching experiment that included 60 trees from different species and varying sizes (Ribeiro et al., 2016). We considered CWS values estimated from a dynamic model (i.e. profile method) assuming that tree mortality associated with convective systems can create canopy gaps that contribute to increasing turbulence and wind propagation into the forest. In this model, tree spacing/density, leaf-area index and wind profiles are updated progressively after each tree failure (Peterson et al., 2019). The reported CWS ranged from 10.75 ms^{-1} to 34.5 ms^{-1} .

Before analyzing the micrometeorological, meteorological and profile data, we conducted the following quality control: (i) testing for record completeness and checking for the presence of error flags as proposed by Zahn et al. (2016) and (ii) detection of spikes and dropouts with eventual removal of corrupted/damaged data as suggested by Vickers and Mahrt (1997).

1.2.2 Turbulence regimes and flux estimates

In this study we examine the turbulence regimes at three heights. The first one was defined as near as possible to the top of the forest canopy (i.e. ≈ 35 m; Fuentes et al., 2016), and the other two at the maximum heights where the sensors were installed on the tower (40 and 48 m, respectively). We focused only on the nighttime period (i.e., nocturnal boundary layer) because the diurnal relationship between wind speed and turbulence rarely deviates from linearity (not shown here), since typical daytime conditions at our study site are close to neutrality.

We used two methods for identifying the nocturnal turbulence regimes. The method proposed by Sun et al. (2012, 2016) considers the possibility of external agents (e.g. convective clouds) acting on the atmospheric boundary layer and disturbing its turbulent fields (Garstang et al., 1998; Garstang and Fitzjarrald, 1999; Betts et al., 2002; Dias-Júnior et al., 2017a). Sun's method accounts for the average relationship between the turbulent velocity scale (V_{TKE}) and the mean horizontal wind speed (V). Where $V_{TKE} = TKE^{0.5} = [0.5(\sigma_u^2 + \sigma_v^2 + \sigma_w^2)]^{0.5}$ and $V = (u^2 + v^2)^{0.5}$, in which u , v and w are zonal, meridional and vertical wind components, σ represents the standard deviation of each variable and TKE represents the turbulent kinetic energy. Sun's method identifies the wind speed threshold value (V_L) from the slope change of a straight line around which the data are clustered. For this, we performed a segmented linear regression and selected the V_L with the best coefficient of determination (R^2). In this case, turbulence is compared at a given level with the wind speed at that level.

The second method, reported by Acevedo et al. (2016) is based on the sign inversion of the vertical gradient of V_{TKE} (ΔV_{TKE}). This procedure accounts for the relationship between V_{TKE} and V , but in contrast to Sun's method, turbulence is examined at all levels in terms of the wind speeds at a fixed reference level. When classified by the V at a reference level, the wind speed crossover threshold (V_T) lies where the average ΔV_{TKE} reverses sign for the entire tower layer. This change occurs because in very stable regime the V_{TKE} increase with height and the importance of low-frequency (sub mesoscale fluctuations) phenomena at higher levels contributes to this pattern. In a weakly stable regime, on the other hand, the classical pattern of decreasing turbulence with height is dominant. Therefore, the parameters V_L and V_T are the wind speed values that determine the change from very stable (decoupled

boundary layer) to weakly stable regime (the classical boundary layer coupled to the surface). Here, we assumed ΔV_{TKE} as the difference between 48 m and 35 m.

Fluxes were estimated by the Eddy Covariance method. We calculated the turbulent fluxes of sensible heat as $H = \rho c_p \overline{w'T'}$ and momentum as $\tau = \rho(\overline{u'w'^2} + \overline{v'w'^2})^{0.5}$, where ρ is the density of air (assumed to be 1.225 kgm^{-3}) and c_p is the specific heat of air at a constant pressure. The variables T' and w' , represent the fluctuations of virtual temperature ($^{\circ}\text{C}$) and vertical wind speed (ms^{-1}), respectively. As suggested by Sun et al. (2002) and Vickers et al. (2010), we used mean values over 5 min intervals for all the different statistical moments of the turbulence.

To investigate statistical differences between the very stable and weakly stable regimes, we first used the Mann Whitney U-test considering the identified regimes as grouping variables. Therefore, six tests were performed, one at each height (48 m, 40 m and 35 m) and for the dry and wet seasons. In a second analysis, we applied a factorial analysis of variance (ANOVA) followed by Tukey tests to assess the effects of the interaction between the seasons, heights and turbulence regimes on observed patterns of atmospheric turbulence. Here we use V_{TKE} as dependent variable. For all tests we considered a significance level of 5% and used the software Matlab and Jamovi for data processing and analysis.

In this work V_{TKE} was only used in the analysis of the transition of turbulence regimes, but not as an indicator of the tree damage/mortality. To assess the potential of extreme wind gusts as a mechanism of tree damage/mortality, we compared the CWS data available for our study site and the maximum wind speed observed within our study period, as detailed in section 2.4.

1.2.3 Identification of convective cloud downdrafts

Previous studies have demonstrated that convective downdrafts transport air with high O_3 concentrations and lower T (cold pool) rapidly from the lower-middle troposphere to the surface (Betts et al., 2002; Gerken et al., 2016; Dias-Júnior et al., 2017a; Bezerra et al., 2021).

It is known that in environments with high moisture and strong convection such as in the Amazon, the equivalent potential temperature (θ_e) is also important under unstable atmospheric conditions. Since θ_e changes as a function of moisture (through

mixing ratio) and temperature, the downward transport of cold and dry air leads to an important decrease of θ_e (Garstang and Fitzjarrald, 1999; Betts et al., 2002). However, moisture data were not available for our study site. Alternatively, we used virtual temperatures.

The identification of convective cloud downdrafts in our study was performed using a two-step procedure. We first analyzed the time series of horizontal wind speed, air temperature and surface concentration of O_3 to investigate atmospheric dynamics and the occurrence of gust fronts associated with convective clouds. For this analysis, we averaged data at 48 m (V and T) and 40 m (O_3) height for 5-minute intervals. The data series were mapped by an algorithm that identifies a subtle increase of O_3 concentrations of at least 3 parts per billion (ppb), and a simultaneous decrease of T values of 2 °C within a 1h time window of respective events (Gerken et al., 2016; Dias-Júnior et al., 2017a). During these events, the highest horizontal wind speed registered within one second interval and the five minutes average comprising such an interval were retained. False detections were manually identified and removed. In the next step, GOES-13 imagery were visually inspected for the presence of convective clouds over the study site. Here, we only considered images acquired on days for which abrupt changes in weather variables were detected as described in the previous step. We analyzed the effect of downdrafts associated with turbulence regimes by checking the relationship between V_{TKE} and V. Here, we assumed the minimum and maximum V values to determine the start and end time of respective wind gusts.

1.2.4 Extreme winds as a mechanism of tree mortality

To assess the potential of extreme wind gusts as a mechanism of tree damage/mortality, we compared the CWS data and the maximum wind speed (V_o) measured by our anemometers. The V_o is the maximum horizontal wind speeds (or extremes winds) measured at 5 min time-intervals at 48 m height. These data were analyzed for the different seasons (dry and wet).

Since downdrafts can last from few seconds to (rarely) minutes, we calculated V_o at two other time intervals: 1-minute and 30-seconds windows. To check if V_o differs with the size of time-intervals, we compared computed values (response variable) using ANOVA of repeated measures. Subsequently, we executed an ANOVA to verify potential variations of V_o as a function of seasonality.

1.3 Results

1.3.1 Turbulence regimes in the nocturnal boundary layer

We found a positive relationship between V_{TKE} and V for both dry and wet seasons, and a clear distinction between the two turbulence regimes (Figure 1-2). For events classified as weak turbulence (very stable regime), V_{TKE} increased less than V as indicated by a smaller slope. As V reached a threshold value (V_L), turbulence changed from weak to strong (weakly stable regime) as indicated by a larger increase of V_{TKE} as a function of V and larger slope. V_L increased nonlinearly with the distance from the forest canopy in both seasons. The observed values of V_L at 35 m and 48 m were near 0.7 and 2.3 ms^{-1} (dry) and 0.5 and 1.9 ms^{-1} (wet), respectively. These results are in agreement with previous studies that found that V_L increases nearly logarithmically with height (Sun et al., 2012; Acevedo et al., 2016).

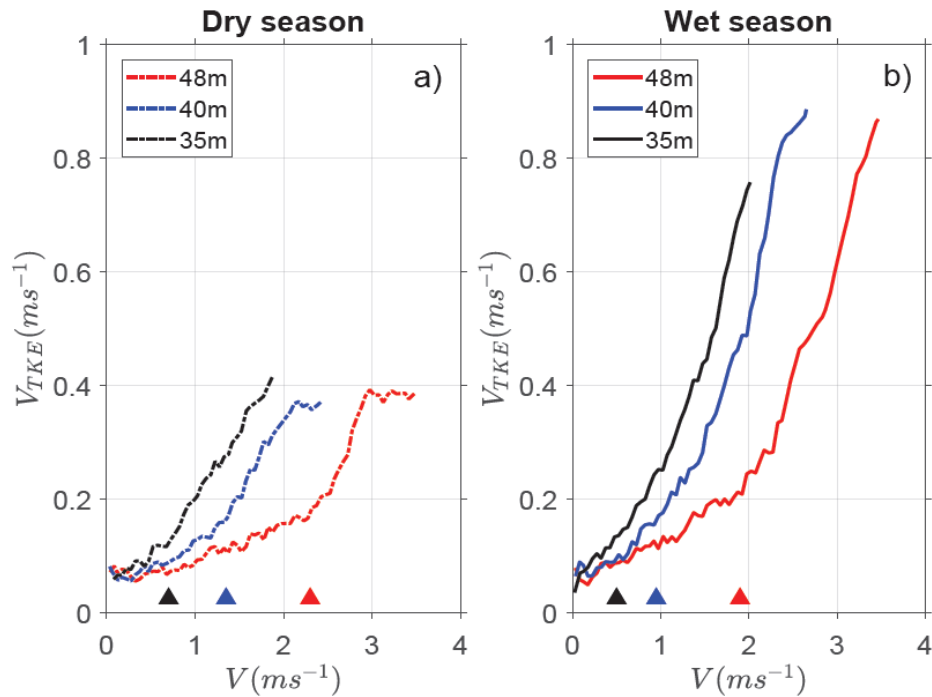


Figure 1-2. Relationship between turbulence velocity scale (V_{TKE}) and mean wind speed (V) at different heights in the (a) dry and (b) wet season during the nighttime at the EEST, Manaus, Central Amazon. Triangles indicate the threshold wind speed (V_L) at which the very stable changed to weakly stable regime.

Although the observed increase of V_L with height supports previous research on non-vegetated surfaces, the thresholds identified at 35 m in our study site were significantly smaller than those observed at a pasture area at the Federal University of Santa Maria (in south Brazil) (V_L of 3.0 ms^{-1} at 30 m height) (Acevedo et al., 2021) and at project FLOSS II (Fluxes over Snow Surfaces) conducted in Colorado (US) (V_L of 5.3 ms^{-1} at 30 m height) (Acevedo et al., 2016). Such differences show that the structure and roughness of dense forests influence the change of the turbulence regime. For a proper comparison we considered the observation level with respect to the zero plane displacement height (d). We used $d=31.5 \text{ m}$ based on estimates provided by Viswanadham et al. (1990) and Chor et al. (2017) for central Amazon, in which $d=0.88h$ and $d=0.9h$, respectively, where h is the canopy height (35 m at our study site). Doing that, the V_L values found at 35 m ($z-d = 3.5 \text{ m}$) in both seasons are smaller than the value of 2.04 ms^{-1} found at 2 m in FLOSS II (Acevedo et al., 2016) and that of 1.5 ms^{-1} found at 1.5 m in the CASES-99 (experiment carried out in southeast Kansas, US) (Sun et al., 2012). Similarly, the V_L found for both seasons at the EEST at 48 m ($z-d=16.5 \text{ m}$) are smaller than those at 15 m in FLOSS (5.04 ms^{-1}) and at 10 m in CASES-99 (4.5 ms^{-1}). The V_L found at 48 m at EEST site in the wet season is also smaller than the value reported by Acevedo et al. (2021) at 14 m in Santa Maria (2.2 ms^{-1}). Still, this value is close to that we found during the dry season at the EEST. In general, the V_L values between turbulence regimes was smaller at the EEST site than that reported in previous studies at similar heights, even when the zero-plane displacement height is taken into account. This relates to V_L being generally smaller above rough surfaces compared to smooth surfaces (Mahrt et al., 2013; Vignon et al., 2017; Guerra et al., 2018), a simple consequence of turbulent mixing being larger above a rough surface than above a smooth one at a same mean wind speed.

From the nighttime turbulence regimes identified at 48 m height, around 77% (dry) and 65% (wet) of them were associated with very stable regime. In contrast, weakly stable regime corresponded to 23% (dry) and 35% (wet) of the events. Furthermore, V_L was larger during the dry season than in the wet season at the three investigated heights (Figure 1-2 and Table 1-1). This result is in agreement with Acevedo et al. (2021), who showed that V_L increases linearly with net radiative loss at the surface at three mid-latitude sites. According to these authors, the fully turbulence regime occurs when the mean wind speed is large enough to support heat fluxes

capable of transferring back to the surface part of the energy lost radiatively. Therefore, such a minimum heat flux and corresponding minimum wind speeds must be larger when the net radiative loss is also larger. The relationship between V_L and R_N is further discussed in Section 4.

Table 1-1. Wind speed threshold (V_L) at studied heights for the dry and wet seasons at the EEST, Manaus, Central Amazon.

Season	V_L at 48 m (ms^{-1})	V_L at 40 m (ms^{-1})	V_L at 35 m (ms^{-1})
Dry	2.3	1.35	0.7
Wet	1.9	0.95	0.5

We also identified the transition between regimes as a function of V at 35 m, for which the ΔV_{TKE} reverses sign for both dry and wet seasons (Acevedo et al., 2021). Figure 1-3 shows that the V_{TKE} increased with height for weak winds in both seasons. When V at 35 m exceeds a sharp threshold, this pattern is reversed, so that V_{TKE} decreases with height under sufficiently strong winds. Such a threshold was approximately 0.65 ms^{-1} and 0.50 ms^{-1} for the dry and wet seasons, respectively. However, when the same analysis is performed in terms of V at 48 m (Figure 1-4), there is no gradient reversal and V_{TKE} decreases with height for all observed wind speeds.

These findings suggest that the relationship between V and V_{TKE} at the different heights is not trivial and may be associated with the occurrence of phenomena such as low-level jets and density currents (Greco et al., 1992; Dias-Júnior et al., 2017b; Corrêa et al., 2021). When the wind speeds are weak at higher levels, the wind profile is often distorted and maximum values can occur near the surface (Acevedo et al., 2016).

The V_T values identified by the vertical gradient method at 35 m height during the dry (0.65 ms^{-1}) and wet (0.50 ms^{-1}) seasons were similar to the V_L values obtained with Sun's method. This result corroborates the existence of nighttime turbulence regimes at the EEST during the two seasons. As the values of V_T and V_L were similar, hereafter we used the values of V_L as an indication for changes of turbulence regimes, since it was possible to identify it at all heights investigated here.

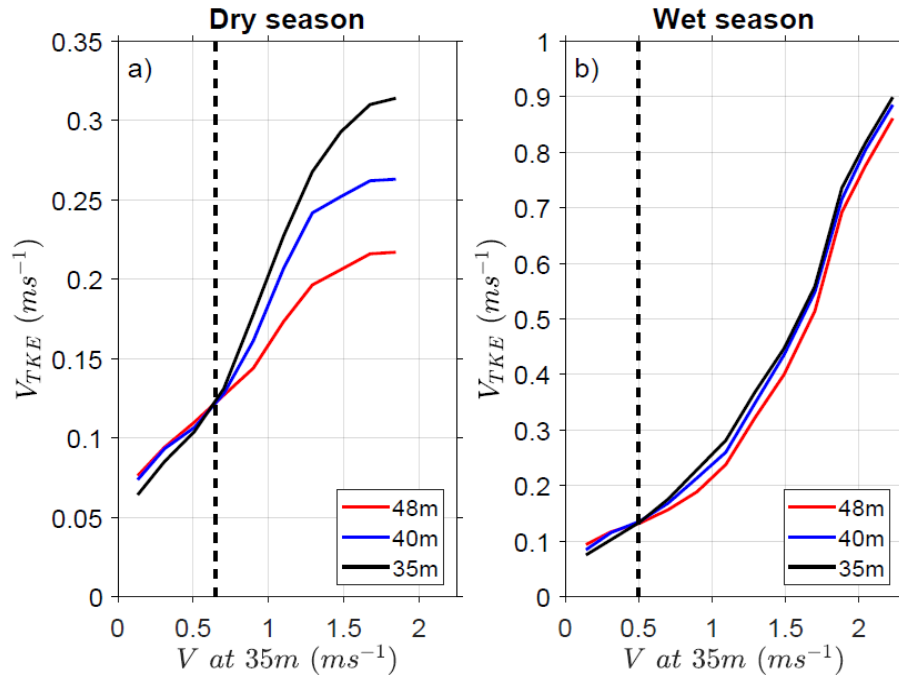


Figure 1-3. Turbulence velocity scale (V_{TKE}) as a function of the mean wind speed (V) at 48 m during the nighttime at the EEST, Manaus, Central Amazon. The solid lines are V_{TKE} at different heights in the (a) dry and (b) wet season. Black vertical dashed line indicates when V_{TKE} reverses its sign and exceeds the crossover threshold (V_T) at which the very stable changes to weakly stable regime.

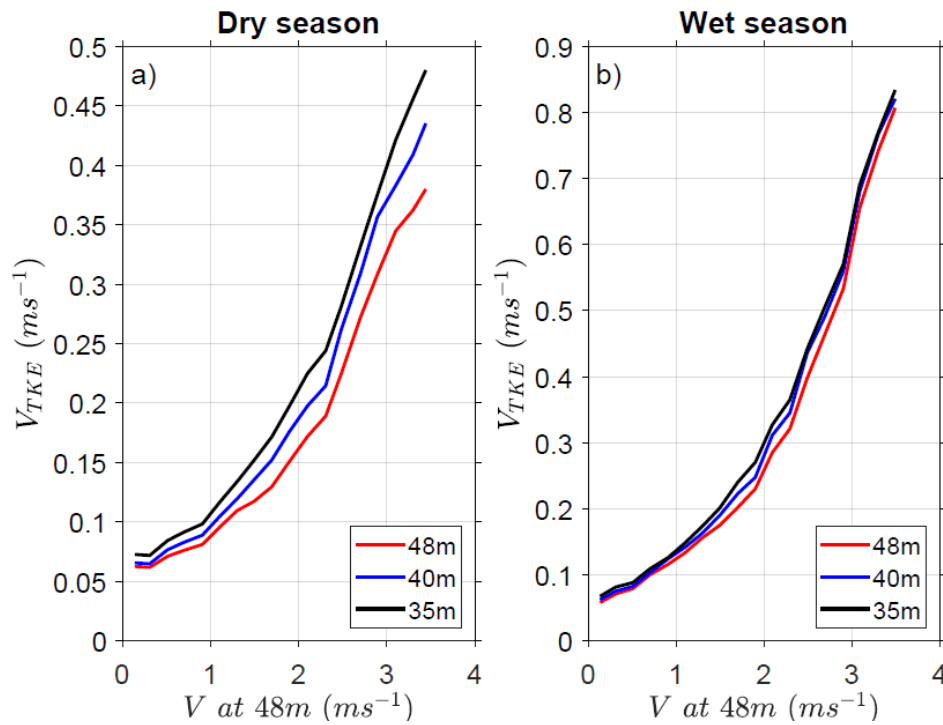


Figure 1-4. Turbulence velocity scale (V_{TKE}) as a function of the mean wind speed (V) at 48 m during the nighttime at the EEST, Manaus, Central Amazon. The solid lines are V_{TKE} at different heights in the (a) dry and (b) wet season.

Our results evidenced the existence of different nocturnal turbulence regimes in the terra-firme Amazon forests. Moreover, they indicated that turbulence intensity decreased with the proximity to the canopy (Figure 1-5a). V_{TKE} during the weakly stable regime was larger than in the very stable regime (Figure 1-5b). Overall, the wet season had larger values of V_{TKE} than the dry season (Figure 1-5c).

These larger values of V_{TKE} in the wet season are likely associated with a lower wind speed threshold between regimes in the wet season (Figure 1-2). Since the threshold is smaller, it is plausible considering that it is easier to overcome this threshold and establish “large” V_{TKE} in this case. Another reason that could result in larger values of V_{TKE} in the wet season would be simply because the wet season has higher wind speeds values than the dry season. Nonetheless, this was not the case, as can be seen in Figure 1-5d-f. Our result may be related to other factors, such as radiative loss, thermal gradient, among others.

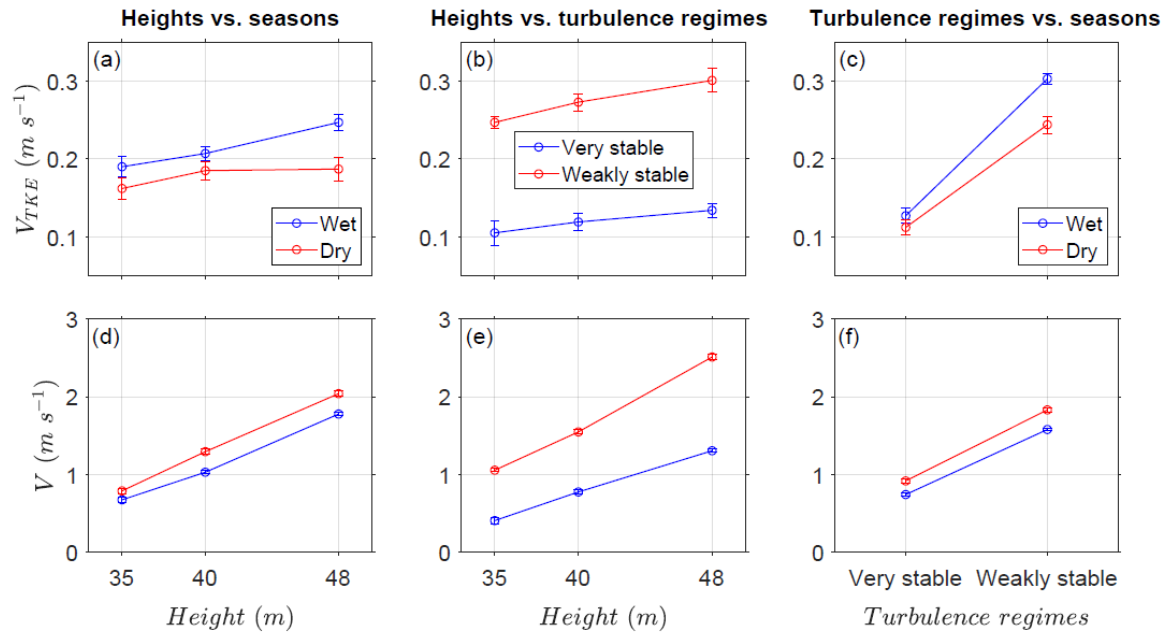


Figure 1-5. Turbulence velocity scale (V_{TKE}) and mean wind speed (V) as a function of (a,d) heights vs. seasons, (b,e) heights vs. turbulence regimes and (c,f) turbulence regimes vs. seasons during nighttime at the EEST, Manaus, Central Amazon. Legend: circles and vertical bars indicate mean values and the 95% confidence intervals, respectively.

1.3.2 Turbulent fluxes in the nocturnal boundary layer

We investigated the differences between the turbulent fluxes during the occurrence of very stable and weakly stable regimes, and found an interesting relationship between V and the sensible heat (H), and momentum (τ) fluxes at 35 m height during both seasons. Bins of V each 0.2 ms^{-1} were used to calculate the average values and standard deviations of the fluxes. H values (Figure 1-6a, Figure 1-6b) changed in response to the turbulence shifting from the weak to the strong in both seasons (Table 1-2). On average, the H flux in the weakly stable regime corresponded to about 88% of the total H observed in both seasons. Our findings corroborate those of previous research conducted in Eastern Amazon (Dias-Júnior et al., 2017b). Importantly, H values were more intense during the dry season ($p = 0.001$). This result is possibly associated with larger net radiative loss due to reduced cloud cover and atmospheric column water vapor load (Collow and Miller, 2016), which leads to less longwave radiation reaching the canopy.

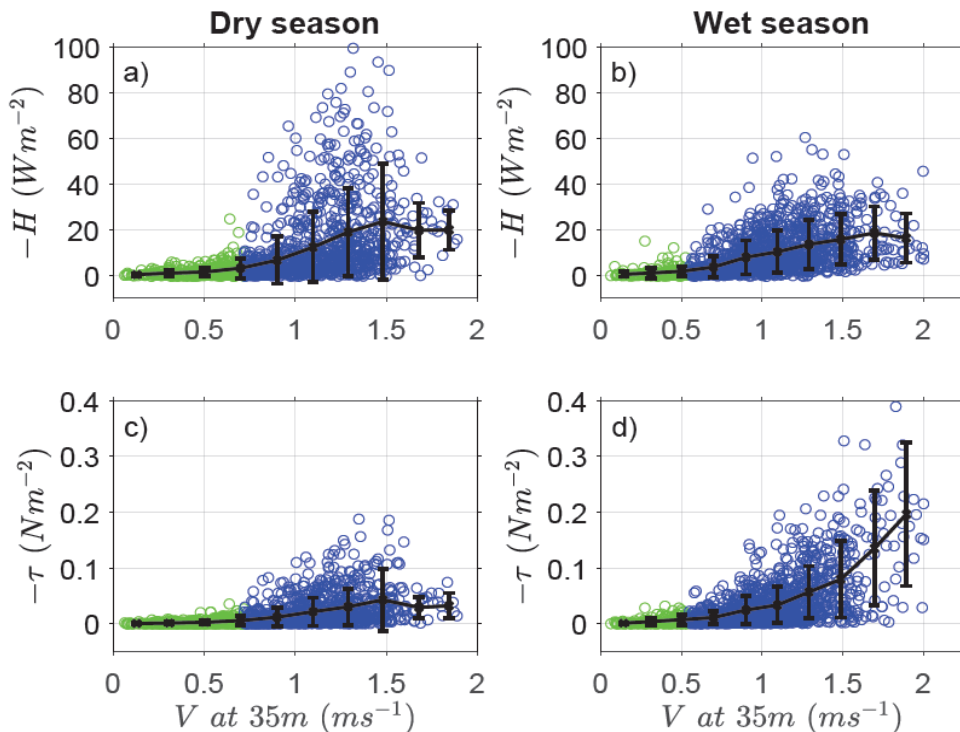


Figure 1-6. Mean and standard deviation of sensible heat flux (H) and momentum flux (τ) as a function of the wind speed (V) at 35 m during nighttime at the EEST, Manaus, Central Amazon. The left and right panels indicate the dry and wet season, respectively. Green and blue circles indicate the very stable and weakly stable regimes, respectively.

The τ fluxes showed a similar behavior to that of H , in which the flux in the weakly stable regime was larger than in the very stable regime ($p \leq 0.001$), and reached on average 89% of the total τ flux in each season (Figure 1-6c and Figure 1-6d). Such variations may be related to atmospheric stability conditions. The wet season is less stable (on average), that is, the strong turbulence events (i.e. weakly stable regime) occurs more frequently and, consequently, the momentum fluxes are greater. Overall, the weakly stable regime contributed more significantly to sensible heat and momentum fluxes than the very stable regime (Table 1-2), in both seasons.

Table 1-2. Mean values ($\pm$ 95% confidence interval) of sensible heat flux (H) and momentum flux (τ) for the dry and wet seasons at the EEST, Central Amazon, Brazil.

Turbulence regimes	H Dry (Wm^{-2})	H Wet (Wm^{-2})	τ Dry (Nm^{-2})	τ Wet (Nm^{-2})
Very stable	-1.8 ± 2.5	-1.3 ± 2.0	-0.3 ± 0.4	-0.5 ± 0.6
Weakly stable	-13.4 ± 17	-9.8 ± 9.6	-2.3 ± 3.0	-4.1 ± 5.4

1.3.3 Relationship between deep convection and nocturnal turbulence regimes

During the night of 06 August 2014 and 12 April 2014, gust fronts from downdrafts reached the tower around 2110 LST (dry season) and 2325 LST (wet season), respectively (Figure 1-7). These events were evidenced from observations of (i) decreasing T (around 2°C at both seasons) and (ii) simultaneous increasing O_3 of approximately 12 ppb (dry) and 15 ppb (wet). The pre-gust measurements of V ($\approx 2 \text{ ms}^{-1}$) were similar on both nights (Figure 1-7c and Figure 1-7g). However, V reached maximums of 7.5 (dry) and 15 ms^{-1} (wet) during the observed downdrafts. At this time, the standard deviation of wind vertical velocity (σ_w) increased substantially (Figure 1-7d and Figure 1-7h). As the surface cools and uncouples the boundary layer from the above troposphere, the near-surface O_3 values are low (3-5 ppb). This pattern allows for a clear recognition of night downdrafts transporting air with higher ozone and lower temperature to the surface. We observed this pattern on other nights and identified 16 downdrafts (eight different nights in each season) over the studied period. Overall, these events occurred between 2000 LST and 2300 LST. Finally, the GOES-

13 imagery indicates the existence of cloudiness at the times of the events (Figure 1-8) and also on the other 16 nights that we studied here.

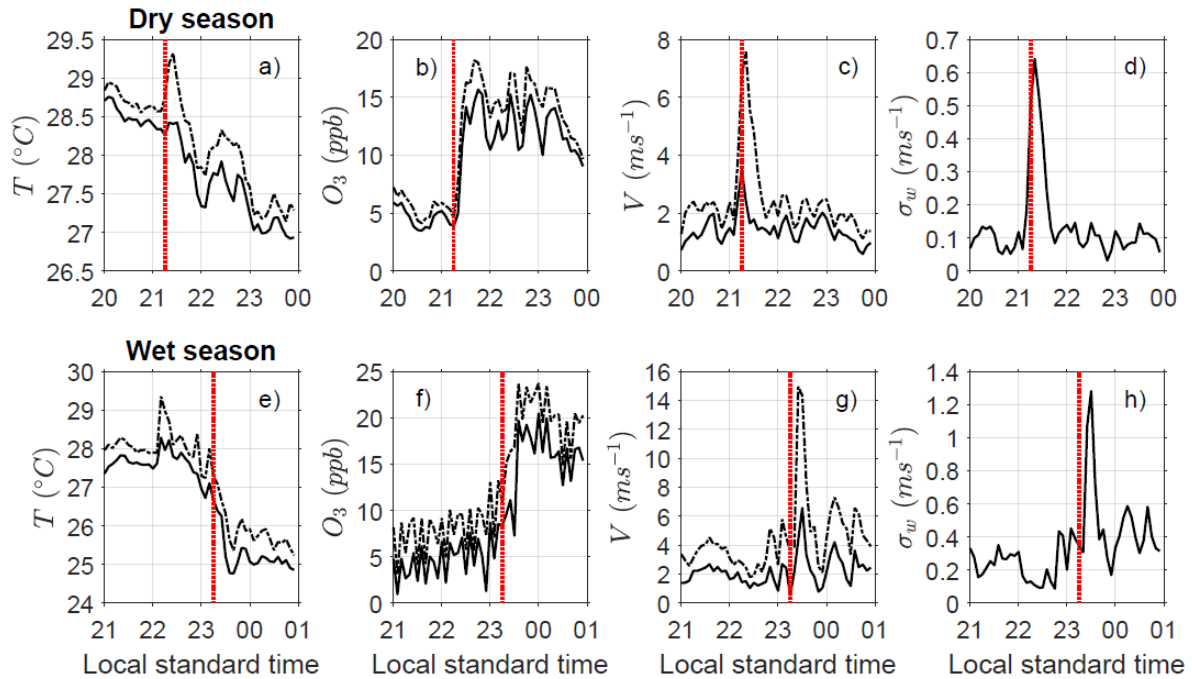


Figure 1-7. Times series of key variables used to identify downdrafts at the EEST, Central Amazon, Brazil. The upper and bottom panels show the downdrafts from 06 August 2014 and 12 April 2014, respectively. Legend: virtual temperature (T), ozone (O_3), horizontal wind speed (V) and standard deviation of wind vertical velocity (σ_w). Solid and dashed lines indicate the mean and maximum values, respectively. Vertical red-dotted lines indicate the starting time of downdrafts.

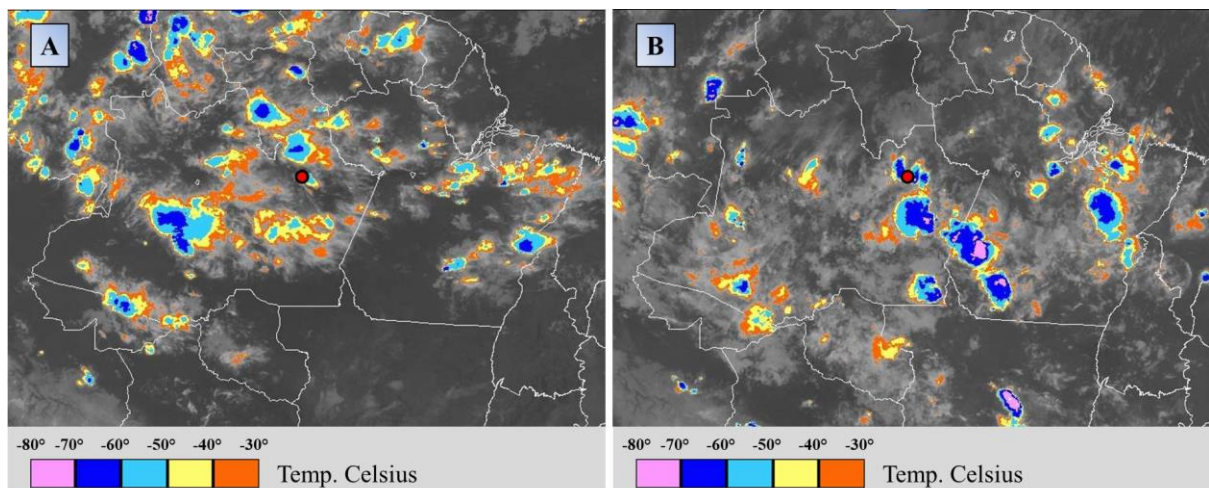


Figure 1-8. GOES 13 imagery for (a) 07 August 2014 at 0100 UTC and (b) 13 April 2014 at 0300 UTC when downdrafts reached the micrometeorology tower at the EEST (red dot), Central Amazon, Brazil.

These results support previous studies, such as that by Betts et al. (2002) who found that nighttime convective downdrafts coupled the surface to the lower troposphere and transported down air with larger O_3 and lower equivalent potential temperature. Dias-Júnior et al. (2017a) also reported that the downdrafts produce O_3 enhancement events and an increase in V values, in addition to the occurrence of air divergence during the horizontal propagation of density currents.

We observed that all downdrafts happened during or after the transition to the weakly stable regime (Figure 1-9). This pattern evidenced that downdrafts may influence the turbulence characteristics near the surface since they are associated with a weakly stable regime (strong turbulence). However, the weakly stable regime was not represented completely by these events, which suggest that the strong turbulence may be associated with the occurrence of other phenomena. Similar results were previously reported for Central Amazon by Bezerra et al. (2021), who observed that downdrafts generated by a squall line occurred only during the strong turbulence regime.

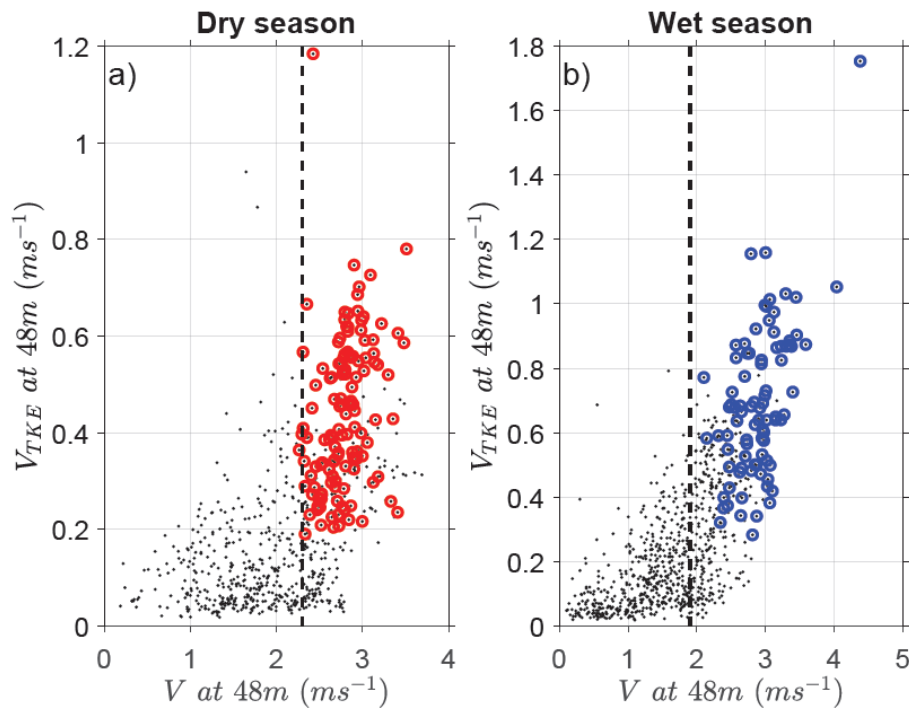


Figure 1-9. Turbulence velocity scale (V_{TKE}) as a function of the mean wind speed (V) measured at 48 m in the (a) dry and (b) wet season during nighttime at the EEST, Manaus, Central Amazon. The dots correspond to mean values calculated over 5-min periods. Thick circles indicate the range of occurrence of the downdrafts. Vertical

black-dashed lines mark the threshold wind speed (V_L) at which the very stable changes to weakly stable regime.

1.3.4 Extreme winds as a mechanism of tree mortality

During the weakly stable regime associated with nocturnal downdraft events, the speeds and thus destructive potential of winds varied between seasons. The greatest wind speeds were identified during the wet season ($V_o = 14.96 \text{ ms}^{-1}$). This value was approximately four times higher than on nights without downdrafts, and exceeded the CWS of three out of the studied trees by Peterson et al. (2019). In contrast, maximum wind speeds in the nocturnal period of the dry season ($V_o = 7.57 \text{ ms}^{-1}$) did not exceed the CWS of the studied trees.

Here we did not assess the direct effect of high wind-speeds on trees, but rather evaluated the destructive potential of these based on observational data acquired at the same study site. Importantly, the occurrence of excessive wind speeds does not necessarily result in tree damage and mortality. Further studies are needed to understand the link between wind speeds, canopy structure and tree motion in these diverse forests. Still, the strong dissipation of air to layers below the forest canopy (presented in Section 4) are rarely observed in the absence of downdrafts. Therefore, the wind gusts described in our study not only reached extreme speeds but also penetrated the forest canopy, and had the potential to cause damage and mortality of trees of different sizes, both directly and indirectly.

The mean V_o values varied significantly between the analyzed time-intervals ($p \leq 0.001$, Figure 1-10a). As expected, the values were higher for the 30 seconds interval ($5.97 \pm 2.85 \text{ ms}^{-1}$, mean $\pm$ 95% confidence interval). For the 1 min and 5 min intervals, V_o was $5.35 \pm 2.68 \text{ ms}^{-1}$ and $3.63 \pm 2.45 \text{ ms}^{-1}$, respectively. Results from subsequent ANOVA showed that the interaction between seasons and time intervals was significant ($p \leq 0.001$, Figure 1-10b). Post-hoc Tukey tests showed that observed variations in V_o were not significantly different between seasons at both 1 min ($p = 0.057$) and 5 min ($p = 1.000$) intervals. Nonetheless, V_o varied as a function of seasonality for our shorter time interval (i.e. 30 seconds). These results show that for extreme wind gusts that may last a few seconds such as those causing tree damage and mortality in the Amazon, our 30 second interval is likely too large and may have underestimated maximum V_o values.

Although daytime patterns were not a focal aspect in our study, wind speed reached the highest values (up to 22 ms^{-1}) during this period, in the dry season. In contrast to the relatively low destructive potential of nocturnal winds, this value could topple 73% of trees previously investigated in our study site (Ribeiro et al., 2016; Peterson et al., 2019). These observations reinforce the importance of extreme wind as a major natural mechanism of tree damage and mortality in these forests (Nelson, 1994; Chambers et al., 2009, 2013; Negrón-Juárez et al., 2017, 2018; Marra et al., 2018).

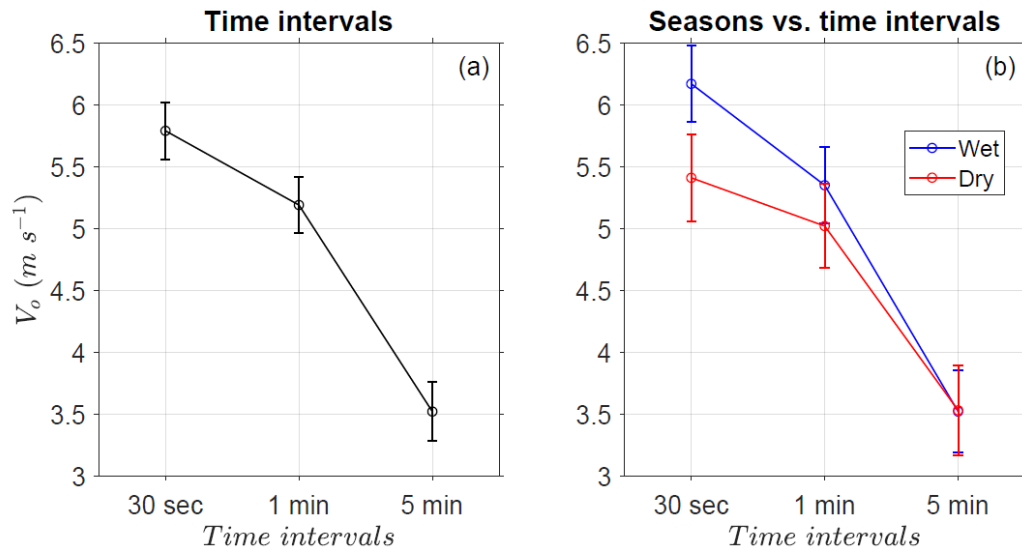


Figure 1-10. Observed wind speed (V_o) as a function of the (a) three time intervals and (b) seasons vs. time intervals at the EEST, Manaus, Central Amazon. Legend: circles and vertical bars indicate mean values and the 95% confidence intervals, respectively.

In fact, a higher destructive potential of diurnal winds may be expected, since in moist environments such as the Amazon forest, surface warming promotes upward movements that increase the low-level moisture convergence and intensify convection. Moreover, the results from the nighttime period provide evidence on the importance of downdrafts on the propagation of extreme winds downward below the canopy.

1.4 Discussion

Coupling between the canopy and the atmosphere occurs when turbulence provides continuous mixing. At nighttime, however, the mixing can be inhibited by the

presence of a stable layer. Therefore, two contrasting regimes are observed. In the weakly stable regime, the wind speed provides heat fluxes that are large enough to continuously transfer the energy lost by radiation back to the surface. In the very stable regime, heat fluxes are not transferred continuously to the surface, leading to strong temperature drops. This allows for the establishment of an enhanced thermal gradient, which further inhibits mixing. Thus, the coupling between the forest and the atmosphere is favored by the continuous turbulence observed in the weakly stable regime.

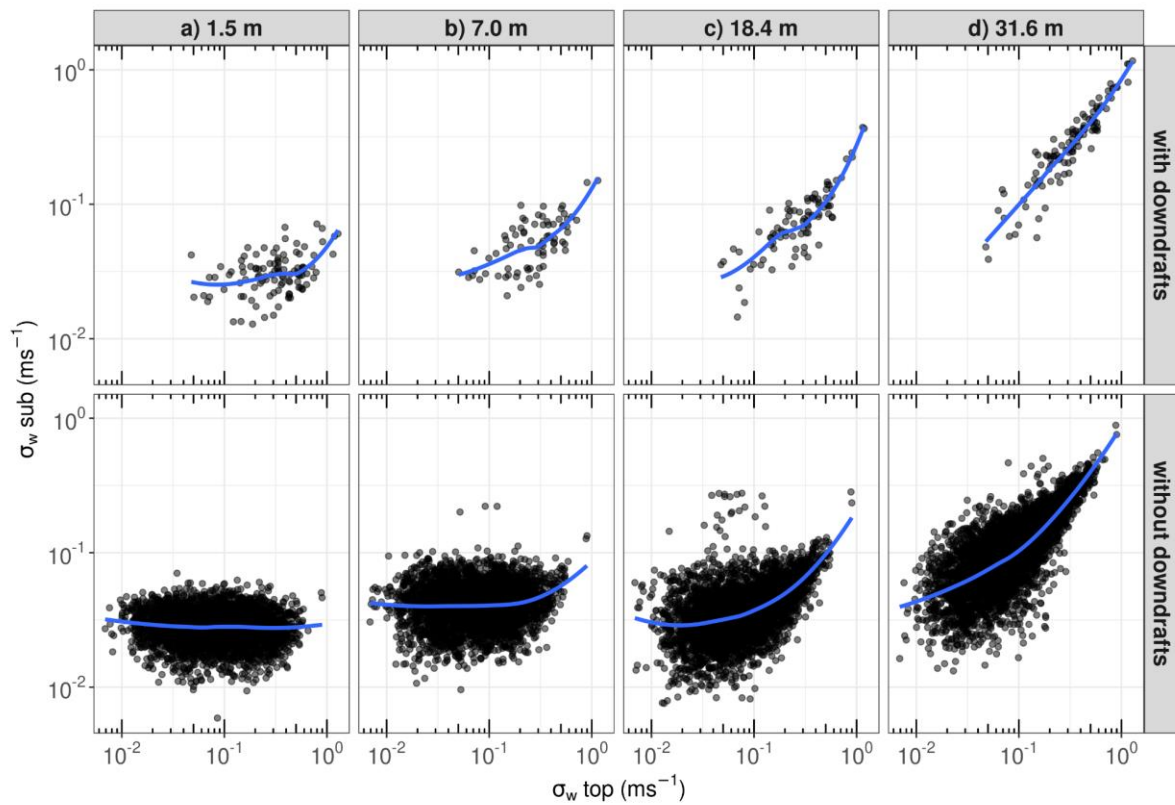


Figure 1-11. Standard deviation of wind vertical velocity (σ_w) between the forest understory (sub) and above-canopy (top at 48 m) heights. The dots correspond to periods of 5 min data. Dataset in which downdrafts occurred (Top panels). Dataset without the occurrence of downdrafts (Bottom panels). The columns a, b, c and d indicated the investigated sub-canopy heights.

We evaluated the degree of coupling in two ways. First, a comparison of σ_w on above (48 m) and sub-canopy heights following Thomas et al. (2013). Figure 1-11 shows the comparison between such variables but it is not easy to visualize the trends with only plotting the data points. Thus, we added a Locally Weighted Regression

(LOWESS) to the graph (solid line). Here we focus on 4 forest understory heights (1.5 m, 7.0 m, 18.4 m and 31.6 m) and on datasets with and without downdrafts. During weak winds (i.e. without downdrafts), the turbulence strength at 1.5 m and 7.0 m was largely independent of that observed above canopy. After 18 m height, the σ_{wsub} was linearly correlated with σ_{wtop} (Figure 1-11c) indicating the occurrence of a coupled canopy condition. On the other hand, the extreme winds associated with downdrafts were propagated into the canopy at all heights, and the threshold of σ_{wtop} (i.e., when the correlation became linear) increased as flow above the canopy reached the ground.

Second, in order to quantify the coupling between the canopy and the atmosphere we calculated the temperature gradient (ΔT) between 51 m and 35 m height, in the dry and wet seasons. The dependence of ΔT on V (Figure 1-12) shows the differences of coupling and subsequent mixing for the two regimes. A large thermal gradient occurs under the lowest wind speeds. This was observed when ΔT averaged 0.58 °C and 0.25 °C during the dry and wet seasons, respectively.

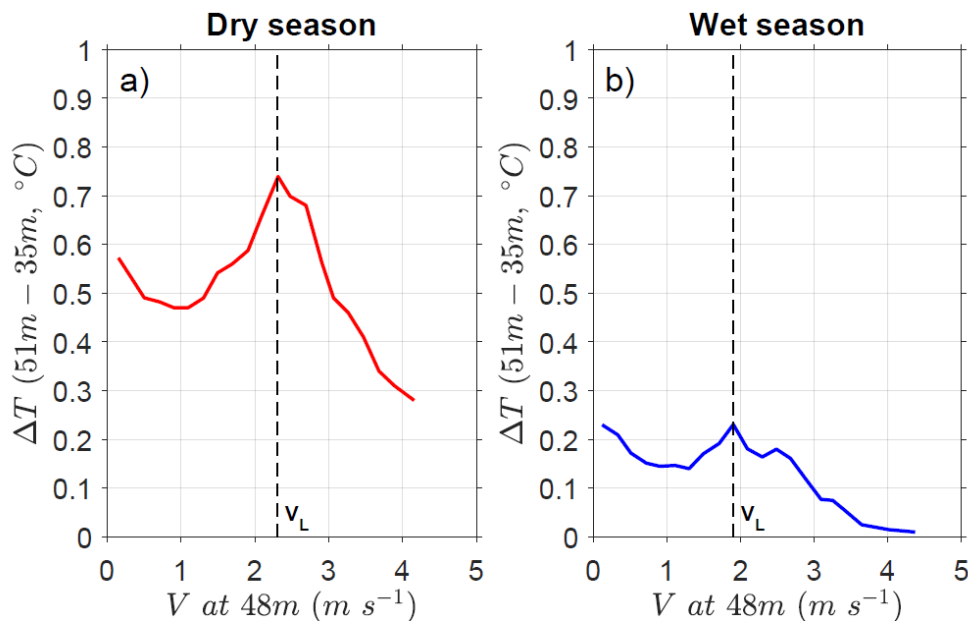


Figure 1-12. Differences of mean virtual temperature between 51 m and 35 m height as a function of the mean wind speed (V) at 48 m in the dry (a) and wet (b) season at the EEST, Manaus, Central Amazon. Dashed lines indicate the threshold wind speed (V_L) at which very stable change to weakly stable regime.

In the dry and wet seasons, ΔT peaked at V_L and reached 0.73 °C and 0.25 °C, respectively. A similar ΔT maxima at the transition between regimes was observed by

Acevedo et al. (2016) for the FLOSS II dataset. These authors argue that this pattern was associated with an enhanced heat-flux convergence at similar ranges of wind speed reported by Acevedo et al. (2021) for the CASES-99 dataset. Smaller gradients typical of mixed conditions occur when wind speeds are higher.

The relationship between turbulence and net radiation is a reliable criterium for distinguishing regimes. When the radiative loss is high, wind speeds tend to grow proportionally to allow compensative heat-fluxes (Acevedo et al., 2021). It is known that cloud cover plays an important role in R_N . Von Randow et al. (2004) showed that in the southern Amazon, reduced cloud cover in the dry season results in increased radiative loss. The opposite occurs in the wet season. In this context, we have assessed the frequency distribution of R_N (30-min averages) for our studied period (from 8pm to 5am) and only for the nights in which downdrafts occurred (2h before and after respective events). In the dry season, the two lowest values of R_N (highest radiative loss) were observed between -40 and -60 Wm^{-2} (Figure 1-13a). This variation indicates that there were relatively fewer clouds over the site. By contrast, R_N was more uniform during the wet season, with values ranging from -10 to -50 Wm^{-2} (Figure 1-13b). Radiative loss was also higher in the 2-hour interval before and after downdrafts observed in the dry season (Figure 1-13c,d).

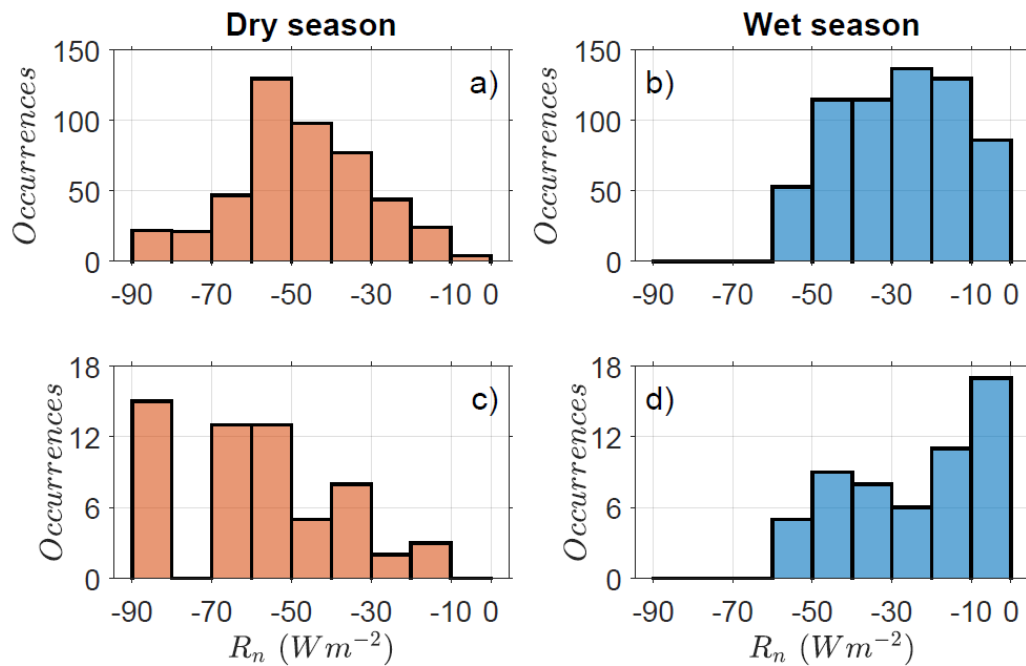


Figure 1-13. Distribution of the net radiation (R_N) for all studied days (top panels) and for those on which downdraft were identified (bottom panels) in the dry (red bars) and wet (blue bars) season during the nighttime.

This R_N pattern is related to the change in cloud cover and moisture loads at each season (Collow and Miller, 2016). Above the Amazon forest, single-cell clouds are frequent in the dry season. This contrasts mesoscale convective-systems (multiple cells) that are more frequent during the wet season (Gerken et al., 2016). Multiple cells and water vapor can trap some of the outgoing infrared radiation emitted by the Earth and radiate it back downward, which can reduce the radiative loss at the surface. This explains why the transition between regimes occurs at higher wind speed V_L in the dry season. During this period, the shallower cloud cover and the lower water vapor load of the atmospheric column allows for a larger loss of radiation than that observed in the wet period.

Both single and multiple cells are known to produce downdrafts (Gerken et al., 2016; Dias-Júnior et al., 2017b). Furthermore, we showed in this study that downdrafts are one of the main causes of transition from turbulence regimes above the Amazon forest (Figure 1-9). We investigated the profile of four turbulent parameters during a night-time downdraft (July 24, 2014). H values, which were initially close to zero, turned strongly negative when the downdraft reached the tower at around 11 pm local time (Figure 1-14a). Similarly, there was an increase in parameters associated with the intensity of turbulence, such as σ_w , TKE and friction velocity (u_*) (Figure 1-14b, c and d, respectively). Furthermore, this strong dissipation of air to strata/layers below the forest canopy are rarely observed in the absence of downdrafts. The air layer from the soil surface is largely decoupled from layers above the canopy (Thomas et al., 2013; Freundorfer et al., 2019; Cava et al., 2022). Santana et al. (2018) provided evidence that atmospheric eddies generated above the canopy can hardly penetrate the region below 0.5 h (h is the canopy top). This pattern was reported for different sites in the Amazon. However, observations (Bezerra et al., 2021) and numerical simulations (Serra-Neto et al., 2021) showed that under strong wind conditions, turbulence below the forest canopy was intensified and the scalar mixing more efficient.

The penetration of wind gusts inside the canopy increases the probability of tree damage and mortality. However, since the peak velocities may be underestimated when using single tower-measurements and we did not have data on risk of tree mortality, further studies are needed to describe the return frequency of such gusts and the relationship between speed and the disturbance severity.

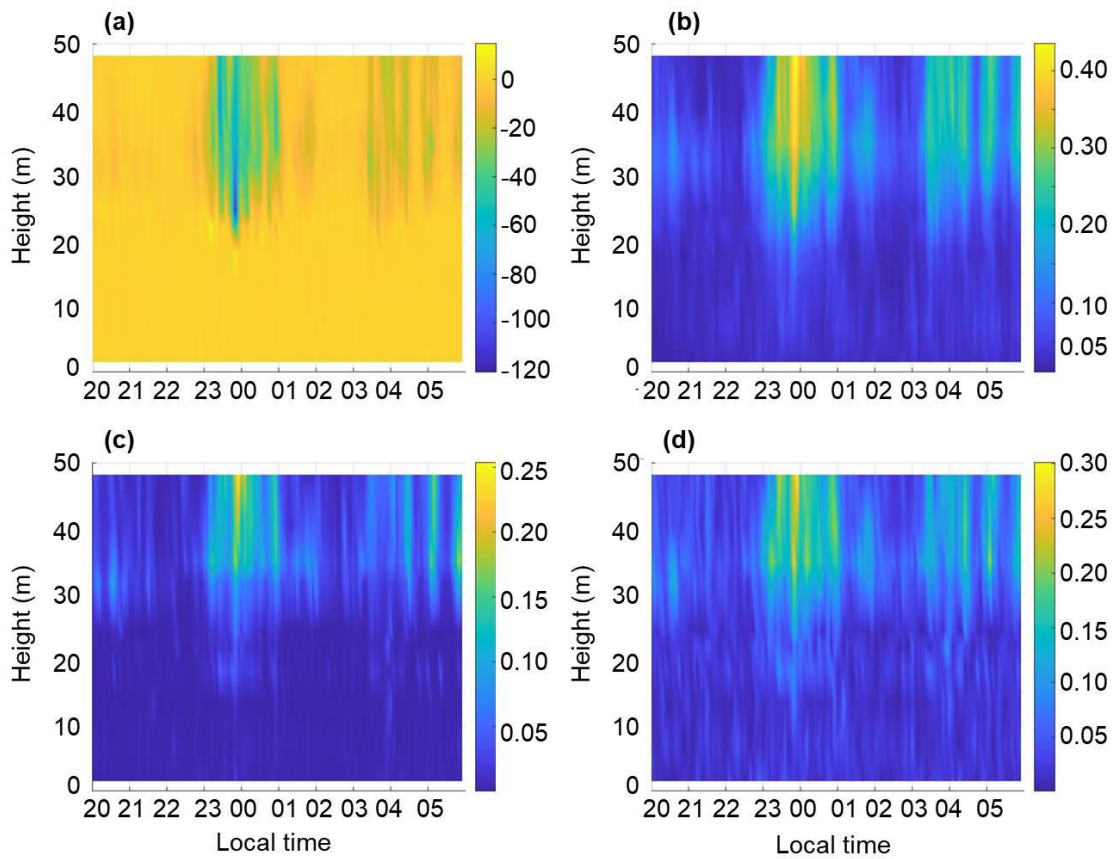


Figure 1-14. Vertical profile of: (a) sensible heat flux (Wm^{-2}), (b) standard deviation of vertical wind (ms^{-1}), (c) Turbulent Kinetic Energy (m^2s^{-2}) and (d) friction velocity (ms^{-1}) at the night of July 24, 2014.

Our study has limitations which shall be addressed in future research. First, the patterns we described for the central Amazon may not occur in other regions with different vegetation structure. This highlights the need of studies based on extended datasets from other regions and for heights above 48 m. Second, when analyzing fast-response data from a single tower, the downdrafts may be not fully captured, and their magnitude and duration may be underestimated. Last, the link between wind speed, canopy structure and tree motion in these diverse forests is currently unknown. Nonetheless, our study stress the importance of datasets including a range of heights and seasons for detecting processes and mechanisms regulating turbulence and wind-tree interactions in dense tropical forests. A better understanding of these interactions is key for the parameterization of more robust and realistic in numerical models. In addition, our findings provided insights into the importance of wind gusts to the ecology and dynamics of Amazon forest.

1.5 Conclusions

This study provides three novel contributions. The first is the identification of different turbulence regimes and their patterns in terms of seasonality and proximity to the forest canopy in the nocturnal RSL. The second is the assessment of the effects of near surface wind gusts (propagated from downdrafts) on the organization of turbulence regimes. Finally, it provides evidence on the occurrence of extreme wind gusts associated with convective downdrafts, with potential to promote damage and mortality of canopy trees. These aspects highlight the strong interactions between atmospheric and biospheric processes and mechanisms regulating forest structure and dynamics.

Two turbulence regimes were identified: the very stable (weak turbulence) and weakly stable regime (strong turbulence). The wind speed threshold that marks the transition between the regimes increases nonlinearly with the distance from the ground under non-vegetated surfaces (Sun et al., 2012; Acevedo et al., 2021). Our study provides evidence that such pattern also occurs in closed canopy forests. In addition, new knowledge was obtained:

- i) The average wind speed threshold for turbulence regime varies seasonally, and was relatively larger in the dry season at all heights as a consequence of a higher radiative loss from the surface during this period. Furthermore, the change of turbulence regime was influenced by the structure and roughness of the forest. This pattern was highlighted by relatively lower thresholds of wind speed compared to previous studies at mid-latitudes, and can be explained by the greater turbulent mixing above rough surfaces for a given mean wind speed.
- ii) Near-surface wind gusts (convective downdrafts) occurred only during the weakly stable regime and were one of the main drivers of the observed turbulence regimes transition. Nevertheless, not all weakly stable regime were associated to such events.
- iii) Full coupling state of wind flow among layers above and within the canopy occur during downdrafts. At nights without extreme winds, coupling along the canopy profile occurred only above 18 m height.

iv) The destructive potential of winds propagated during downdrafts was approximately four times higher than on nights without downdrafts in the wet season. By contrast, the wind speeds during daytime downdrafts were more intense in the dry season (not reported). These gusts would be sufficient to topple 73% of the previously investigated trees at our study site, which emphasize the importance of wind disturbances on controlling forest structure and diversity in central Amazon.

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Estimation of the nocturnal boundary layer height over the central amazon forest using turbulence measurements

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Abstract

The nocturnal boundary layer height (h_N) was investigated using one year of data (2022) collected by sonic anemometers installed at 11 heights, above the canopy top on the towers of the Amazon Tall Tower Observatory (ATTO) in Central Amazon. Unlike previous assessments relying on indirect methodologies, in the present study h_N was directly estimated from measurements of turbulent fluxes of momentum and kinematic sensible heat. Our findings highlighted the dynamic effect of forest topography: under north-east winds, associated with lower roughness, h_N varied between 81 m during very stable stratification and 172 m in neutral conditions. Conversely, under southeast winds, where roughness is higher, h_N ranged between 81 and 223 m. These estimates reveal the significant control exerted by atmospheric stability and topography on the h_N variability. Interestingly, under neutral and weakly stable stratifications our finds align with the theoretical parameterization proposed in previous works. However, discrepancies emerged in very stable stratification and when the boundary layer structure is influenced by topography.

Keywords: Amazon forest, nocturnal boundary layer, turbulent fluxes, topographic effects, atmospheric stability

2.1 Introduction

The height of the atmospheric boundary layer (BL) exhibits dynamic variations throughout the diurnal cycle, significantly impacting weather patterns, air quality, and energy exchange between the surface and the atmosphere (Driedonks, 1982; Kotthaus and Grimmond, 2018; Levi et al., 2020). While the top of the BL during the daytime is well-defined as the height of an inversion layer (Hennemuth and Lammert, 2006), the height of the Nocturnal Boundary Layer (NBL) is not so easily identified due to its characteristics of weak vertical mixing, low and intermittent turbulence, and stable atmospheric conditions.

Fundamentally, the height of the BL represents the vertical distance from the surface at which turbulence remains active, being therefore defined by turbulence-based relationships. Based on its turbulence properties, the NBL height (h_N) can be identified at the level at which the turbulence intensity sharply decreases with respect to its surface value (Lenschow et al., 1988; Stull, 1988; Vickers and Mahrt, 2004). However, such determination of h_N requires measurements of turbulent quantities with high vertical resolution, which are usually difficult to obtain.

Observations obtained by aircraft during field campaigns have provided high-frequency turbulence measurements for estimating the height of the atmospheric boundary layer over different surface types (Poulos et al., 2002; Dai et al., 2014). Additionally, previous studies have proposed analytical models to characterize turbulent quantities and estimate NBL height (Caughey et al., 1979; Lenschow et al., 1988). These models assume a constant geostrophic wind with height and neglect drainage flow effects associated with terrain slope. As a result, they do not account for how variations in topography might modify wind patterns, which may lead to inaccuracies in predicting how heat and momentum are transported vertically. Despite these simplifications, Lenschow et al. (1988) incorporated radiative cooling effects into their model, a key process in NBL dynamics, providing a robust theoretical basis for NBL height estimation.

In cases where turbulent quantities are unavailable, different parameters measured with different instruments (and resolutions) have been widely employed. The methods often include analysis of mean profiles of temperature, humidity, wind speeds obtained from radiosondes (Seibert et al., 2000; Fisch et al., 2004; Carneiro and Fisch, 2020, among others), tethered balloons (Garstang et al., 1990; Culf et al.,

1997; Acevedo et al., 2004, among others) and ground-based remote sensing such as ceilometer, Sodar and Radiometer (Eresmaa et al., 2006; van der Kamp and McKendry, 2010; Dias-Júnior et al., 2022; Souza et al., 2023, among others). These estimates are considered indirect because they are based on characteristics of the mean profile that are influenced by the vertical extent of turbulence (Mahrt et al., 1979; Vickers and Mahrt, 2004).

The estimates presented in this work are based on direct measurements of turbulent characteristics, made possible by the multi-level high frequency measurements of the Amazon Tall Tower Observatory (ATTO) (Andreae et al., 2015; Dias-Júnior et al., 2019). Two ATTO towers (325-m and 80-m heights) instrumented with sensors installed at different heights enable flux measurements across the NBL. We aim to determine h_N by analyzing vertical profiles of the kinematic sensible heat and momentum turbulent fluxes measured above a pristine forest during 2022. Additionally, we investigate how atmospheric stability and surface roughness (influenced by the interaction between wind direction and topography) affect NBL height. Specifically, we seek to answer the following questions:

1. What is the height of the NBL based on turbulence properties?
2. How do atmospheric stability and surface roughness influence NBL height?

By addressing these questions, we intend to contribute to a deeper understanding of NBL dynamics above the Amazon forest. These estimates hold significant relevance for future research, particularly for studies conducted at the ATTO site, such as those examining aerosol and greenhouse gas concentrations in the interface between the forest and the atmosphere (Botia et al., 2020; Yanez-Serrano et al., 2020; Barbosa et al., 2022).

2.2 Study site and data acquisition

Our data were measured at the ATTO site (<https://www.attoproject.org/>) located about 150 km northeast of Manaus, in Central Amazonia (Andreae et al., 2015). The site is in a region of pristine tropical forest where the average canopy height is around $h = 35$ m (Gomes Alves et al., 2023). Sonic anemometers were installed on two towers (located at 2.1468 °S, 59.0068 °W and 2.1448 °S, 59.0008 °W, respectively) that are 670 m distant from each other (Dias-Júnior et al., 2019).

The 80-m tower (named as INSTANT) and the 325-m tower (named as TALL) are situated on a plateau approximately 130 m above sea level, with an average slope of $H_h/(2L_h) \approx 0.06$, where L_h represents the horizontal half-length of the topography and H_h denotes the height relative to the surrounding terrain (Chamecki et al., 2020). A comparative analysis between measurements taken at similar heights (81 m at INSTANT and 83 m at TALL towers) showed negligible differences in wind speed and temperature data (Figure A1 and A2 in the Supplementary Material, included here as Appendix A), indicating that the data from both towers are spatially consistent with each other and can be reliably used together at the same profile.

The data were acquired using Campbell Scientific Instruments CSAT-3B (CSAT3, Logan, USA), and Thies Ultrasonic Anemometer 3D (THIES, Gottingen, Germany). The measurement heights in meters above ground were T-035, C-050, C-081 (measured at the INSTANT tower) and T-83, T-100, T-127, T-151, T-172, T-223, T-247, T-274 and T-298 (obtained at the TALL tower), where C stands for the CSAT3, and T for the THIES. Data from the three wind velocity components (u , v and w) and temperature measured at a sampling frequency of 10 Hz were analyzed. Additionally, temperature data from thermohygrometers (CS215-L, Campbell Scientific Inc.), sampled at 0.1 Hz were obtained at the same heights as the anemometers. Data measured from 8:00 pm to 5:00 am LT (Local Time = UTC - 4h) were analyzed to avoid transitional periods. Surface topography (Figure 2-2) at the experimental site was assessed using geomorphometric variables derived from SRTM (Shuttle Radar Topographic Mission) data available on <https://earthexplorer.usgs.gov/>.⁹⁶

2.3 Methodology

2.3.1 Quality Control and Data Processing

The raw data were initially divided into 30-minute subsets (18000 records) for quality control (QC) and tilt correction, aligning with practices in other micrometeorology studies. The QC procedures followed established protocols (Vickers and Mahrt, 1997; Zahn et al., 2016; Dias et al., 2023) and were applied to each wind velocity component and the anemometric temperature. These procedures involve checking the completeness of the data record, identifying error flags, identification and removal of spikes and the application of stationarity test. The wind

velocity components were rotated into the streamwise coordinate system (double rotation method) following Foken (2017), ensuring that the mean vertical wind component was negligible, while the horizontal components were aligned with the mean wind direction. Following QC and tilt correction, turbulence statistics were calculated in 5-minute time windows. The QC procedures focused on data preparation, ensuring the reliability of the data before proceeding to the calculation of turbulence statistics in subsequent steps. The protocols established at the ATTO site are described in detail by Zahn et al. (2016) and Dias et al. (2023).

Periods that indicate the presence of low-level jets (LLJ) were excluded from the analysis. Depending on the height and magnitude, LLJ can significantly influence surface turbulence and the vertical distribution of mass and energy (Banta, 2008; Corrêa et al., 2021). In our analysis, LLJs were identified as the lowest wind speed maximum (or jet nose) in the vertical profile characterized by a decrease in wind speed above and below this height, following criteria proposed by Andreas et al. (2000). However, while Andreas et al. (2000) required a wind speed decrease of at least 2 ms^{-1} both above and below the jet nose, we opted not to impose this threshold, as it led to some LLJ profiles not being identified. Instead, we required a decrease in wind speed at least three measurement heights above the jet nose, without imposing any specific speed threshold above, below, or at the jet nose. We analyzed hourly average wind speed profiles to identify the occurrence of LLJs, and about 13% of the 8905 profiles were excluded.

The potential temperature (θ) was calculated using the equation: $\theta = T_h + (g/C_p)h_r$ (Monson and Baldocchi, 2014), where T_h represents the temperature measured by the thermohygrometer (units of K), g is the gravitational acceleration (9.81 ms^{-2}), C_p is the specific heat of dry air at constant pressure ($\text{J kg}^{-1} \text{ K}^{-1}$), and h_r represents the height (in meters) at which the temperature (T_h) was measured. This equation provides an approximation of the θ based on the dry-air adiabatic lapse rate (g/C_p), which is appropriate for near-surface atmospheric layers where variations in pressure and air density with height are minimal (Monson and Baldocchi, 2014).

2.3.2 Nocturnal Boundary Layer Height Estimation

The h_N values were determined from vertical profiles of kinematic sensible heat and momentum fluxes, where h_N was defined as the lowest height where the turbulent

flux dropped below 5% of its surface values (Caughey et al., 1979; Lenschow et al., 1988). The turbulent fluxes were estimated by the covariance of the fluctuations of the vertical wind component (w) with sonic temperature for kinematic sensible heat and with the horizontal wind component (u) for momentum fluxes. All variables, including fluxes, were evaluated over a five-minute time window (Vickers and Mahrt, 2006).

In addition, the NBL height was estimated using the model proposed by Lenschow et al. (1988). Although turbulent measurements offer direct information on NBL dynamics, the Lenschow model provides a theoretical basis for estimating NBL height. Lenschow et al. (1988) suggested that the turbulent fluxes profiles can be parameterized as $(1 - z/h_N)^r$, where r is 1.5 and 1.75 for heat and momentum flux, respectively. In our analysis, we assume z to be equivalent to $z - h_c$, where h_c denotes the height near the canopy top with the highest intensity of turbulent flux. This adaptation was made due to the presence of the tall canopy (forest) at our study site. Subsequently, we compared the adapted model (referred to as Lenschow's model) with data measured at our study site and estimated the NBL height using both observational data and Lenschow's model outputs.

2.3.3 Atmospheric Stability and Surface Roughness

Two main factors were considered for estimating the NBL height: atmospheric stability and surface roughness. Atmospheric stability was evaluated using the stability parameter (h/L), where L represents the Monin-Obukhov length, defined as $L = -u_* / ((kg/T) \overline{w'T'})$, with u_* being the friction velocity, κ the von Kármán constant (0.41), g the gravitational acceleration, T the sonic temperature and $\overline{w'T'}$ the kinematic sensible heat flux. The variables u_* , $\overline{w'T'}$ and T were defined at 35 m above the ground.

The stability classes were empirically defined based on the methodology proposed by Mahrt (1998) and Cava et al. (2022), which involves analyzing the relationship between the kinematic sensible heat flux ($\overline{w'T'}$) and momentum flux ($\overline{u'w'}$) as a function of the stability parameter (h/L). This approach allowed for the identification of specific h/L thresholds that define the stability classes, based on observed changes in flux behavior from measurements at 35, 50, 81, 172, and 298 m (discussed in Section 4.1). Therefore, the identified thresholds reflect characteristics specific to the local atmospheric stability conditions at our study site.

The second factor, surface roughness, was assessed using the roughness length (z_0). This parameter was derived using the traditional logarithmic law (Kaimal and Finnigan, 1994), which describes the wind speed $V(z)$ at height z above tall vegetation as $V(z) = (u_*/\kappa)\ln((z-d)/z_0)$, where d represents the displacement height. The wind speeds measured at six different heights (35, 50, 81, 100, 127, and 151 m above ground) were logarithmically interpolated assuming $d = 0.75h$ (Dias-Júnior et al., 2015).

To calculate z_0 , only data from nights with near-neutral stability conditions were used. This approach minimizes the effects of atmospheric stratification, ensuring that z_0 primarily represents the roughness of the surface. Roughness classes were then defined based on the variability of z_0 (detailed in Section 4.1).

2.3.4 Statistical Analyses

A factorial Analysis of Variance (ANOVA) was conducted to evaluate the mean differences among the classes: atmospheric stability, roughness, and measurement height. Specifically examining how these factors influenced the variability in wind speed and potential temperature. The analysis was performed on the residuals obtained from an AutoRegressive Integrated Moving Average (ARIMA) model, which allowed us to account for the autocorrelation inherent in time series data. The normality of the distributions was assessed using the Shapiro–Wilk test. Subsequently, Tukey tests were used to analyze interactions among these factors, with a significance level of 5%. To ensure that extreme values do not affect the analyses, outliers were identified and removed using the Tukey et al. (1977) method. We used the $\overline{w'T'}$ flux as a reference to identify and remove outliers at each measurement height.

2.4 Results and Discussion

2.4.1 Data Classification for NBL height Estimation

Figure 2-1 illustrates the variations of $\overline{w'T'}$ and $\overline{u'w'}$ at five heights as a function of h/L evaluated at the canopy top. The boundaries that define the stability classes, based on h/L , are represented by black dashed lines. Three main stability regimes were identified as follows:

i) Near-Neutral regime (NN: $0 \leq h/L < 0.15$): This regime is characterized by maximum $|\overline{u'w'}|$ flux near the canopy top and small magnitudes of $|\overline{w'T'}|$ flux (Figure 2-1, grey-shaded area). Turbulence near the ground is produced primarily by wind shears (Stull, 1988) and the $|\overline{w'T'}|$ flux decreases due to reduced temperature fluctuations, as the vertical thermal gradient is weak.

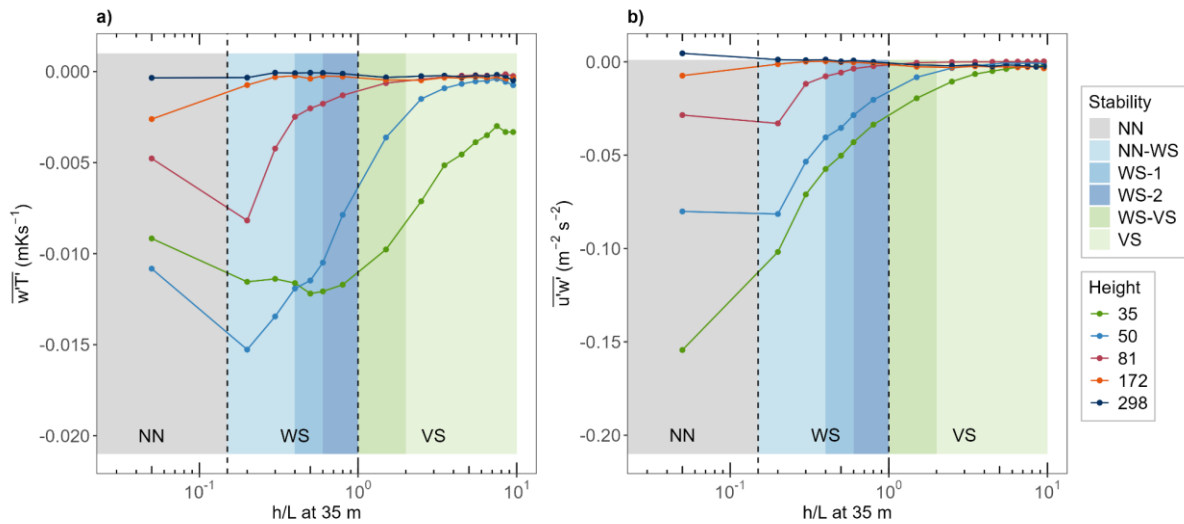


Figure 2-1. Average values of the (a) kinematic sensible heat ($\overline{w'T'}$) and (b) momentum ($\overline{u'w'}$) fluxes at different heights as a function of the stability parameter (h/L) evaluated at 35 m. Background color marks six classes of atmospheric stability.

ii) Weakly Stable regime (WS: $0.15 \leq h/L < 1$): In this regime, the $|\overline{w'T'}|$ flux near the canopy top increases, while the $|\overline{u'w'}|$ flux gradually decreases with increasing h/L (Figure 2-1, blue-shaded area). As the vertical temperature gradient becomes more pronounced, temperature fluctuations intensify, leading to an increase in $|\overline{w'T'}|$ flux.

iii) Very Stable regime (VS: $1 \leq h/L < 10$): This regime is characterized by weak or intermittent turbulence and a strong thermal inversion (Mahrt, 1998). The strong stratification inhibits vertical mixing of heat, causing the $|\overline{w'T'}|$ flux to decrease. Similarly, the $|\overline{u'w'}|$ flux becomes small as vertical momentum transport is suppressed (Figure 2-1, green-shaded area). Additionally, this regime is marked by the coexistence of turbulent structures dominated by submeso motions (Mortarini et al., 2019).

Additionally, to further investigate the variability of h_N within narrower stability ranges, we empirically subdivided the Weakly Stable (WS) regime into NN-WS ($0.15 \leq h/L < 0.4$), WS-1 ($0.4 \leq h/L < 0.6$), and WS-2 ($0.6 \leq h/L < 1$), and the Very Stable (VS) regime into WS-VS ($1 \leq h/L < 2$) and VS ($2 \leq h/L < 10$), totaling six stability classes (Table 2-1). While the three main stability classes effectively captured the general trends, these subdivisions were introduced to better represent subtle transitions in flux behavior within each stability regime. By refining these categories, we aimed to highlight the gradual decrease in h_N with increasing stability, as illustrated in Figure 2-3 and Figure 2-7.

Table 2-1. Classes of atmospheric stability: Near-Neutral (NN), Weakly Stable (WS) and Very Stable (VS) regimes. Percentage of occurrences and median (IQR, Interquartile Range) of sensible heat flux ($\overline{w'T'}$) and momentum flux ($\overline{u'w'}$) measured at 35 m for each stability class.

Atmospheric	Percentage	$\overline{w'T'}$ (Wm^{-2}) Median (IQR)	$\overline{u'w'} \times 10^{-2}$ (Nm^{-2}) Median (IQR)
NN	7.2%	-12.8 (16.2)	-15.2 (15.5)
Transition NN-WS	10.6%	-14.3 (15.0)	-8.3 (7.0)
WS-1	8.9%	-15.0 (15.8)	-5.7 (4.3)
WS-2	14.4%	-14.4 (15.3)	-4.1 (3.0)
Transition WS-VS	25.9%	-12.0 (13.3)	-2.3 (2.0)
VS	33%	-6.2 (8.2)	-0.7 (0.8)

Although atmospheric stability was not influenced by wind direction, we observed that roughness surface varied across sectors with distinct topographic features, influenced by the interaction of atmospheric flow with these features. This relationship is illustrated in Figure 2-2b, which shows a bivariate polar plot of mean roughness length (z_0) classified by horizontal wind speed (V) and wind direction at 100 m height. While it is challenging to isolate the specific contribution of topography in a complex environment like the Amazon rainforest, our analysis shows that sectors with higher z_0 values (Figure 2-2b) correspond to regions where the terrain elevation (Figure 2-2a) probably increases surface roughness. These findings align with the hypothesis outlined by Corrêa et al. (2021) that south-southeast flow, when interacting

with the extensive topographic gradient, creates a circulation mechanism that can be detected at the ATTO towers. Chamecki et al. (2020) used experimental data and modeling to show that the vertical structure of turbulence above the Amazon forest is strongly influenced by two distinct regions within the roughness sublayer: one associated with the forest and the other with topography. While both factors modulate turbulence and influence z_0 , the variability in z_0 may primarily be driven by different topographic features, as no significant variability in Leaf Area Density (LAD) was observed across the transects around the ATTO towers (Gomes Alves et al., 2023, their Figure 2-2b).

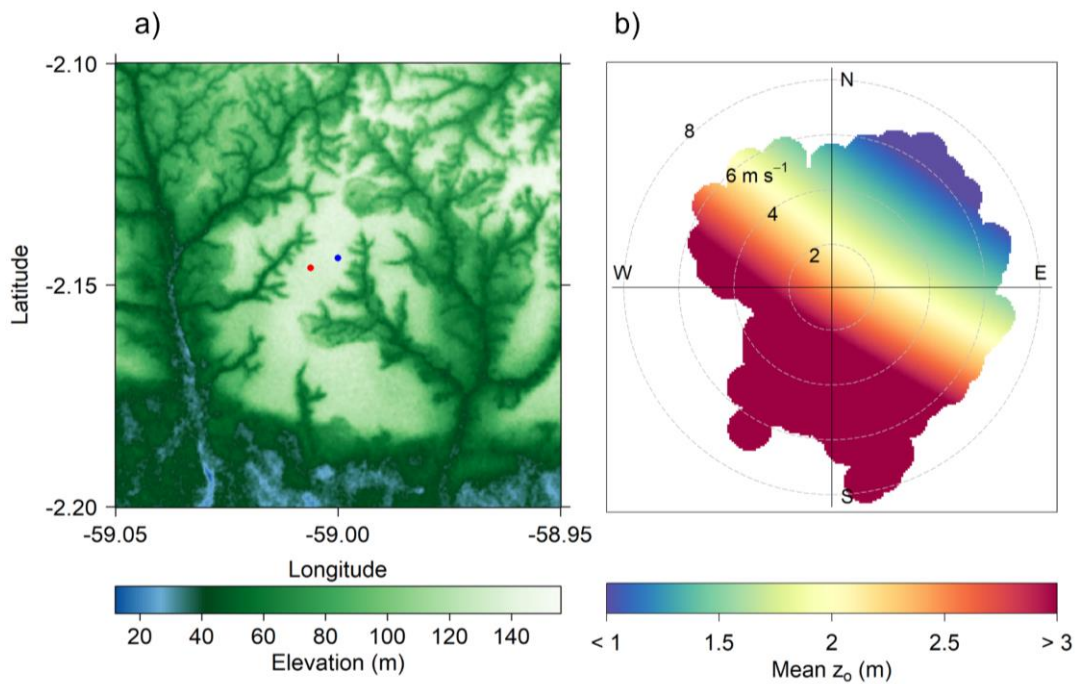


Figure 2-2. (a) Topography map around the ATTO site showing the locations of the 325-m tower (red dot) and 80-m tower (blue dot). (b) Polar plot illustrating the mean roughness length (z_0) as a function of the wind speed (radial scale, m s^{-1}) and direction (angle) at 100 m.

Understanding the variability of z_0 is crucial because it provides insights into how surface characteristics affect the turbulence, consequently influencing the development of the boundary layer during the nighttime period. Given this, in addition to the stability classes, our data were also categorized based on z_0 values and their variation across different wind direction sectors. First, we focused on wind directions from 0 to 180 degrees due to the predominance of easterly winds in the region

(Andreae et al., 2015). This range was chosen to minimize potential flow distortion caused by the tower structure, particularly from the west. Consequently, wind directions from 180 to 360 degrees were excluded to avoid such distortions.

Wind directions from 0 to 70 degrees were classified as “lower roughness” ($z_0 \approx 1$), while those from 90 to 180 degrees were considered “higher roughness” ($z_0 > 2$). We included both $z_0 \approx 2$ and $z_0 \approx 3$ in the high roughness category to ensure a sufficient amount of data, enabling a more robust analysis under these conditions given the predominance of easterly winds. Furthermore, wind directions between 70 and 90 degrees were excluded to ensure differentiation between the two classes.

2.4.2 Influence of topography and atmospheric stability on vertical profiles

The vertical profiles of $\overline{w'T'}$ and $\overline{u'w'}$ are presented for two distinct roughness classes in Figure 2-3. These profiles were normalized relative to the flux near the top of the canopy (35 m). Our results establish that the fluxes are larger near the canopy top, decreasing with height. Specifically, under NN conditions, the $\overline{w'T'}$ fluxes reached 5% of the surface value at a height of 172 m for lower-roughness (Figure 2-3a, solid red line) and 223 m under higher-roughness (Figure 2-3a, dashed red line). This observation suggests that the $\overline{w'T'}$ profiles changed influenced by distinct roughness length inherent to the topography. As stability increased, the $\overline{w'T'}$ fluxes reached 5% of its surface value at progressively lower heights, approaching 81 m in the VS regime for both types of roughness (Figure 2-3f, red lines).

Similar behavior was observed in the $\overline{u'w'}$ fluxes (Figure 2-3, blue lines). However, these fluxes decayed to 5% of their surface values at heights comparatively lower than those observed for the $\overline{w'T'}$ fluxes. This pattern can be explained by the vertical wind profile becoming constant at a specific height, while the potential temperature continues to increase. For example, in the NN regime under lower roughness (Figure 2-4c), the wind speed reached a maximum at 223 m, while the potential temperature gradient remained positive at this height. As a result, the $\overline{u'w'}$ flux approached zero, while the $\overline{w'T'}$ flux was still positive. Vickers and Mahrt (2004) reported that estimates of h_N based on the $\overline{u'w'}$ profile are considered less reliable than $\overline{w'T'}$ profile, since any layer with turbulence in communication with the

surface should experience significant cooling, even if the momentum flux becomes small or not.

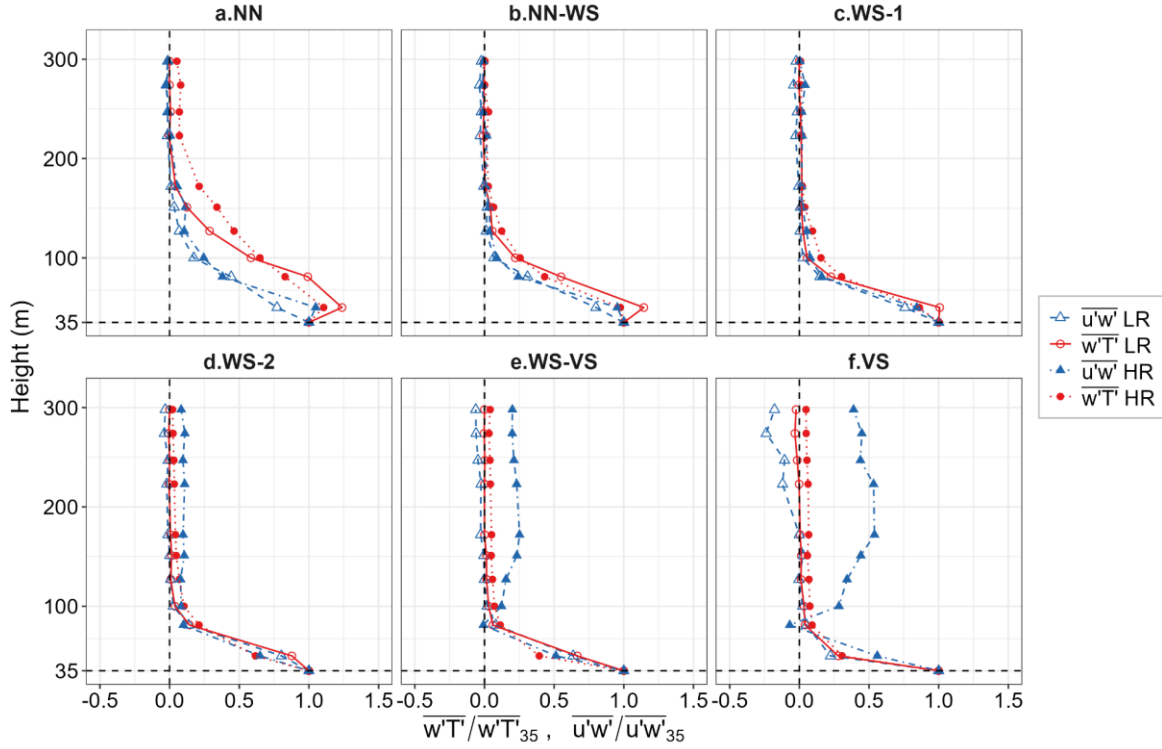


Figure 2-3. Average vertical profiles of normalized kinematic heat flux ($\overline{w'T'}$, red lines) and momentum flux ($\overline{u'w'}$, blue lines) for different stability classes. Solid lines illustrate profiles at lower roughness (LR), while dashed lines represent those at higher roughness (HR).

Especially, within the WS-2 (Figure 2-3d), WS-VS (Figure 2-3e), and VS (Figure 2-3f) regimes, where weak turbulent mixing prevails, the momentum fluxes are larger near the canopy top. $\overline{u'w'}$ initially decrease with height, and then remain constant or even increase with height when winds from the higher-roughness direction prevailed. This can possibly occur when the main source of turbulence is shear generation associated with non-turbulent fluctuations referred to as submeso motions (Mahrt, 2009). In the Amazon, these motions significantly impact flow dynamics under stable conditions (Acevedo et al., 2014; Oliveira et al., 2018; Corrêa et al., 2021; Mortarini et al., 2022) and tend to propagate to lower levels as stability increases (Cava et al., 2022; Mendonça et al., 2023). It is important to note that submeso activity is frequently associated with low wind speeds, but this is because the majority of the studies were conducted close to the surface. Figure 2 in Mortarini et al. (2019) clearly shows that

submeso activity at higher heights can exist at wind speeds of the order of 5 ms^{-1} or larger. Furthermore, Figure 7 in Cava et al. (2022) confirms the presence of submeso activity at the upper levels of the ATTO tower for atmospheric stability: neutral, weakly stable and very stable stratification.

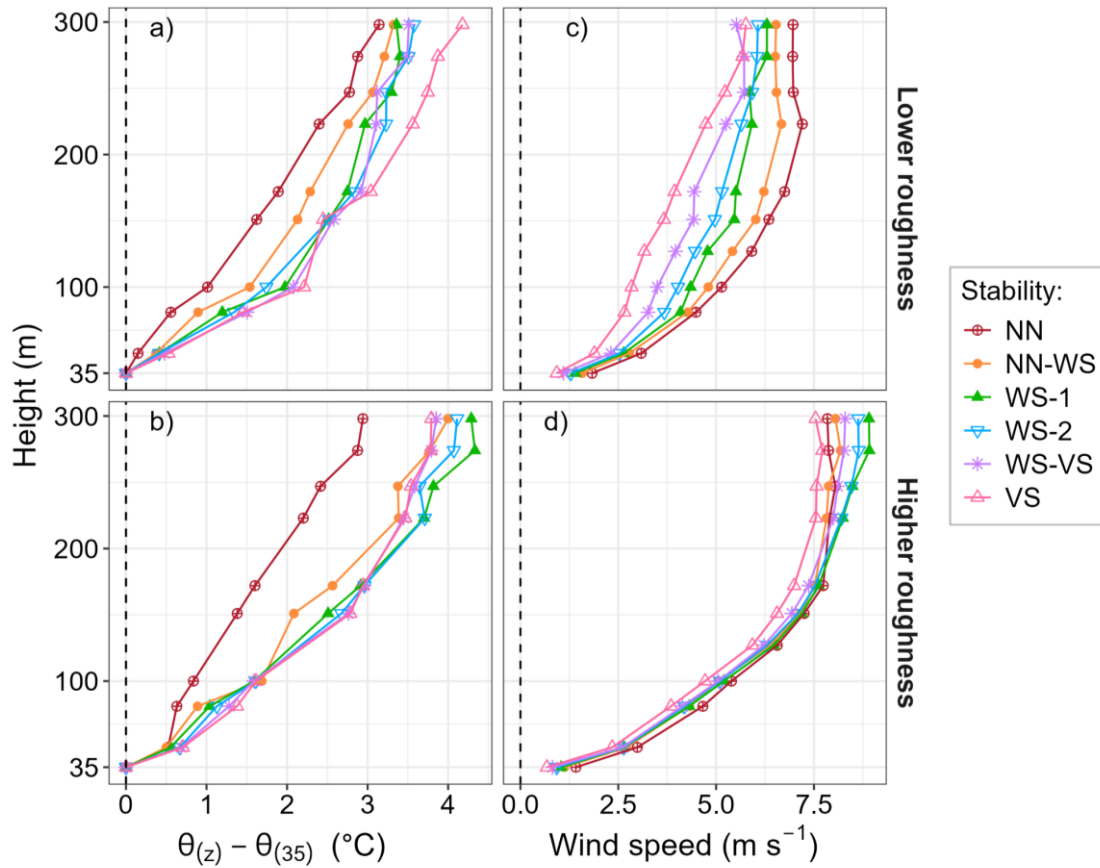


Figure 2-4. Vertical profiles of potential temperature (θ) difference from the value observed at 35 m (a)(b) and horizontal wind speed (c)(d) under lower (top panels) and higher (bottom panels) roughness. Lines colors show different stability classes.

Figure 2-4 presents a comparative analysis of roughness classes, illustrating the six stability regimes for vertical profiles of potential temperature (θ) differences between the value observed at 35 m ($\theta = \theta_z - \theta_{35}$) and the wind speed (V). The temperature increases with height across all measurement levels of the tower, indicating the presence of a stable layer extending above the tower independent of the type of roughness and stability classes. However, turbulent profiles approached zero at the lower measurement levels of the tower (Figure 2-3), revealing that the turbulence decreased systematically below the inversion layer. This result highlights

that the height through which surface-based turbulence extends is not associated with the height of the thermal inversion layer. This implies that the sole use of potential temperature profiles is likely misleading for the determination of the h_N above the canopy top of Amazon forest.

Analyzing the vertical profiles of V (Figure 2-4), we found conformity with characteristics of canopy turbulence, as described by Finnigan (2000). Results from subsequent ANOVA analysis revealed that when the wind direction was from sectors with lower roughness, wind speeds under VS conditions were lower compared to NN conditions (Figure 2-4c). For example, at a height of 151 m the wind speed difference was 2.7 ms^{-1} ($p < 0.01$), indicating a significant reduction under VS regime. Conversely, under higher roughness, the average difference was smaller (0.7 ms^{-1} , $p < 0.01$) (Figure 2-4d). This discrepancy suggests a stronger influence of stability on speeds when winds came from directions with lower roughness, possibly because the reduced “surface” friction in these areas allows stability effects to more directly impact wind speeds.

Analysis within the same stability class showed that when the wind direction was from the higher-roughness area, the values of V exceeded those under lower roughness starting from 151 m, with an increase of approximately 1.0 ms^{-1} under NN ($p < 0.01$) and 2.9 ms^{-1} under the VS ($p < 0.01$) regime (Figure 2-4c,d). These higher wind speeds resulted in θ values being closer together across different stability classes (Figure 2-4b).

2.4.3 Turbulent vertical profiles using an analytical function

The comparison between our observational data (smoothed line) and the results from the adapted Lenschow model (dashed line) is shown in Figure 2-5 and Figure 2-6. To the right of the vertical dashed line (i.e., within the boundary layer), the observational data closely align with the model predictions. Conversely, to the left of the line, the model deviates from the data, showing lower values.

In regions characterized by lower roughness (Figure 2-5), the observational data align well with Lenschow’s model, except in the VS regime. This deviation may be attributed to a reduced turbulence or the relative limited data used for model fitting. These findings are consistent with the observations of Lenschow et al. (1988). In contrast, under higher roughness (Figure 2-6), Lenschow’s model did not perform

consistently for WS-2, WS-VS and VS conditions. This result reveals a notable disparity between the observational data and the vertical structure predicted by the model. In these conditions, the trend of smoothed lines changes. Initially, within the NN regime, both the observational and model profiles exhibit a gradual decrease in flux with increasing height. However, as stability increases, observed flux drops more sharply with height, reflecting the effect of stable stratification, while the modeled profile shows limited sensitivity to these effects.

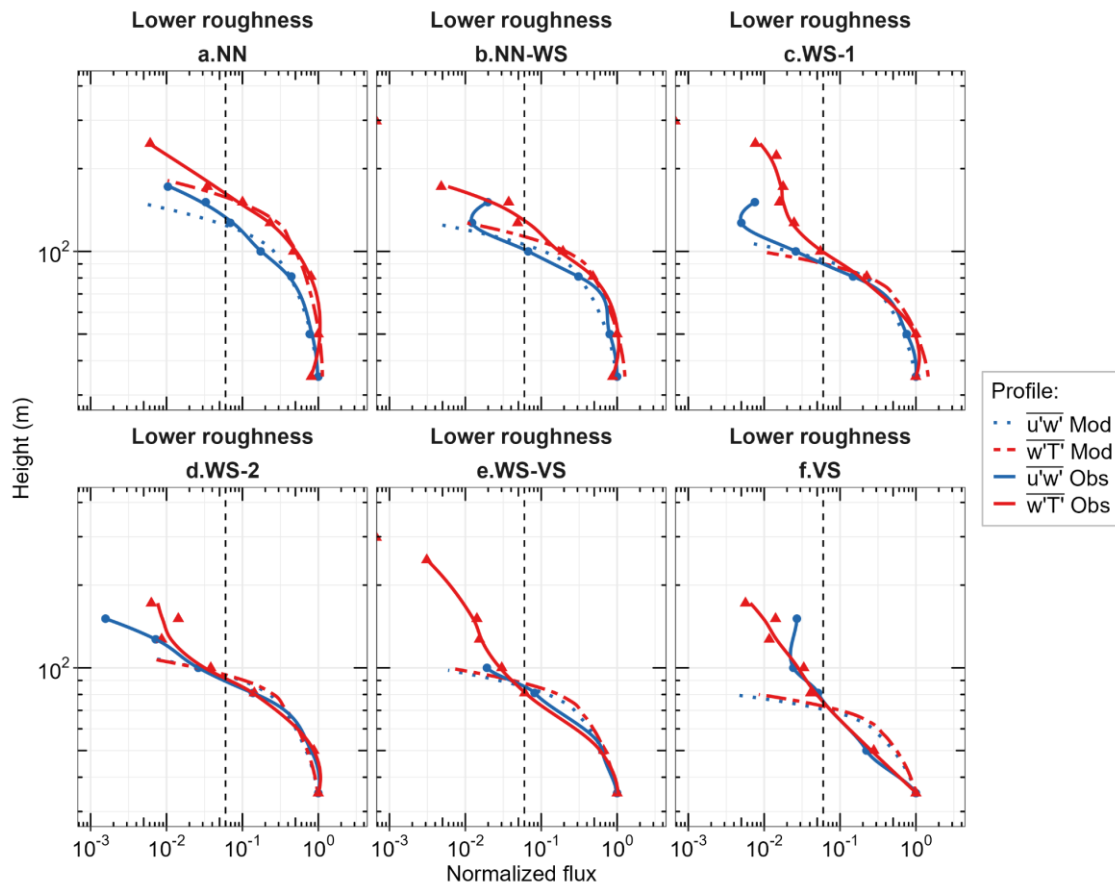


Figure 2-5. Normalized vertical profiles of kinematic sensible heat flux ($\overline{w'T'}$, red lines) and momentum flux ($\overline{u'w'}$, blue lines) in different stability classes based on lower roughness. The dashed line represents the Lenschow model-fitted flux profile. Dots depict observed data, while the smoothed line shows the overall trend. Vertical dashed lines correspond to 5% of the maximum flux.

Focusing specifically on vertical profiles of $\overline{u'w'}$ in areas with higher roughness (Figure 2-6, blue lines), within WS-2, WS-VS and VS regimes, Lenschow's model does not perform consistently due to intrinsic difficulties of defining a boundary layer when the shear generation is associated with submeso motions as discussed earlier

(Figure 2-3, section 4.2). The significance of these motions becomes more pronounced in very stable conditions, mainly due to reduced turbulence.

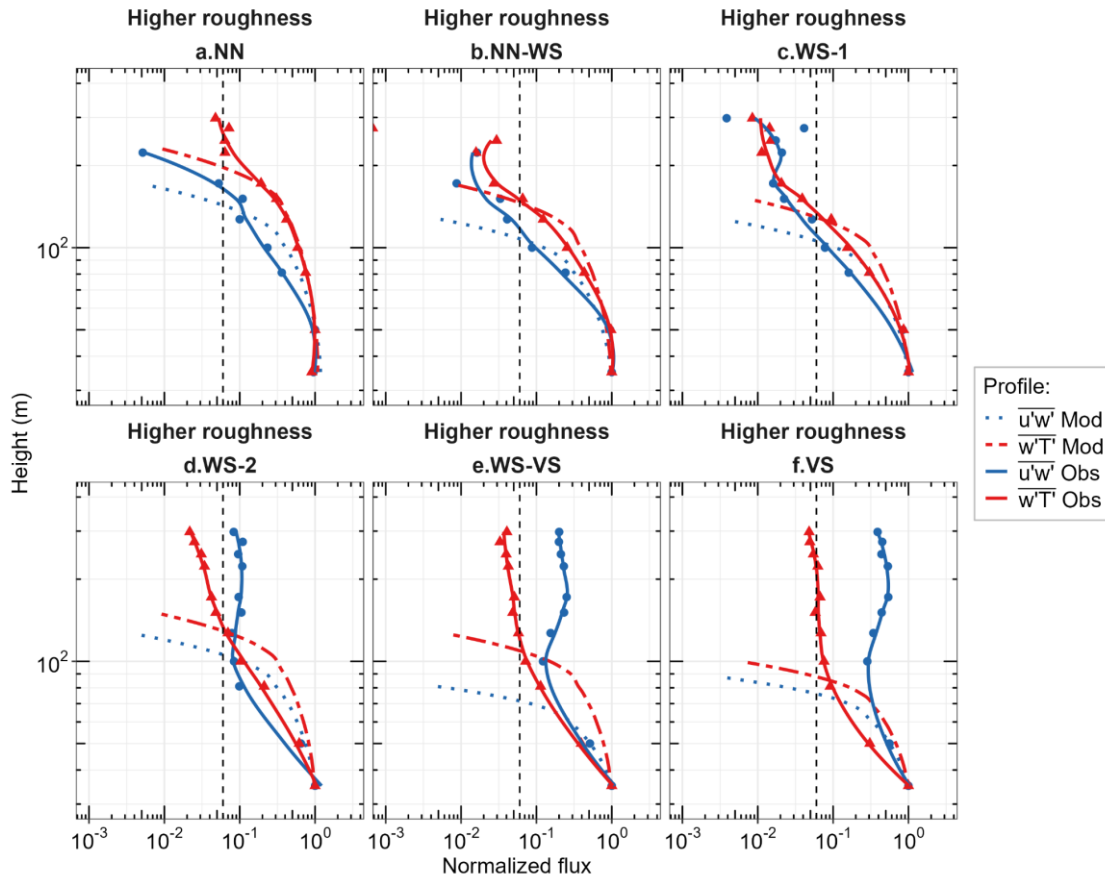


Figure 2-6. Same as Figure 2-5 but for higher roughness.

At the ATTO site, the wind profile (Figure A3 in the Supplementary Material) exhibits a pattern similar to that described by Lenschow et al. (1988) (Figure 13 in their paper), despite differences in magnitude. However, assumptions in Lenschow's model, such as a constant geostrophic wind with height and the neglect of drainage flow effects, can introduce inaccuracies in predicting the vertical transport of heat and momentum. Although the model was developed for temperate-zone latitudes, where atmospheric characteristics differ from those in tropical regions, our findings indicate that it performed well in representing the NBL height at the ATTO site, particularly under lower roughness. In this case, discrepancies were observed only during very stable stratification, consistent with the results reported by Lenschow et al. (1988). However, under higher roughness, divergences occurred even under weakly stable stratification, likely related to local topographic effects. While the model is not suitable

for very stable conditions, our findings show that Lenschow's model can be applicable to tropical conditions, with limitations under high roughness.

2.4.4 Estimates of NBL height

The NBL height estimates were compared across the stability and roughness classes (Figure 2-7), considering the method of estimation ($\overline{w'T'}$ or $\overline{u'w'}$ profiles). The results presented here focus on h_N estimates derived from observational data due to the limitations of the adapted Lenschow model described in the previous section. The subscripts "wT" and "uw" represent the h_N values derived from kinematic sensible heat and momentum flux vertical profiles, respectively. This distinction holds throughout the subsequent discussion.

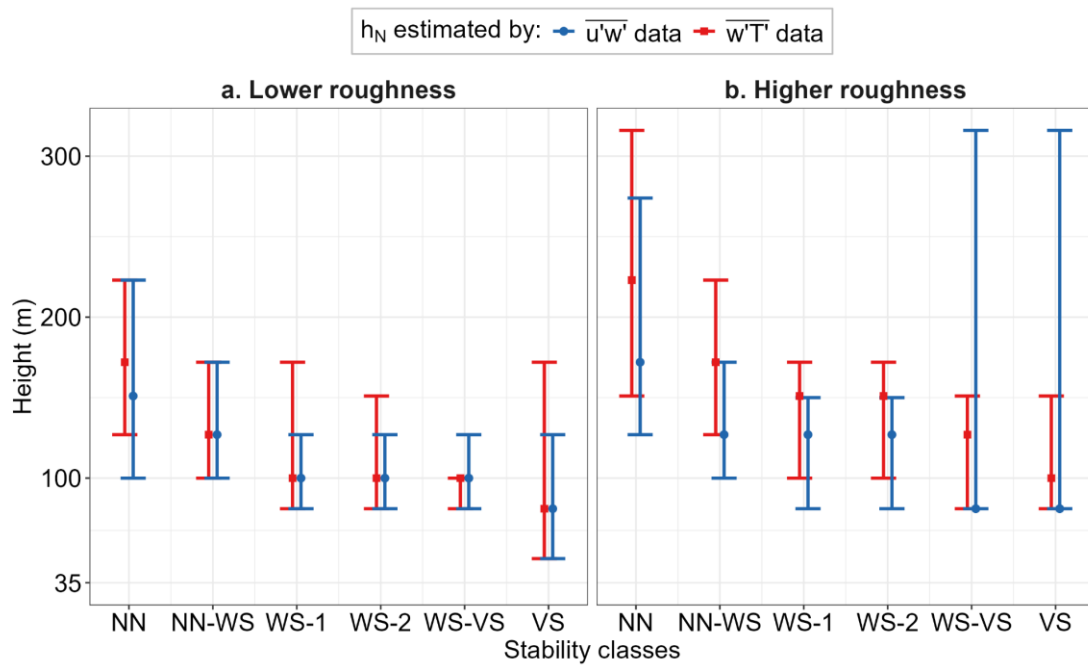


Figure 2-7. Median and interquartile range of the NBL height (h_N) from vertical profiles of kinematic sensible heat flux ($\overline{w'T'}$, in red) and momentum flux ($\overline{u'w'}$, in blue) as a function of the atmospheric stability classes under (a) lower and (b) higher roughness conditions.

When investigating the types of turbulent profiles employed, h_{N_uw} (blue circles in Figure 2-7) exhibited lower values than h_{N_wT} (red circles in Figure 2-7), since the $\overline{u'w'}$ flux approached zero, while the $\overline{w'T'}$ flux remained positive, as discussed previously (Section 4.2). The difference was more pronounced in NN, NN-WS, and

WS-1 conditions under higher roughness (24 to 51 m), in contrast to lower roughness conditions (21 m). Particularly under higher roughness, h_{N_UW} was 81 m with the third quartile exceeding 300 m during WS-VS and VS regimes (blue circles in Figure 2-7b) indicating that h_{N_UW} estimates are less reliable than h_{N_WT} . This discrepancy emerged because the $\overline{u'w'}$ profiles were more sensitive to non-turbulent fluctuations.

Focusing specifically on results for h_{N_WT} our analysis reveals variations in h_N values associated with different roughness and atmospheric stability classes. The h_{N_WT} estimates (red circles in Figure 2-7) ranged from 81 m (VS) to 172 m (NN) under lower roughness and from 81 m (VS) to 223 m (NN) under higher roughness. These findings provide insights into the NBL height variations concerning roughness and atmospheric stability. Specifically, the ATTO site may experience a deeper NBL when winds originate from directions characterized by highest roughness (or 90 to 180 degrees). During NN-regime, the interquartile range was greater under higher roughness, indicating that h_{N_WT} values may exceed the height of the tower. In contrast, under lower roughness regions, h_{N_WT} values remained below 250 m.

The height of the NBL in the Amazon region has been investigated in a limited number of studies (Acevedo et al., 2004; Santos et al., 2007; Neves and Fisch, 2011, 2015; Carneiro and Fisch, 2020). Despite this, indirect estimations derived from radiosonde data, launched during experimental campaigns, have provided insights about h_N . For example, in the southwest Amazon, the h_N was determined to be 172 ± 81 m over a deforested area (Neves and Fisch, 2011). Meanwhile, a h_N of 250 m was observed over a pasture area in the Central Amazon (Carneiro and Fisch, 2020). By contrast, values of h_N derived from different instruments/methods (e.g., wind profiler, sodar, lidar and microwave radiometer profiler) under- or overestimated the h_N when compared with in situ radiosonde measurements (Carneiro and Fisch, 2020).

Our results reveal a pattern in which h_N reaches its maximum under NN conditions and gradually decreases to a minimum under the stable stratifications. This observed consistency holds true for both roughness classes and types of flux profiles used (Figure 2-7). Notably, h_N values are larger with increasing roughness, and this difference is reduced with stronger stability. These findings are in line with a previous study by Chamecki et al. (2020) - that topography at ATTO site plays an important role in the organization of turbulence within the atmospheric boundary layer.

2.5 Conclusions

This study investigated the height of the NBL over the Amazon rainforest at the ATTO site using direct measurements of turbulent momentum and kinematic sensible heat fluxes. Our findings highlight that NBL height varied by 152 ± 71 m due to atmospheric stability and topographic effects. Specifically, the interaction between wind direction and different topographical features affects the height of the NBL, reflecting the influence of distinct roughness length. While estimates from an analytical model developed for rolling terrain align well with our observational data, they exhibited limitations during very stable conditions or when the NBL structure is influenced by topography.

We emphasized the practical significance of understanding the NBL height. In the context of a tropical rainforest, a shallower NBL and reduced mixing conditions facilitate the accumulation of high concentrations of gases and organic compounds within a thin layer above the canopy top. This highlights the relationship between boundary layer characteristics and their potential impact on atmospheric composition in the Amazon rainforest. Despite the limitations addressed in this study, such estimates would not be possible to obtain using indirect methods like radiosondes or tethered balloons.

Our results provide insights for improving NBL parameterizations in atmospheric models, particularly in regions like the Amazon where data are limited. From an observational perspective, future research could extend these findings by incorporating nights with low-level jets to assess the role of jets in the h_N variability. Additionally, studies using models could further explore the influence of topography at the ATTO site across different spatial scales.

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Is Low-Level Jet height a good approximation for the top of the nocturnal boundary-layer?

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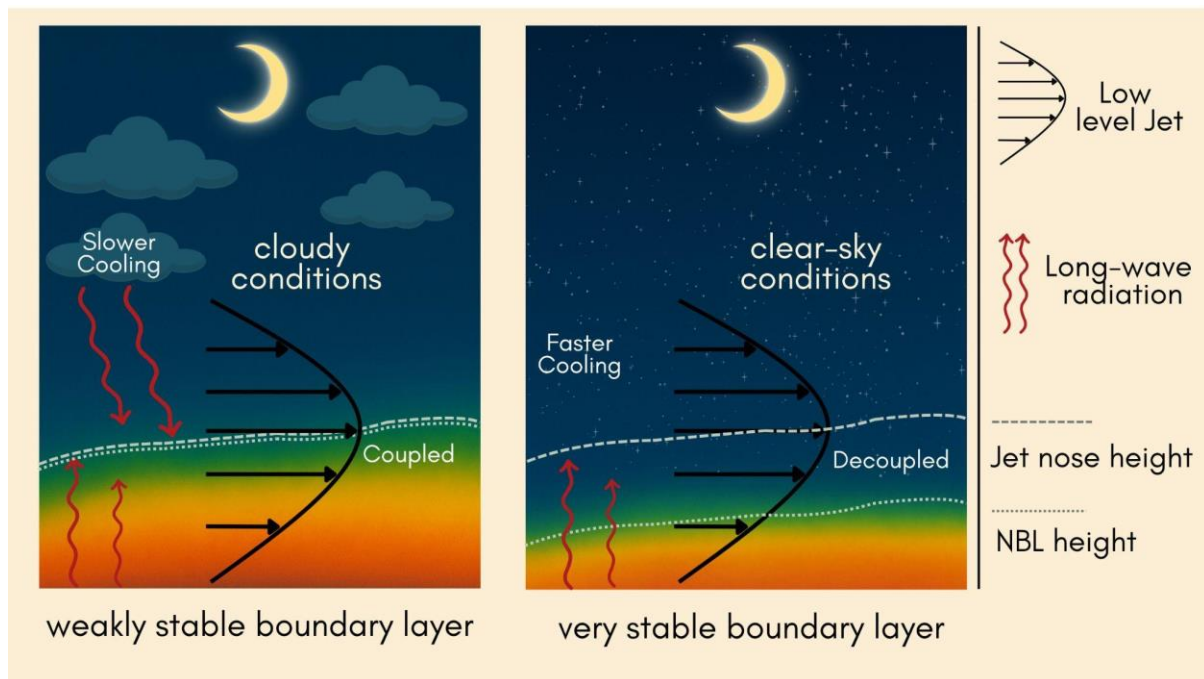
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Abstract

Low-Level Jets (LLJs) influence the dynamics of the Nocturnal Boundary Layer (NBL) by enhancing mechanical turbulence below the jet nose through vertical wind shear. In this study, we have evaluated whether the jet nose height (h_{N_jet}) can serve as a reliable proxy for estimating the NBL height (h_N), defined as the level where turbulence diminishes to negligible magnitudes (h_{N_flux}). We have used unique high-resolution measurements from sonic anemometers installed at 11 heights above the canopy at the Amazon Tall Tower Observatory (ATTO) in the central Amazon during 2022-2023. Nights with LLJ were selected based on the presence of a wind speed maximum (jet nose), along with a decrease in speed for at least two levels above and below. Given that both the height and intensity of LLJs vary over time, h_{N_jet} and h_{N_flux} were determined from mean profiles grouped by jet nose heights: 81–100 m, 151–172 m, and 223–247 m. We have also considered the predominant wind directions associated with LLJs at the ATTO region: N–NE and S–SE. Our results show that, for the same jet nose height, N–NE jets are generally weaker and occur under more turbulent and less stable conditions, with h_{N_jet} being comparable to h_{N_flux} . In contrast, S–SE jets are stronger and associated with enhanced stable thermal stratification, which likely suppresses turbulence and results in h_{N_jet} values higher than h_{N_flux} . These differences are likely influenced by radiative cooling above the canopy, as S–SE jets occurred under larger net radiative loss. We conclude that the LLJ nose identified from wind speed profiles does not provide a good approximation of the NBL height under strongly stratified conditions, when turbulence above the canopy is significantly suppressed.

Keywords: Amazon Forest, Low level Jets, Nocturnal Boundary Layer, Wind shear, Canopy radiative cooling

Graphical Abstract



3.1 Introduction

The Atmospheric Boundary Layer (ABL) is the region of the lower troposphere that is directly influenced by the Earth's surface through turbulent exchange. Its height defines the volume of air where various atmospheric constituents are mixed. Consequently, ABL height is a key parameter for many applications such as weather forecasting, assessment of air quality and atmospheric chemistry, including the dispersion and concentration of trace gases and pollutants (Stull, 1988; Kotthaus and Grimmond, 2018; Levi et al., 2020).

The ABL height changes dynamically throughout the daily cycle. During the day, surface heating leads to the development of a turbulent layer (also called a mixed layer) whose height is well defined by a clear separation between the well-mixed turbulent layer and the inversion layer above it (Hennemuth and Lammert, 2006). In contrast, at night, this separation becomes less distinct as the surface cooling induces stable thermal stratification, which partially suppresses turbulence, so that the ABL is not as effectively mixed as during the day. As a result, the weaker and intermittent nature of turbulence make the determination of the nocturnal boundary layer (NBL) height more challenging than it is at daytime.

As turbulence is the process responsible for transporting the effects of physical interactions that occur at the planet's surface towards larger altitudes, the ABL height can be precisely defined as the vertical distance from the surface at which turbulence remains active. Specifically, the NBL height corresponds to the level where turbulence diminishes to negligible magnitudes or a small percentage of its surface-layer value (Lenschow et al., 1988; Stull, 1988; Vickers and Mahrt, 2004). However, its determination based on such precise definition requires measurements of turbulent quantities with high vertical resolution, which are usually difficult to obtain.

In the absence of direct turbulence observations, alternative methods using data from different instruments (and resolutions) have been proposed, including radiosondes, tethered balloons and ground-based remote sensing instruments such as ceilometers, sodar and radiometers (e.g., Seibert et al., 2000; Acevedo et al., 2004; Carneiro and Fisch, 2020; Dias-Júnior et al., 2022; Souza et al., 2023). The methods often include the analysis of mean profiles of variables such as temperature, humidity, wind speed or aerosol concentration. However, the estimates using these parameters

are indirect rather than direct, because they are only proxies for the amount of turbulence (Mahrt et al., 1979; Vickers and Mahrt, 2004).

As an alternative approach, previous studies have used the height of Low-Level Jets (LLJs) as a key diagnostic for determining the NBL height (Seibert et al., 2000; Vickers and Mahrt, 2004; Pichugina and Banta, 2010; Carneiro and Fisch, 2020, among others). LLJs are characterized by a maximum wind speed (jet nose) in the lower troposphere and play an important role in the NBL dynamics, influencing the transport of moisture, carbon dioxide, ozone, and atmospheric pollutants (Banta, 2008; Karipot et al., 2008; Corrêa et al., 2021). Moreover, the vertical wind shear below the jet nose may generate turbulence, which may be transported downward and affect the exchange of properties between the surface and atmosphere (Prabha et al., 2008; Santana et al., 2015; Dias-Júnior et al., 2017). At the jet nose, turbulence typically diminishes, providing an approximate indication of the NBL top.

Nocturnal LLJs have been documented in many locations around the world. In general, they arise when nocturnal thermal stratification suppresses turbulence, so that the surface decouples from the air above it. Consequently, the drag effects become minimal, allowing the wind speed to accelerate and producing the jet nose. This accelerating process may be influenced by one or a combination of different mechanisms. One of the primary ones causing LLJs is inertial oscillation, as described by Blackadar (1957), when the Coriolis force is the main accelerating process, leading to wind oscillations around the geostrophic wind speed. Other mechanisms include: baroclinicity over sloping terrain, breeze effects, among others (Stull, 1988; de Oliveira and Fitzjarrald, 1993, 1994; Shapiro et al., 2022).

Greco et al. (1992) documented LLJs in the Amazon region, despite the negligible influence of the Coriolis effect and the insignificant baroclinicity at low latitudes. They observed that the LLJs were closely associated with intense thermal stratification near the surface, caused by radiational cooling of the rainforest canopy. This cooling induces turbulent decoupling and reduces frictional drag above the inversion layer, leading to the formation of a nocturnal wind maximum. Furthermore, they found that pressure gradients driven by thermal differences between forest and the river/city, were responsible for much of the nocturnal wind acceleration. Subsequent studies reported that LLJs in the Amazon can also develop in response to mesoscale processes, such as breeze circulation (de Oliveira and Fitzjarrald, 1993, 1994), the propagation of density currents (Dias-Júnior et al., 2017), and convective

systems (Anselmo et al., 2020), highlighting the complexity of LLJ formation in tropical regions. While the mechanisms driving LLJ development remain poorly understood, their occurrence strongly influences NBL dynamics, which are the focus of this study.

Most previous studies have relied on indirect methods to infer NBL dynamics, often without direct turbulence measurements. To fill this gap, the dataset used in this study comprises unique high-resolution measurements from sonic anemometers installed at 11 heights (up to 325 m) above the canopy at the Amazon Tall Tower Observatory (ATTO) in the central Amazon. These data allowed us to determine the NBL height (h_N) by analyzing vertical profiles of wind speed and turbulence during LLJ events. Specifically, we examine whether the LLJ nose height (h_{N_jet}) corresponds to the top of the NBL, defined as the height where turbulence becomes negligible (h_{N_flux}), and under which conditions h_{N_jet} provides a good approximation of h_{N_flux} . This study contributes to the broader discussion on whether LLJ nose can be used as a reliable proxy for NBL depth in the absence of turbulence data. It also advances the understanding of NBL dynamics in the Amazon, since accurately estimating h_N is crucial for understanding how the vertical extent of the NBL affects the accumulation or dilution of greenhouse gases and other scalars.

3.2 Study site and data acquisition

Data were collected from Jan 2022 to Dec 2023 at the ATTO site (<https://www.attoproject.org/>) located about 150 km northeast of Manaus, in Central Amazonia. The site is situated in a region of old-growth tropical forests, where canopy structure and species composition vary strongly with soil attributes and water availability. The average canopy height is approximately 35 meters (Andreae et al., 2015; Gomes Alves et al., 2023).

Instruments were installed on two towers located on a plateau: the 80-m INSTANT tower (2.1468° S, 59.0068° W) and the 325-m TALL tower (2.1448° S, 59.0008° W), which are 670 m apart (Dias-Júnior et al., 2019) (Figure 3-1). Table 3-1 summarizes the instruments, variables, and measurement heights (in meters above ground). Specifically, the measurements at 35, 50, and 81 meters correspond to the INSTANT tower, while the remaining heights were measured at the TALL tower. The data from both towers are spatially consistent, as measurements at comparable heights (81 m at INSTANT and 83 m at TALL) showed no statistically significant

differences ($p > 0.05$), allowing them to be reliably used together in the same profile (Mendonça et al., 2025). Only data collected during nighttime (between 5:00 pm and 6:00 am LT (Local Time = UTC - 4h) were used in the analysis. Additionally, vertical profiles of equivalent radar reflectivity factor (Ze) were obtained from the Cloud Radar Profiler (Metek MIRA-35C), a vertically pointing Ka-band (35 GHz) radar installed at the ATTO site. The Ze dataset has a temporal resolution of 5 s.

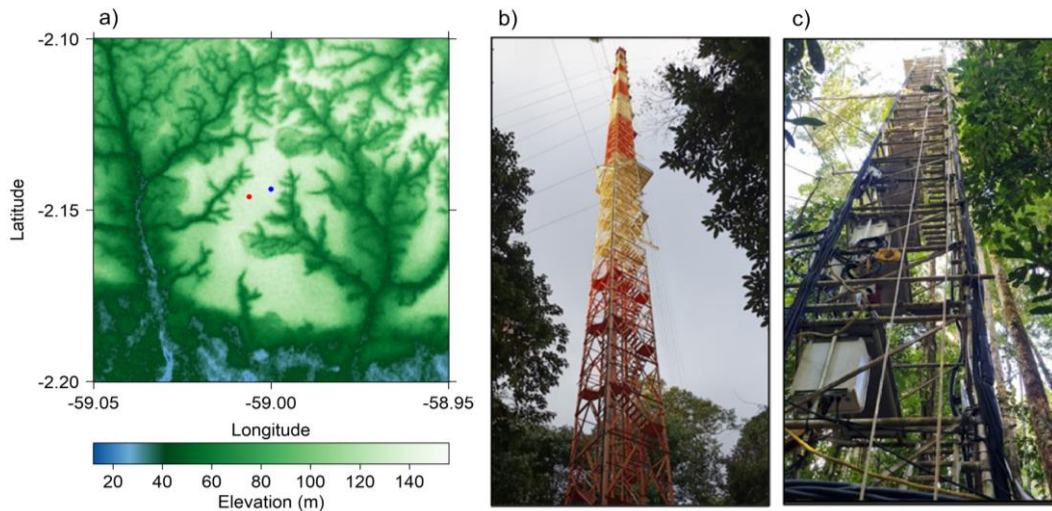


Figure 3-1. (a) Topographic map of the area surrounding the ATTO site showing the locations of the 325-m TALL tower (red dot) and the 80-m INSTANT tower (blue dot). (b-c) Images of the TALL and INSTANT towers. Topography was derived from Shuttle Radar Topography Mission (SRTM) data.

Table 3-1. Summary of instruments installed at the INSTANT and TALL towers, measured variables, measurement heights, and sampling rates.

(Continua)			
Instrument	Measured variable(s)	Height(s) (m)	Sampling rate
3D Ultrasonic Anemometer (THIES Clima)	Sonic temperature and wind componentes	35, 100, 127, 151, 172, 223, 247, 274, 298	10 Hz
Sonic Anemometer (CSAT3B, Campbell Sci.)	Sonic temperature and wind componentes	50, 81	10 Hz
Thermohygrometer (CS215-L, Campbell Sci.)	Air temperature and relative humidity	35, 50, 81, 100, 127, 151, 172, 223, 247, 274, 298	0.01 Hz

Table 3-1. Summary of instruments installed at the INSTANT and TALL towers, measured variables, measurement heights, and sampling rates.

(Conclusão)			
Instrument	Measured variable(s)	Height(s) (m)	Sampling rate
Pyranometer (CMP21, Kipp & Zonen)	Shortwave radiation (incoming and reflected)	81	0.01 Hz
Pygeometer (CGR4, Kipp & Zonen)	Longwave radiation (atmospheric and terrestrial)	81	0.01 Hz

3.3 Methodology

3.3.1 Quality Control and Data Processing

The first step of the analysis focused on data preparation, ensuring the reliability of the data. The data from the sonic anemometer were split every 18000 records (30-minute blocks) and then quality control (QC) and tilt correction methods were applied. The QC followed the procedures proposed by Zahn et al. (2016). They involve checking the completeness of the data record, identifying error flags and removing spikes.

Next, the turbulence statistics were calculated using 5-minute time windows as detailed in Section 3.2. Following this, two important aspects for the subsequent characterization of nights with LLJs were analyzed: the completeness of the vertical profiles (i.e., data availability at all heights) and the availability of data during the entire night. For each 5-minute profile, heights with missing values (NaN) were initially filled using cubic spline interpolation, which better represents the natural smooth transitions in the data. However, to avoid introducing unrealistic values or biases, interpolation was applied only when at least six of the eleven levels contained valid data, according to the following criteria: i) when a NaN occurred at height h with valid data at $h + 1$ and $h - 1$; ii) when 35 m or 298 m was NaN and the adjacent level contained valid data; or iii) when no more than two consecutive levels were NaN. Profiles that remained incomplete after applying these criteria were excluded. Completeness was first performed for each 5-minute profile using the wind speed as the reference variable, given its key role in identifying LLJ events. The same procedure was subsequently

applied to temperature, which exhibited recurrent gaps at 127 m due to intermittent thermohygrometer failures. In both cases, if a profile remained incomplete after interpolation, the entire 5-minute profile (including all variables) was removed. This procedure ensured temporal alignment across variables and preserved the physical consistency of the dataset.

A total of 111437 profiles were analyzed, with 26% excluded and 21% interpolated (of which 17% were interpolated at a single height along the vertical profile). Specifically, profiles from April and May 2022 did not meet the interpolation criteria, and were removed due to excessive missing values. No measurements were available for December 2022. Because of the thermohygrometer failure, approximately 46% of the temperature profiles required interpolation at 127 m, while 3% were interpolated at two consecutive heights. Subsequently, nights with less than 4 hours of data were excluded, while all remaining nights had more than 9 hours of data. As a result, the dataset used in this study consists of 508 nights, with 241 nights from 2022 and 267 nights from 2023 (Table 3-2), totaling 81134 profiles.

Table 3-2. Number of nights studied per month and year, and percentage (T%) of nights available in each year.

Year	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	T
2022	28	28	22	-	-	29	31	31	30	24	18	-	66%
2023	14	17	24	28	9	30	12	20	29	24	30	30	73%

3.3.2 Turbulence Statistics and Computed Variables

Turbulence statistics were calculated as follows: the mean horizontal wind speed (V) was computed as $V = (u^2 + v^2)^{0.5}$ where u and v are the streamwise and cross-stream wind components (ms^{-1}), respectively. The turbulent sensible heat flux (H) was estimated as $H = \rho C_p \overline{w'T'}$ (Wm^{-2}) where ρ is the air density (assumed to be 1.225 kgm^{-3}) and C_p is the specific heat of dry air at constant pressure ($1004 \text{ Jkg}^{-1}\text{K}^{-1}$). This formulation assumes constant ρ and C_p , as their variability has a negligible impact on the estimated H . The components w' and T' are the fluctuations of the vertical wind component and sonic virtual temperature, respectively. The turbulent kinetic energy (TKE, m^2s^{-2}) was calculated as $\text{TKE} = 0.5(\sigma_u^2 + \sigma_v^2 + \sigma_w^2)$ where σ represents the

standard deviation of each wind component. The kinematic momentum flux was estimated by the covariance $\overline{u'w'}$ (m^2s^{-2}).

To assess stratification within the NBL, the potential temperature (θ) was calculated as: $\theta = T_h + (g/C_p)h_r$. This equation provides θ from the dry-air adiabatic lapse rate (g/C_p), which is appropriate for near-surface atmospheric layers where variations in pressure and air density with height are minimal (Monson and Baldocchi, 2014). The variable T_h represents the temperature measured by the thermohygrometer (units of K), g is the gravitational acceleration, and h_r is the height (in meters) at which T_h was measured. Net radiation (R_N) was calculated as the sum of measured shortwave and longwave components (Table 3-1), to evaluate surface radiative cooling. Additionally, vertical profiles of the equivalent radar reflectivity factor (Z_e) were analyzed to assess cloud presence, which modulates surface radiative cooling.

3.3.3 Criteria for Identifying LLJs

Several criteria have been proposed in the literature for identifying LLJs, such as a minimum jet nose speed (e.g., greater than 3.0 ms^{-1}), or minimum difference (threshold) in wind speed between the levels above and below the jet nose (e.g., Andreas et al., 2000). These criteria vary depending on the objectives of the analysis and the resolution of the measurements. For example, Andreas et al. (2000) used thresholds around 2.0 ms^{-1} with radiosonde data, while Banta et al. (2002) applied thresholds of 0.5 ms^{-1} for Doppler Lidar and 1.5 ms^{-1} for sodar/profiler systems.

We tested fixed-threshold criteria, defining the jet nose as a local wind speed maximum exceeding adjacent levels by 0.3 ms^{-1} to 2.0 ms^{-1} . However, visual inspection revealed that low thresholds (e.g., 0.3 ms^{-1}) led to false detections, while high thresholds (e.g., 2.0 ms^{-1}) excluded many valid cases. Given that the sonic anemometers were installed at multiple levels with vertical spacing of 15-51 m, much finer than the effective vertical resolution of typical radiosonde or sodar profiles (50-200 m), we adapted the threshold criteria to better suit our data, as described below.

To systematically identify LLJs, the main requirement was the presence of a wind speed maximum (jet nose), followed by a decrease in wind speed at least at two levels above and below the jet nose. This requirement is qualitative and ensures that the vertical wind profile exhibits the typical LLJ shape described in previous studies

(Banta et al., 2002; Pichugina and Banta, 2010). However, in order to process the large data set efficiently, we implemented a four-step algorithm:

- Step 1 (jet nose identification): The objective is to ensure that the wind-speed profile contains a well-defined maximum. Each 5-minute wind profile was analyzed to detect a jet nose following the main requirement described above. When detected, two primary characteristics were recorded: jet nose height (h_{nose}) and jet nose speed (V_{nose}).

- Step 2 (jet nose validation): The objective is to ensure that the detected jet nose is physically meaningful. A jet nose was considered valid when the increase in wind speed between the jet nose and the level immediately below it exceeded $\Delta V > 0.3 \text{ ms}^{-1}$. This value was selected to ensure that the detected maximum exceeds the accuracy of the sonic anemometers (CSAT3: $\pm 0.08 \text{ ms}^{-1}$; THIES: $\pm 0.1 \text{ ms}^{-1}$).

- Step 3 (hourly LLJ validation): The objective is to ensure the temporal consistency of the LLJ within each hour. We computed the percentage of 5-minute profiles with a valid jet nose, defined as $P_{\text{jets}} = N_{\text{jets}}/N_p$, where N_{jets} is the number of profiles with valid jet noses and N_p is the total number of profiles within that hour. An LLJ was considered valid during that hour if $P_{\text{jets}} \geq 0.5$ (i.e., if at least 50% of the profiles within the hour contained a jet nose). Lower values of the criterion (e.g., $P_{\text{jets}} \geq 0.4$) increased false positives, while higher values (e.g., $P_{\text{jets}} \geq 0.6$) excluded valid cases. The value of 0.5 provided the best agreement with visual inspections.

- Step 4 (LLJ classification): The objective is to ensure that only persistent LLJs are classified as valid events. An LLJ event was considered valid for a given night if it persisted for at least three consecutive hours. This condition ensured that only persistent LLJs were included in the analysis. Longer durations (e.g., $\geq 4 \text{ h}$) reduced the number of identified nights but did not significantly alter the overall trends. Empirical analysis also showed that, in some cases, the jet nose was intermittently detected, which may reflect variability in LLJ evolution or situations where the jet nose ascended above the highest measurement level (300 m). Further investigation is required to better characterize such cases, which is beyond the scope of this study.

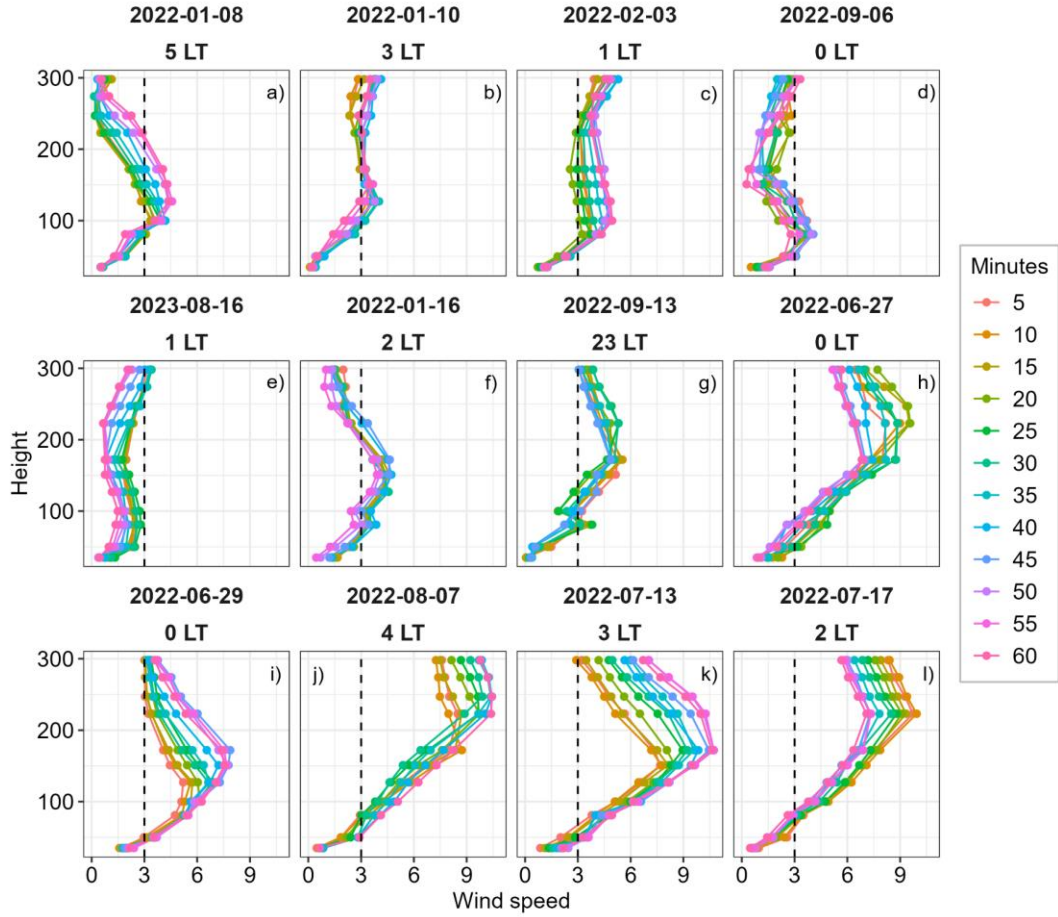


Figure 3-2. Sample of 5-minute wind profiles displaying Low-Level Jets (LLJs). Each panel shows data from a specific night over a one-hour period in local time (LT). The line colors represent individual 5-minute profiles within the given hour. Panel (e) shows an example of a night not classified as LLJs. The vertical dashed line marks the 3.0 ms^{-1} wind speed.

We have identified various types of wind profiles (Figure 3-2). Some showed a well-defined jet nose (Figure 3-2a, d, k, l), while others exhibited minimal wind reduction above the maximum (Figure 3-2b,c). In some profiles, two wind speed peaks were observed; however, only the peak followed by a decrease in wind speed at the two levels immediately above it was classified as a valid jet nose (Figure 3-2f,g), in accordance with the adopted criteria. Temporal variations have also been found: in some cases, jet height and speed decreased over time (Figure 3-2h), while in others, both variables increased (Figure 3-2i, j). There were also cases when the jet height remained constant while the nose speed increased (Figure 3-2k) or decreased (Figure 3-2l). Conversely, some profiles showed a jet nose with weak winds, so that the jet nose speed was below 3.0 ms^{-1} (Figure 3-2e). Such profiles (which occurred during

very stable conditions) have been excluded from the analysis. All data processing and analysis were performed using the RStudio environment (R Core Team, 2024).

3.3.4 Nocturnal Boundary Layer Height Estimation

The NBL height (h_N) has been defined as the lowest height where the turbulent flux dropped below 5% of its surface value (Lenschow et al., 1988; Stull, 1988). This turbulence-based approach provides an estimate of h_N consistent with the fundamental definition of the boundary layer. In this study, this criterion was applied to the vertical profiles of sensible heat (H) to estimate h_N , hereafter denoted as h_{N_flux} (Mendonça et al., 2025). The TKE profile was examined as a complementary variable, since it decreases with height but does not become negligible at the top of the NBL, due to non-turbulent motions that still contribute to the velocity variances (Acevedo et al., 2016).

Additionally, we used the jet nose height to estimate h_N from the wind speed profile, referred to as h_{N_jet} . This approach was based on the premise that turbulence and wind shear are typically enhanced below the LLJ core and suppressed above it (Banta et al., 2003). Therefore, we used h_{N_flux} estimates as a reference to evaluate h_{N_jet} (Figure 3-3a).

Our analysis was structured in two complementary parts. The first focused on the mean behavior of the NBL and LLJ characteristics across different classes. The dataset was classified by jet nose height (h_{nose}) into three classes (low, mid and high) and by wind direction at the jet nose (WD_{nose}) into two sectors (N-NE and S-SE), because distinct WD_{nose} patterns were observed within each h_{nose} class. For each combination of h_{nose} and WD_{nose} , mean profiles were used to derive h_{N_flux} and h_{N_jet} . These mean profiles are ensemble averages computed from all 5-min wind profiles belonging to each class, regardless of the night to which they belong. The detailed criteria for these classifications are provided in Section 4.1. Furthermore, the following auxiliary variables were examined to evaluate the static and dynamic characteristics of the atmosphere: potential temperature (θ) and its gradient ($d\theta/dz$), wind shear (dV/dz), kinematic momentum flux ($\overline{u'w'}$) and net radiation (R_N). The local Richardson number (Ri) was used to assess the balance between static stability ($d\theta/dz$) and dynamic instability (dV/dz), providing an indication of whether turbulence is suppressed or generated. To compute these gradients, smooth profiles of $\theta(z)$ and

$V(z)$ were fitted using cubic spline interpolation, from which analytical derivatives were obtained (Figure 3-3b).

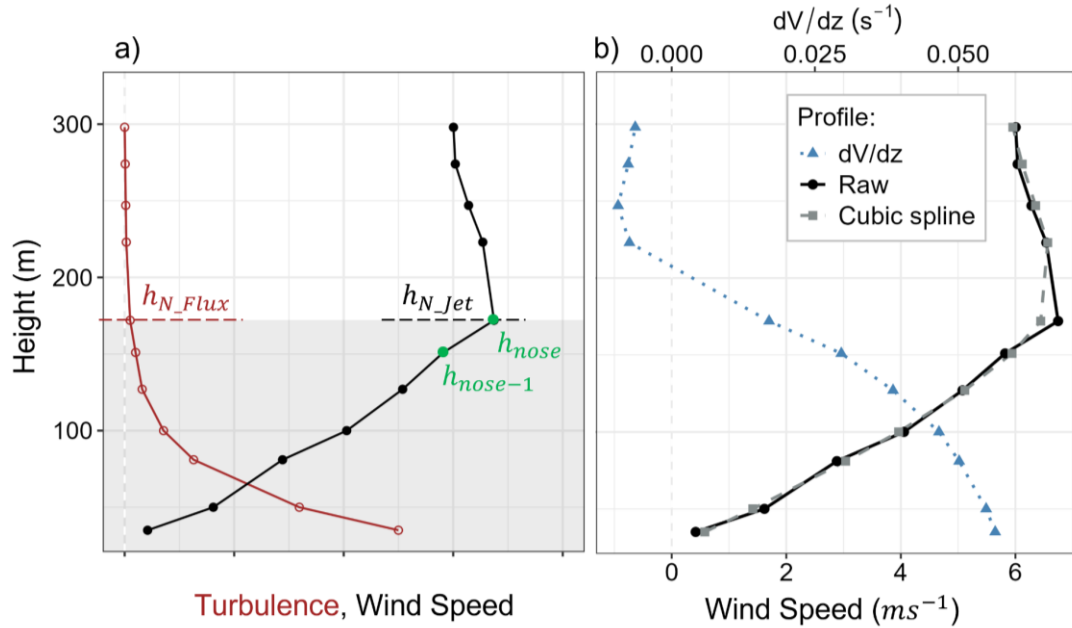


Figure 3-3. Schematic illustration of a typical LLJ event, showing: a) turbulent flux (brown) and wind speed (black), and b) wind speed (black), cubic spline interpolation (grey), and vertical gradient of wind speed (dV/dz , blue). The shaded area represents the vertical extent of the NBL, and the dashed lines indicate the NBL height estimates from turbulence (h_{N_flux}) and jet nose height (h_{N_jet}). h_{nose} and h_{nose-1} refer to the jet nose and the level immediately below it, respectively.

The second part consisted of a case study. To further investigate the relationship between LLJ height and NBL height, two contrasting nights were selected as representative cases: one in which the NBL height estimates derived from h_{N_flux} and h_{N_jet} closely matched, and another where they differed substantially. This case study focused on the temporal evolution of the variables throughout each night. In addition to the previously described variables, the radar reflectivity factor (Z_e) was analyzed as an indicator of cloud presence, which can influence radiative cooling. The Ri was examined in a reformulated bulk version, referred to as $Ri_{bulk} = (g/\theta)(\Delta\theta/\Delta z)(\Delta V/\Delta z)^{-2}$, where Δ represents the bulk difference between the canopy-top level (35 m) and the jet nose height. This bulk formulation provides an integrated measure of the thermal and dynamical balance within the layer below the jet, indicating whether the shear associated with the LLJ is sufficient to overcome the static stability and sustain turbulence. When the value of Ri is less than critical (≈ 0.2 -

0.25), the flow can become turbulent (Freire et al., 2019). Although the critical value of Ri strictly applies to idealized conditions, it is used here as an empirical reference, since TKE clearly increases for $Ri < 0.25$ (not shown here).

3.4 Results

3.4.1 Characteristics of LLJs

From the data set available (508 nights), 106 satisfied the criteria for a LLJ. Histograms of jet nose height (h_{nose}) with the median jet nose speed (V_{nose} , blue dot) are shown in (Figure 3-4a). The h_{nose} values occurred predominantly below 172 m (70%), although the most frequent individual value was 223 m, accounting for approximately 24% of the total observations (noting that only observational levels can be “detected” as possible jet heights). An increasing trend in V_{nose} with height was evident, with the median rising from 3.7 ms^{-1} at 81 m to 7.4 ms^{-1} at 247 m. The interquartile range (IQR) also increased with height, indicating larger variability in V_{nose} at higher levels. This increase in wind speed with height is expected, as drag forces near the canopy reduce wind speeds in the lower layers. These results highlight that LLJs, although weaker (i.e., with lower wind speed at the jet nose) than in previous studies, were frequently observed at the ATTO site. For proper comparison of the lowest observed LLJ, its height was referenced to the zero plane displacement (d). We used $d = 0.9 h$ (approximately 31.5 m) based on Chor et al. (2017) for central Amazon, with canopy height $h = 35 \text{ m}$ at our site. With this correction, this jet at 81 m above the ground corresponds to about 50 m ($z - d$, comparable to heights over flat terrain), indicating that these jets occurred at lower levels in the Amazon forest than elsewhere (Pichugina and Banta, 2010; Dias-Júnior et al., 2017).

The tendency for more intense LLJs to occur at higher levels was also reported by Santana et al. (2015), based on radiosonde data collected during January and February 1999 as part of the Wet Season Atmospheric Mesoscale Campaign (WETAMC) conducted in the southwestern Amazon. However, in their study, the weakest jets had wind speeds between 4.1 and 6.9 ms^{-1} , with jet noses ranging from 140 to 950 m. In contrast, our observations revealed the presence of even weaker and lower jets (e.g. 81-100 m). This difference is likely attributable to the higher temporal

and vertical resolution of our dataset, which allows weak and shallow jets to be captured but would not be resolved by radiosondes.

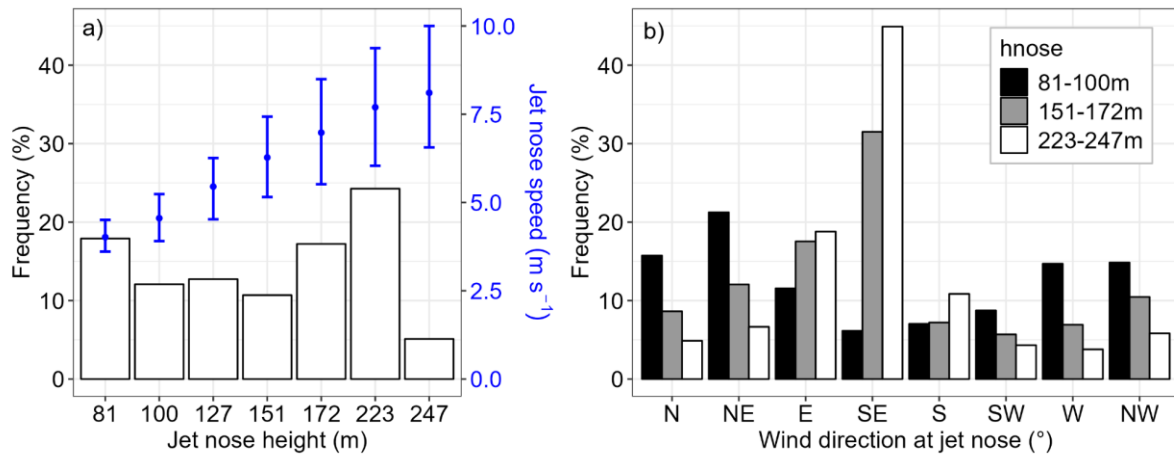


Figure 3-4. Frequency distributions of: a) jet nose height (h_{nose}), with median nose speed (blue dot) and interquartile range (vertical bars); and b) wind direction at jet nose categorized by h_{nose} into 81-100 m (black), 151-172 m (grey) and 223-247 m (white).

The h_{nose} values were empirically grouped into three classes: 81-100 m (30%, low jets), 151-172 m (28%, mid-jets), and 223-247 m (29%, high jets). This approach suits our objective of investigating the relationship between jet nose height and the NBL height. Moreover, the selected ranges ensure a balanced number of observations across the three classes, and the differences observed between these classes support the relevance of this classification for the present dataset. We examined the frequency distributions of wind direction at the jet nose (WD_{nose}) within each h_{nose} classes (Figure 3-4b) and the results indicated that WD_{nose} changed among the different classes.

For jets between 81-100 m, the WD_{nose} was distributed among west (W), northwest (NW) and north (N), each accounting for approximately 14%, with a predominance from the northeast (NE, 21%). On the other hand, in the 151-172 m class, there was a significant change in the predominant direction, with southeast (SE, 32%) and east (E, 18%) becoming dominant. This pattern became even more pronounced in the 223-247 m class, where SE accounted for 45% of occurrences. The increased frequency of higher jets originating from E-SE aligns with the findings of Dias-Júnior et al. (2017), who reported a higher frequency of LLJs from the east

quadrant. However, their analysis was based on 1-hour averaged sodar profiles collected at a rural site near the Amazon River in the central Amazon, resulting in an average h_{nose} of 570 ± 209 m and V_{nose} of 9.2 ± 2.9 ms⁻¹. These elevated jets are likely decoupled from the surface layer and differ from the LLJs detected by the ATTO (< 300 m) due to differences in the vertical resolution of sodar and sonic measurements. Despite these differences, the two datasets are complementary, as both reveal an increased contribution of easterly winds at higher jet heights.

Overall, a clear distinction was observed between jet heights: lower-level jets predominantly originated from the N-NE, while higher-level jets were mainly from the E-SE and were associated with larger wind speeds. Given this, subsequent analyses were conducted by grouping the data according to the predominant wind sectors: 337.5°-89° and 90°-202.5° referred to here as N-NE and S-SE, respectively. Although WD_{nose} is not a physical driver that directly influences the NBL, it serves as a proxy for the underlying physical conditions. We performed an additional classification using physical variables (net radiation and thermal stability), and the same patterns observed with the WD_{nose} classes were reproduced (not shown here). This indicates that WD_{nose} reflects broader physical differences that are not captured by R_N or $d\theta/dz$ alone. Previous studies at the ATTO site have shown that different wind sectors are associated with distinct topographic gradients (Figure 3-1a), which can modulate near-surface circulations (Chamecki et al., 2020; Corrêa et al., 2021), supporting the use of WD_{nose} as a diagnostic grouping variable.

3.4.2 Estimates of NBL height from Turbulence and LLJ Height

To directly address the main objective of the study, we assess whether h_{N_jet} can serve as a proxy for h_{N_flux} , based on the classification by jet height and wind direction at the jet nose (Section 4.1). The corresponding vertical profiles of wind speed, normalized sensible heat flux (H) relative to the value at the canopy top (35 m), and turbulent kinetic energy (TKE) are shown in Figure 3-5.

For the 81-100 m jet class (Figure 3-5a), wind speeds at jet nose were similar for both N-NE and S-SE directions, with a median of approximately 3.8 ms⁻¹ (IQR: 3.3 - 4.5 ms⁻¹). The profiles of normalized H and TKE (Figure 3-5b, c) exhibited higher values near the canopy top, decreasing with height. TKE values at 35 m were larger with winds from N-NE (around 0.2 m²s⁻²) than with winds from S-SE (around 0.05

m^2s^{-2}). However, in both wind direction classes, H and TKE decreased with height, reaching negligible values near 100 m. The difference in θ between 100 and 35 m ($\Delta\theta_{(100-35)}$) was also similar (Figure 3-5d), with a median of 2.0 °C for N-NE and 1.8 °C for S-SE, indicating comparable thermal stratification conditions. These results indicate that when the jet nose was between 81-100 m, the NBL height estimated from turbulence profiles ($h_{\text{N_flux}}$) corresponded to the jet nose height ($h_{\text{N_jet}}$) regardless of the wind direction.

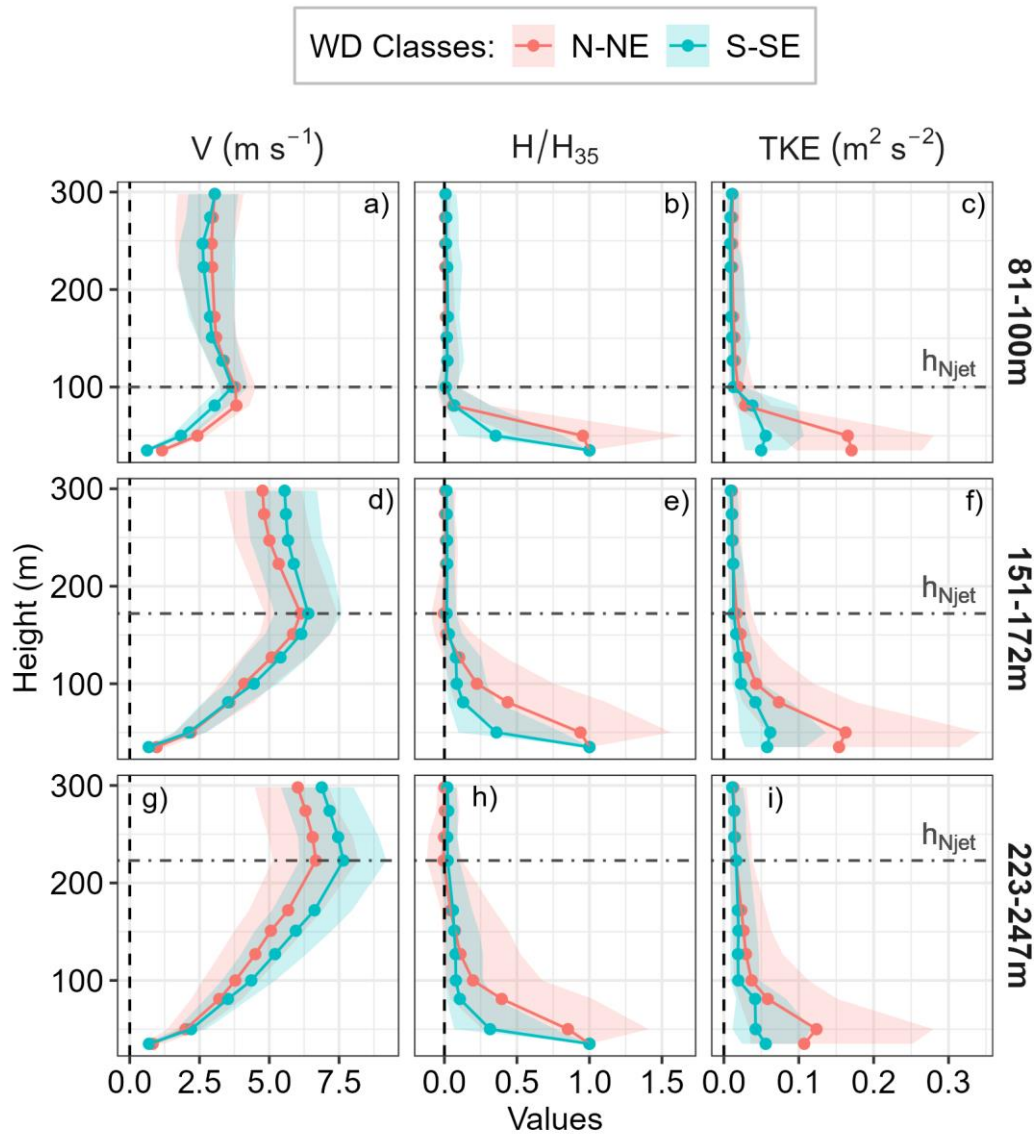


Figure 3-5. Median vertical profiles of wind speed (V), normalized heat flux (H/H_{35}), and turbulent kinetic energy (TKE) for low-level jets associated with wind directions: N-NE (red) and S-SE (blue). Panels from top to bottom correspond to jet nose height-classes. Shaded areas represent the interquartile range.

For the 151-172 m jet class, V_{nose} was larger from S-SE (6.5 ms^{-1} , IQR of $5.2\text{--}7.8 \text{ ms}^{-1}$) than from N-NE (5.6 ms^{-1} , IQR: $4.6\text{--}6.8 \text{ ms}^{-1}$) (Figure 3-5d). Although jets from the S-SE exhibited higher wind speeds, turbulence was stronger and extended higher in N-NE cases. Both TKE and H approached zero around 150 m for N-NE, while for S-SE, turbulence decreased more sharply, reaching near-zero values around 100 m (Figure 3-5e, f). Thus, in this jet class, agreement between h_{N_jet} and h_{N_flux} occurs only for N-NE jets showing that similar jet heights do not ensure similar NBL heights.

Similar patterns were observed in the 223-247 m jet class. The median wind speed profile showed a similar jet nose height (223 m) for both wind directions (Figure 3-5g). Despite that, the median V_{nose} reached 6.3 ms^{-1} (IQR: $4.9\text{--}7.8 \text{ ms}^{-1}$) for N-NE and 7.4 ms^{-1} (IQR: $5.9\text{--}9.0 \text{ ms}^{-1}$) for S-SE, indicating stronger LLJs from the S-SE direction. Although S-SE jets presented higher V_{nose} , the vertical extent of turbulence was deeper in N-NE cases, as both TKE and H decreased to near-zero values around 172 m (Figure 3-5h, i). For jets from S-SE, turbulence was confined below 100 m. Therefore, as in the previous class, h_{N_jet} aligns with h_{N_flux} only for N-NE jets.

Interestingly, θ increased with height across all measurement levels of the tower (Figure 3-6b, f, j), indicating that a stable stratification extended above the height of the jet nose. Moreover, the turbulent layer was confined to levels lower than the region of increasing θ . This pattern was reported by Mendonça et al. (2025) during nights without LLJs at ATTO, and also by Mahrt et al. (1979) in four different field experiments conducted in temperate-zone latitudes, where LLJs developed within a thermally stratified environment, below a persistent inversion layer.

The profiles of the $d\theta/dz$ exhibited two contrastingly distinct shapes, reflecting the differences in the vertical structure of thermal stratification (Figure 3-6a, e, i). For N-NE jets, the maximum in $d\theta/dz$ occurred near the jet nose height, suggesting enhanced thermal stability around the LLJ core. However, the relatively weak stratification above the canopy top allowed turbulence to persist, favoring the coupling between the jet and the canopy top, which may explain the closer agreement observed between h_{N_jet} and h_{N_flux} in these cases.

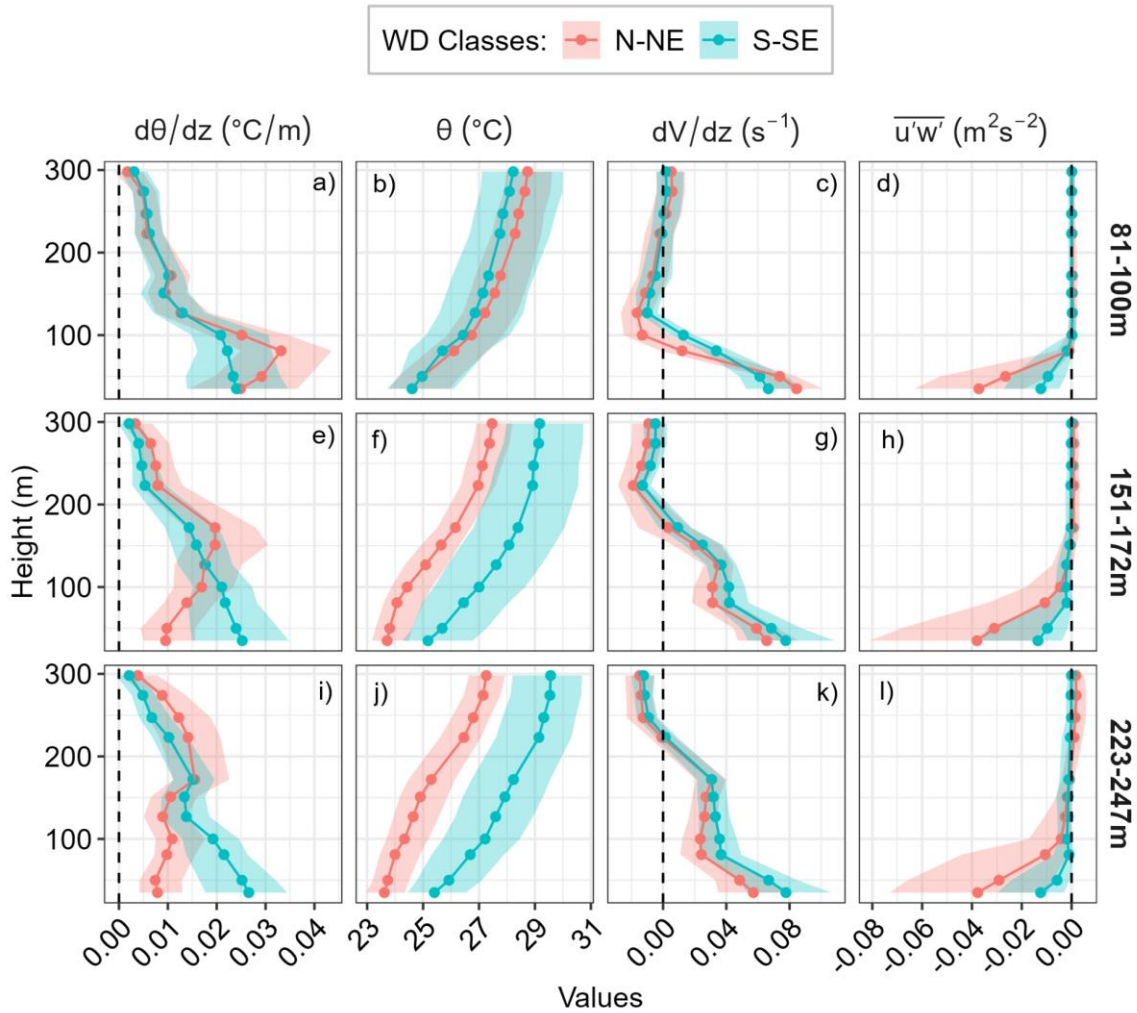


Figure 3-6. Median vertical profiles of potential temperature gradient ($d\theta/dz$), potential temperature (θ), vertical wind shear (dV/dz), and kinematic momentum flux ($\overline{u'w'}$) for low-level jets associated with wind directions: N-NE (red) and S-SE (blue). Panels from top to bottom correspond to jet nose heights classes. Shaded areas represent the interquartile range.

A similar correspondence between h_{N_jet} and h_{N_flux} was reported by Pichugina and Banta (2010) for profiles exhibiting a classic LLJ shape. In their study, the turbulence-based boundary layer depth was estimated from streamwise velocity variance profiles obtained with high-resolution Doppler lidar over the Great Plains of the United States. Although both N-NE and S-SE jets in our study exhibited a classic LLJ shape, only the N-NE cases were consistent with the findings of Pichugina and Banta (2010). In contrast, S-SE jets showed the strongest thermal stratification above the canopy top, suggesting a decoupled boundary layer structure, which likely contributed to the difference between h_{N_jet} and h_{N_flux} values.

Although the vertical shear profiles (dV/dz) were similar between N-NE and S-SE jets (Figure 3-6c, g, k), consistent with the previously presented wind profiles, contrasting turbulence responses were observed. In the S-SE jets, strong thermal stratification inhibited vertical mixing despite similar mechanical forcing, indicating that thermal stratification modulates the efficiency of shear-driven turbulence. According to Pichugina and Banta (2010), in classic LLJ profiles, the zero-crossing in the wind-shear profile (representing the top of the shear layer) closely corresponds to the NBL height. However, in our case, this agreement was observed only for the N-NE jets.

Accordingly, $\overline{u'w'}$ was consistently larger above the canopy during N-NE jets, suggesting a coupled boundary layer where surface-generated turbulence could extend upward. In contrast, limited vertical mixing during S-SE jets resulted in a difference between the shear-based and turbulence-based boundary layer heights (Figure 3-6h, l).

Radiative cooling very likely contributed to the differing thermal stratification conditions observed between LLJs from N-NE and S-SE directions. R_N measured at 81 m showed enhanced radiative loss during S-SE jets (-45 Wm^{-2}) compared to N-NE (-20 Wm^{-2}) (Figure 3-7a). This enhanced cooling was associated with a warmer canopy under S-SE jets, as indicated by θ measured at 35 m (Figure 3-7b). Notably, this pattern persisted even when only the dry-season months (June to August) were considered (not shown here), suggesting that the differences were not solely driven by seasonal variability but were also associated with the prevailing wind direction.

Figure 3-8 presents the NBL height estimates derived from turbulence (h_{N_flux}), jet nose height (h_{N_jet}), and the height at which the $d\theta/dz$ reached its maximum ($h_{maxstab}$), grouped by h_{nose} and WD_{nose} classes. This figure summarizes the h_N estimates previously obtained from the vertical profiles shown (Figure 3-5). For jets from the N-NE (Figure 3-8a), h_{N_flux} increased with increasing h_{N_jet} , suggesting a direct relationship between jet height and the depth of the turbulent layer. In these cases, $h_{maxstab}$ was typically located near the jet nose height, reinforcing the idea of a coupled boundary layer structure. In contrast, jets from the S-SE sector (Figure 3-8b) exhibited consistently lower h_{N_flux} values (median $\approx 100 \text{ m}$), regardless of h_{nose} class. In such cases, $h_{maxstab}$ remained close to the canopy top and exhibited greater variability. These features suggest a shallower and less vertically developed turbulent layer under S-SE winds.

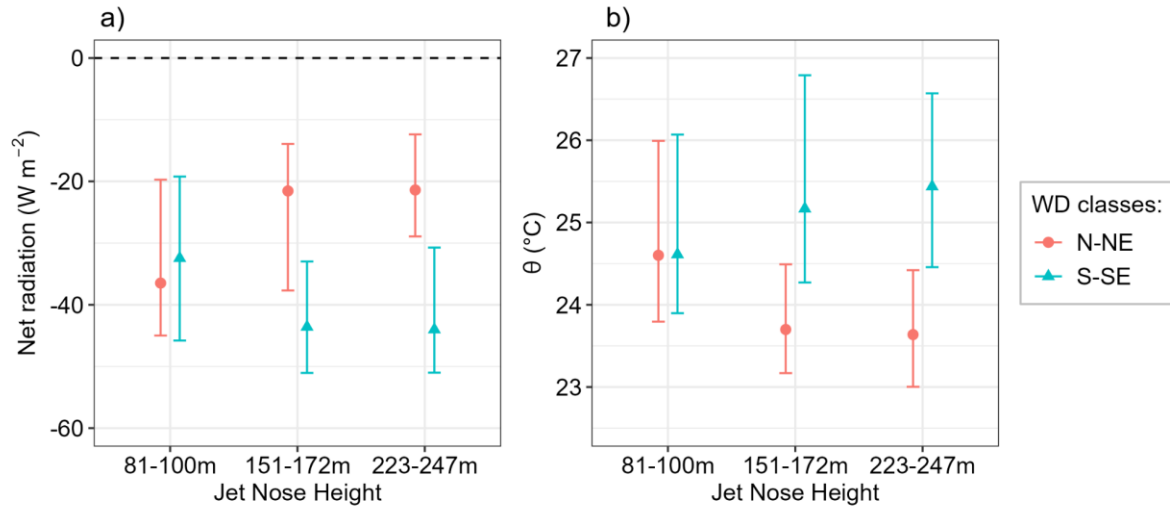


Figure 3-7. Median values of a) net radiation (R_N) and b) potential temperature (θ) corresponding to each jet nose height range, grouped by wind direction: N-NE (red circles) and S-SE (blue triangles). Vertical bars indicate the interquartile range.

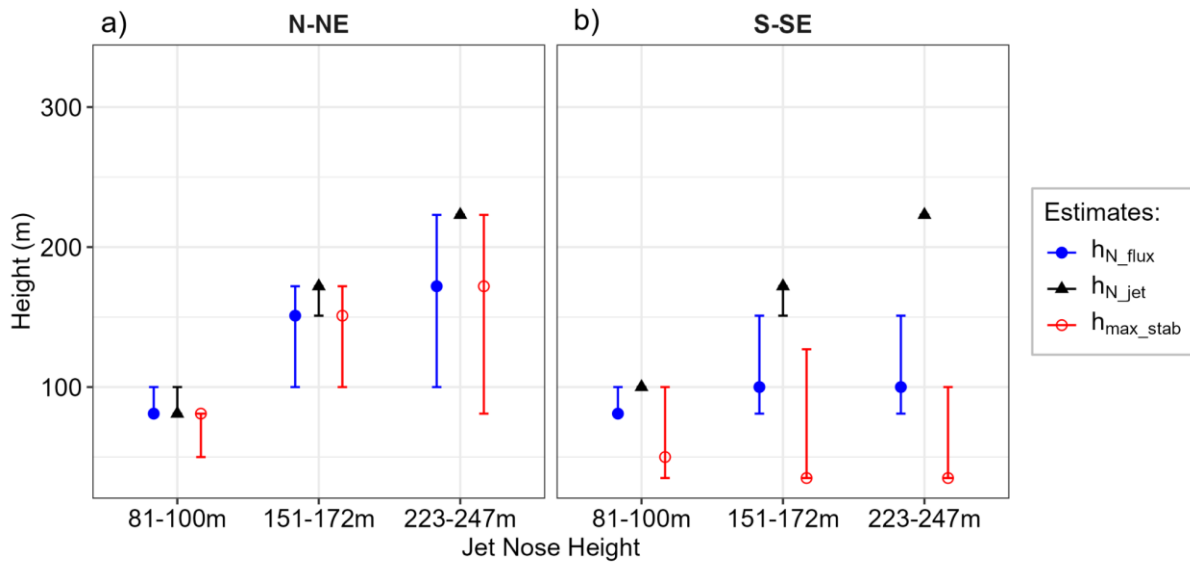


Figure 3-8. Median values of nocturnal boundary layer height (h_N) estimates for each jet nose height range, grouped by wind direction: a) N-NE and b) S-SE. Estimates are based on turbulence (h_{N_flux} , blue circles) and jet nose height (h_{N_jet} , black triangles). h_{max_stab} indicates the height of maximum potential temperature gradient (red open circles). Vertical bars represent the interquartile range.

As a result, when comparing the same h_{nose} class across wind directions, h_{N_jet} and h_{N_flux} were similar for jets with heights between 81-100 m. However, this correspondence was not observed for higher jets (151-172 m and 223-247 m). For the 223-247 m class, h_{max_stab} varied widely (IQR: 81-223 m) for N-NE jets, while for S-SE

jets it remained confined to lower altitudes (IQR: 35-100 m), contributing to the difference between h_{N_jet} and h_{N_flux} estimates. This difference highlights the importance of thermodynamic stratification when using h_{N_jet} as a proxy for estimating the NBL depth in the Amazon forest.

The reasons for the observed discrepancies between h_{N_flux} and h_{N_jet} under some conditions, but not others will be discussed in Section 5.

3.4.3 Time Evolution of Nocturnal LLJs - Case study

To further illustrate the patterns observed in Section 4.2, we analyze two contrasting LLJ cases to assess how both the jet structure and turbulence evolve during the night. Although the formation and dynamics of LLJs are beyond the scope of this diagnostic study, understanding their temporal evolution helps interpret how the jet interacts with turbulence and thermal stratification during nocturnal periods. Figure 3-9 presents time-height cross sections for two representative cases: July 13, 2022 where the NBL height estimates from h_{N_jet} and h_{N_flux} closely matched, and Aug 07, 2022 where they differed substantially. In addition, these nights were selected as representative cases of LLJs events with WD_{nose} from N-NE and S-SE, respectively (Figure 3-9i-j).

On July 13, 2022 (Figure 3-9a), the LLJ emerged around 20:00 LT with a maximum wind speed of 4.2 ms^{-1} at 100 m. Over the subsequent hours, h_{nose} ascended to 151 m and V_{nose} exceeded 6 ms^{-1} . By 04:00 LT, the h_{nose} reached 172 m with a V_{nose} of 10.72 ms^{-1} , representing the highest values observed that night. In contrast, on 07 Aug 2022 (Figure 3-9b), the LLJ was already well established early in the night: its nose was first identified at 172 m around 21:00 LT with 7.8 ms^{-1} , and then remained between 172-223 m with speeds ranging from $8\text{-}11 \text{ ms}^{-1}$.

The vertical distribution of turbulence further distinguishes these two cases. Figure 3-9c-f, presents time–height sections of TKE and kinematic sensible heat flux ($\overline{w'T'}$) with the jet nose height indicated by black dots. On July 13, both variables (Figure 3-9c, e) became negligible near the jet nose height during most of the night, indicating that the turbulence was coupled with the jet structure, and therefore h_{N_jet} provided consistent estimates of h_{N_flux} . This is in line with the expectation that turbulence generated by shear near the jet core can sustain mixing to higher altitudes under moderately stable conditions (Banta et al., 2006) Conversely, on August 07

(Figure 3-9d, f), turbulent activity remained below 100 m for most of the night. In this case, h_{N_jet} estimates were higher than those based on h_{N_flux} . This suggests a decoupling between the surface and the jet core, characteristic of very stable nights.

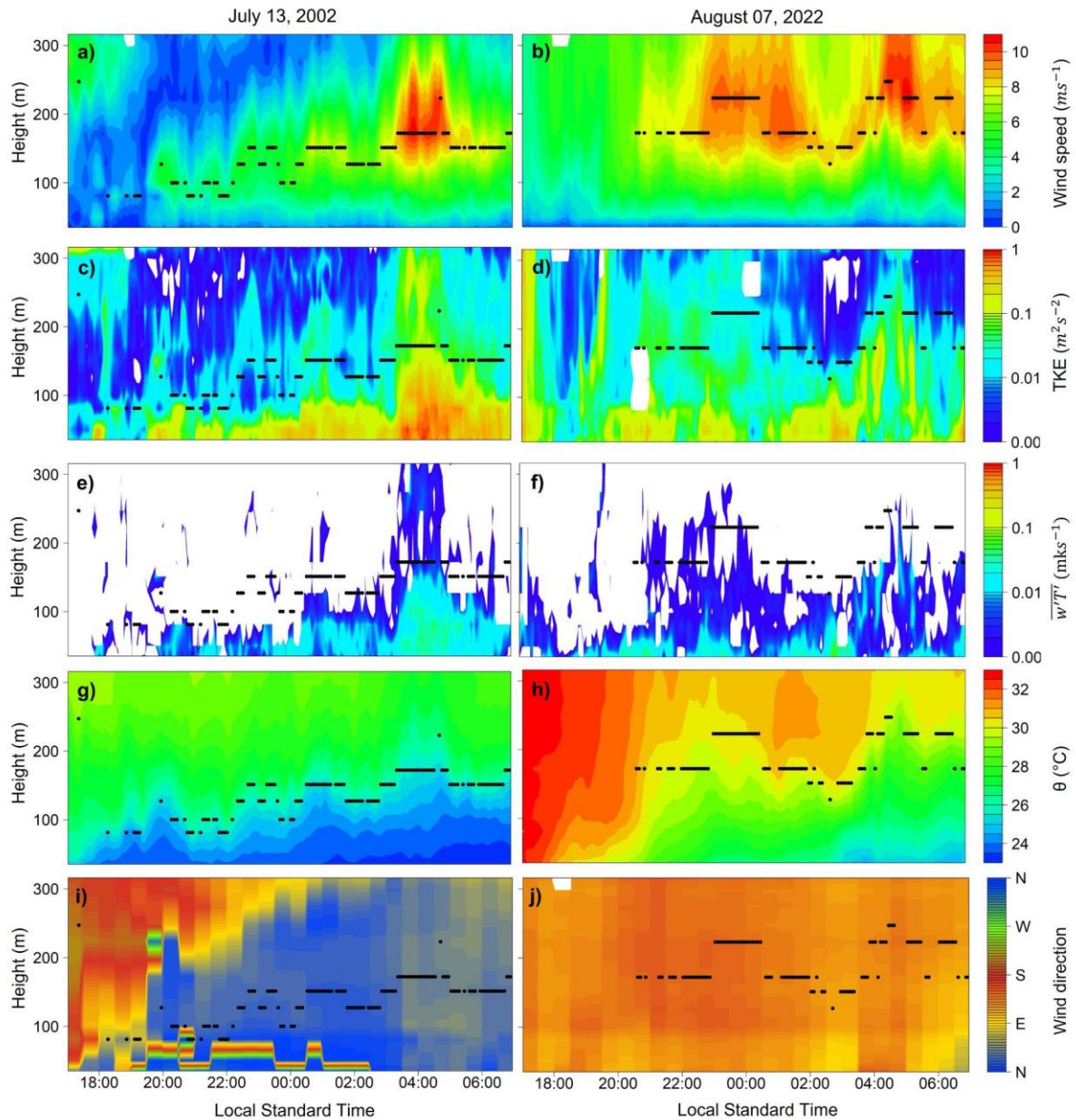


Figure 3-9. Time-height cross section of (a, b) wind speed, (c, d) turbulent kinetic energy (TKE), (e, f) kinematic sensible heat flux ($\overline{w'T'}$), (g, h) potential temperature (θ) and (i, j) wind direction for the nights of (left panels) July 13, 2022 and (right panels) August 07, 2022. The dots represent the height of LLJ. All variables are derived from tower measurements.

Thermal conditions above the canopy also support this interpretation. Figure 3-9g and Figure 3-9h shows time-height cross sections of θ , while Figure 3-10

presents the behavior of R_N measured at 81 m, and stability ($d\theta/dz$) at 35 m for each night. On 07 August, θ reached 32.5 °C above the canopy during the evening transition (17:00-19:00 LT), dropping rapidly to a minimum of 24.8 °C at 04:00 LT. This strong cooling was accompanied by a persistent temperature gradient (median $d\theta/dz \approx 0.03$ °Cm⁻¹), and R_N reached -60 Wm⁻². These results indicate stronger surface cooling and a rapid establishment of stable atmospheric conditions. In contrast, on July 13, θ was lower in the evening (≈ 25 °C), the cooling was less intense (minimum ≈ 23.3 °C), and the stratification weaker (median $d\theta/dz \approx 0.01$ °Cm⁻¹), with R_N remaining closer to -20 Wm⁻².

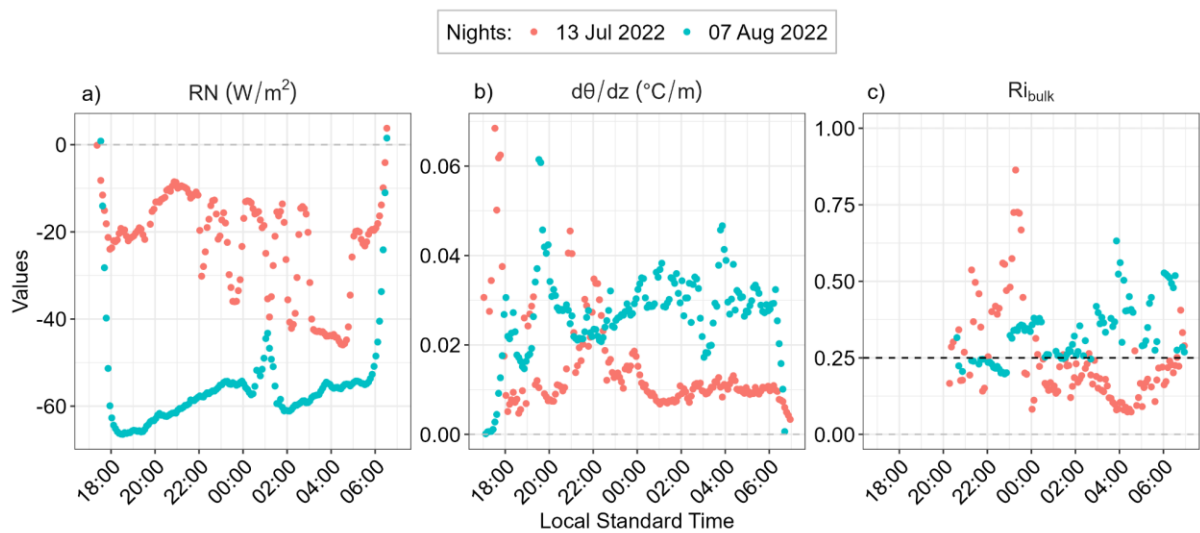


Figure 3-10. Temporal evolution of a) net radiation (R_N) at 81 m, b) potential temperature gradient ($d\theta/dz$) and c) bulk Richardson number (Ri_{bulk}) (both at 35 m) during the nights of July 13, 2022 (red dots) and August 07, 2022 (blue dots).

A potential factor contributing to these differences was the presence of clouds. Although both nights occurred in the dry season, radar reflectivity data (Figure 3-11) indicated more cloud cover on July 13, which may have limited radiative cooling above the canopy and contributed to a weaker inversion. These findings are consistent with the results of Greco et al. (1992), who reported that nocturnal inversions in central Amazonia are primarily driven by radiational cooling of the rainforest canopy and are strongly controlled by cloudiness.

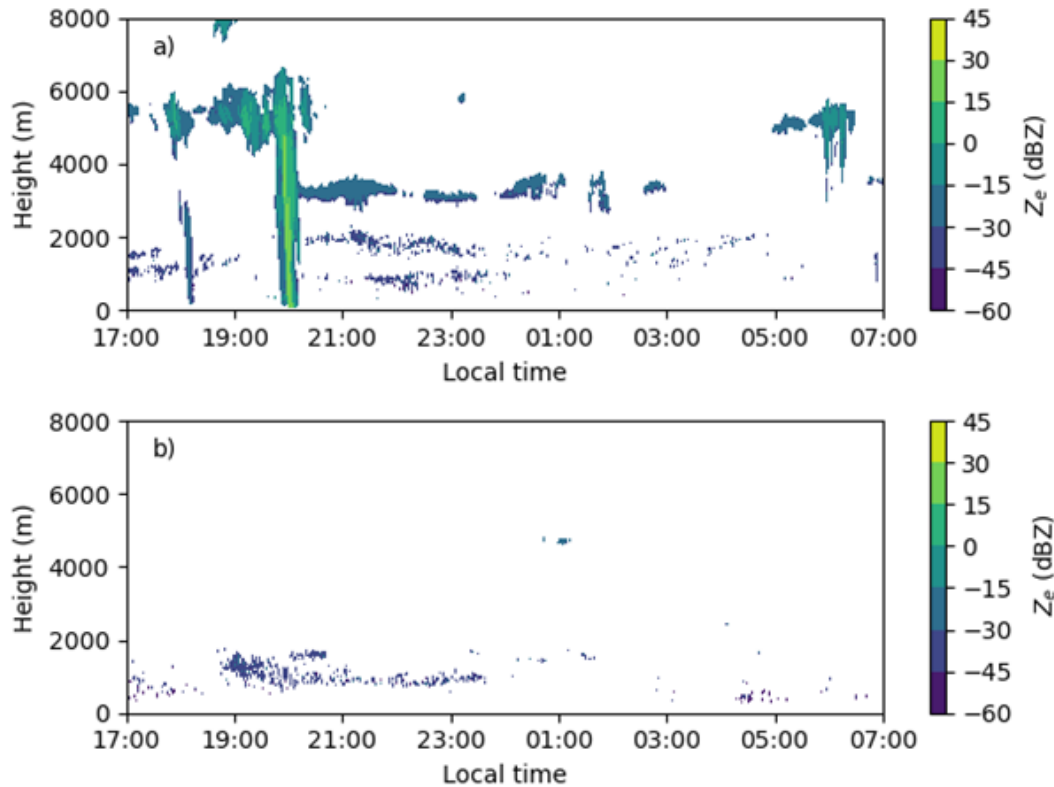


Figure 3-11. Time–height cross sections of equivalent radar reflectivity factor (Z_e) for (a) July 13, 2022 and (b) August 07, 2022. Color shading represents Z_e values in dBZ.

Contrasting states of coupling between the atmosphere and the canopy were also identified through the analysis of the bulk Richardson number (R_{i_bulk}) at 35 m during both nights (Figure 3-10c). On 07 August, R_{i_bulk} values remained consistently above the critical threshold (≈ 0.25), above which shear-generated turbulence is no longer sustained. This indicates suppressed turbulent mixing and a dynamically decoupled layer above the canopy. In contrast, on July 13, R_{i_bulk} initially stayed below 0.25, briefly increased around 22:00 LT, and then decreased again, suggesting prevailing weak thermal stratification and sustained turbulent coupling between the jet and the layer just above the canopy. This behavior is in line with the findings of Banta et al. (2003), who observed that TKE tends to increase as R_{i_bulk} decreases, supporting the interpretation that lower R_{i_bulk} values are associated with sustained turbulence.

In summary, the wind speed at the beginning of the night was lower on 13 July 2022 than on 7 August 2022, suggesting more stable conditions for the July case. However, the LLJ nose was located at a lower height during the July night, leading to enhanced shear below the jet and thus reduced stability. In addition, radiative cooling

was weaker in July but stronger in August, favoring weaker stratification in July and a very stable regime in August. The net effect of shear production and surface cooling, consistent with the shear capacity concept (Van Hooijdonk et al., 2015), was that the July case remained turbulent, whereas the August case transitioned into a shallow and strongly stable NBL.

3.5 Discussion

Both the case studies and average patterns indicate two types of LLJs at ATTO, and the most important distinction between them regards whether or not they are fully coupled by turbulent mixing throughout their depths. Banta (2008) described these two types of LLJs, one typical of weakly stable boundary layers, where there is turbulent coupling from the surface to the LLJ nose, and one during very stable boundary layer, where turbulent mixing is restricted to a layer much shallower than LLJ height.

In the fully coupled LLJ observed at ATTO, finite TKE and heat flux exist from the surface up to the LLJ nose and the temperature gradient is maximum just below that height. The LLJ wind maximum is generally smaller than in the decoupled case, but TKE is larger. These cases are more common under reduced radiative loss, which typically occur when there are clouds in the region. The analysis also indicates that they occur more often when the wind direction is from N-NE. The reduced radiative loss favors a weakly stable boundary-layer, with continuous turbulence throughout its depth, as confirmed by TKE and heat flux profiles. The maximum temperature gradient forms just below the LLJ nose because the heat flux goes from negative below to zero at that height, maintaining cooling below while the layer above remains warmer, producing a local maximum in the temperature gradient even though the heat-flux divergence is not largest there. This is particularly evident for N-NE jets in the class of lowest height (Figure 3-5b and Figure 3-6a). For these reasons, LLJ height is a good indicator of NBL depth in these coupled conditions.

The decoupled LLJ occurs when the radiative loss is larger, as typically occurs in clear sky nights. At ATTO, these are more common when the wind is from S-SE. In such cases, the radiative loss at the surface favors enhanced local cooling, establishing a larger temperature gradient at lower levels and a very stable boundary layer. Turbulent mixing is smaller than in the coupled cases, as confirmed by the smaller magnitudes of TKE and heat flux. More importantly, in this case, the NBL

height, which is where such turbulent quantities decrease in magnitude to zero, occurs at much lower heights than the LLJ nose. There is no cooling by turbulent flux divergence near the LLJ nose, so that the vertical temperature gradient is not affected by this mechanism at those heights. This pattern follows a classical LLJ (Businger, 1973; Banta et al., 2007), in which the wind speed accelerates above the NBL because of reduced drag at levels where the momentum sink associated with the turbulent Reynolds stress is inexistent. For these reasons, in decoupled LLJs, nose height is not a good indicator of NBL depth.

The occurrence of nights with contrastingly different turbulence regimes has been reported at ATTO (Oliveira et al., 2018; Mendonça et al., 2023), but it is still not clear what determines the difference. The present analysis indicates that differences in cloudiness associated with different wind directions may be important. The coupled LLJ, whose height matches NBL height and is common in weakly stable boundary layers, tends to occur with wind directions from N-NE and enhanced cloud cover. In contrast, the decoupled case, where LLJ nose is well above the NBL top, is more common with wind from the S-SE and clear skies. In Section 4, it has been shown that the coupled LLJ case observed in 13 July initiated almost at the same time of an enhanced radar echo at all levels (Figure 3-11), and this fact suggests that the LLJ may be related to the passage of a convective event at the region. To test the hypothesis that coupled LLJs may be typically triggered by nearby convection, we performed an analysis of satellite imagery at nights of LLJs identified in this study. The results are shown in Table 3-3, where cases “with deep convection” refer to those when convection was obviously present at ATTO during the night and “ambiguous” refers to cases when there was convection nearby, but it is unclear whether it affected ATTO directly. Most LLJs in nights with winds from N-NE occurred when there was convection at the site or nearby (75.6% if cases with nearby convection are included). On the other hand, 44.3% of the LLJs with S-SE winds occurred when there was no deep convection of any type at the region.

The results in Table 3-3 indicate a tendency of N-NE LLJs occurring when there is convection at the region, that is further supported by the fact that they tend to occur when there is small radiative loss, indicative of enhanced cloud cover. At the same time, almost half of the LLJs from S-SE are with no convection at the region. Although there is a correspondence between wind direction and cloud cover, there are cases when such fails, meaning N-NE LLJs with no deep convection or S-SE LLJs with deep

convection. This is also supported by some of our previous analysis, such as the finding that low LLJs are usually connected, regardless of the wind direction. We propose, therefore, that when there is no nearby convection, which more commonly occurs with S-SE winds, the clear skies and enhanced radiative loss help establishing a very stable boundary layer, in which LLJs are initiated by the classical mechanism of decoupling of the surface and wind acceleration aloft in the absence of drag effects. The accelerating mechanism is not clear, and the possibilities are horizontal pressure gradient or large scale momentum advection. On the other hand, convection nearby, which more commonly occurs with winds from N-NE, helps triggering coupled LLJs.

Table 3-3. Overview of low-level jet (LLJ) occurrences, grouped by wind directions (N-NE and S-SE), reporting the total number of LLJ events and the occurrence of convective activity for each direction.

Wind direction	Total LLJ events	With deep convection	No deep convection	Convection nearby
N-NE	45	24 (53.4%)	11 (24.4%)	10 (22.2%)
S-SE	61	21 (34.4%)	27 (44.3%)	13 (21.3%)

We can only speculate on the mechanisms through which such triggering occurs. Possibilities include triggering by convectively driven perturbations adjacent to nearby convective systems and by cloud-scale downdrafts underneath storms or developing deep convection. In the former, the sinking motion field and attendant adiabatic warming that often trail deep convective systems (most commonly in mesoscale convective systems) can establish mesoscale low pressure, high wind zones in their wake and flanks. These so-called “wake lows” (e.g., Houze Jr 2004) can typically initiate periods of sustained locally strong, turbulent winds in areas adjacent to deep convection. In this scenario, LLJ initiation may occur even when deep convection passes near the ATTO site without directly impacting it. The latter possibility refers to near-ground intrusion of localized downdrafts from sustained or even developing storms passing at the ATTO site. This possibility implies that kinematic and thermodynamic perturbations produced by transient convective cells can locally disturb the structure and short-term evolution of the lower boundary layer sufficiently to trigger LLJs, which can later be maintained by nocturnal boundary layer dynamics in a weakly stable environment.

In summary, the N-NE LLJs observed at ATTO are likely triggered by nearby convective activity, such as downdrafts or mesoscale wake lows, and are subsequently maintained in a weakly stable boundary layer. In contrast, S-SE LLJs typically develop under clear skies with strong radiative cooling, and may follow the classical decoupling mechanism with wind acceleration above the very stable surface layer, possibly influenced by large-scale pressure gradients or momentum advection. These contrasting mechanisms require further investigation in future studies.

3.6 Conclusions

We investigated the height of the NBL over the Amazon rainforest to assess whether mean wind speed profiles can provide reliable estimates of NBL height (h_N) during nights with the occurrence of Low-Level Jets (LLJs). The LLJ nose height was used to estimate h_N (referred to as h_{N_jet}), and compared against h_{N_flux} , which was derived from turbulent profiles.

The accuracy of h_{N_jet} estimates showed a dependence on thermal stability. Unlike the LLJs from the N-NE direction, the S-SE jets were associated with stronger thermal stratification above the canopy, which likely suppressed turbulent mixing and resulted in a decoupled boundary layer structure, contributing to the difference between h_{N_jet} and h_{N_flux} estimates. Radiative cooling likely contributed to these differences, with more negative net radiation and warmer potential temperature observed under S-SE conditions. Interestingly, regardless of wind direction, LLJs developed within a thermally stratified environment, below the inversion layer.

Overall, our results show that the LLJ nose height can be used as a proxy for the NBL height when turbulent profiles are not available. However, the reliability of this proxy depends on the thermal stability of the boundary layer, which can be assessed from the temperature profile. When the vertical temperature gradient is maximum just below the LLJ nose, turbulent mixing below the jet remains coupled to the surface, making the LLJ nose height a good indicator of the NBL depth. In contrast, when the strongest temperature gradient is located near the surface, a very stable and decoupled boundary layer develops, and the LLJ nose height may overestimate the NBL height.

The high vertical and temporal resolution of the sonic anemometers at ATTO site enabled the detection of LLJs at lower levels (e.g., 81-100 m), which radiosondes

typically used in previous studies failed to detect. However, two observational aspects should be noted. First, because the observations extend up to 300 m, the dataset does not include jets occurring above this height. Such higher-level jets likely arise from different dynamical regimes and thus do not affect the interpretation of the cases analyzed here. Second, the measurements were limited to a single fixed tower. Investigating the mechanisms behind LLJ formation would require spatial coverage (e.g., multiple towers) to resolve horizontal pressure/temperature gradients. Complementary numerical modeling studies could help explore these mechanisms under different spatial scales and atmospheric conditions.

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Atmospheric patterns associated with blowdowns and long-term trends in extreme winds in the Central Amazon

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Abstract

Convective storms in the Amazon can produce intense downdrafts, leading to significant forest disturbance, including tree mortality (blowdowns). However, observational data in remote regions such as the Amazon are limited, restricting detailed studies of atmospheric drivers of such events. In this work, we investigate whether the ERA5 reanalysis captures the thermal and dynamical patterns associated with a well-documented blowdown that occurred between 16 and 18 January 2005 in the Central Amazon. Using hourly ERA5 data, we analyzed wind speed and direction at 10 m and 100 m, 2-m air temperature, and convective available potential energy (CAPE), as well as wind speed at multiple pressure levels, focusing on simultaneous changes indicative of downdrafts. Our results show that ERA5 reproduces the main behavior of these variables during the event: wind intensification, shifts in direction, surface cooling, and elevated CAPE. Focusing on long-term trends, we examined extreme 10-m wind speeds (99th percentile) from 1985 to 2020, revealing a reversal in trends around 2005, with earlier changes during the wet (DJF) and transitional (MAM) seasons. These findings indicate that environments favorable to downdrafts may be becoming more frequent. Despite the limitations of ERA5 resolution and the restricted grid points analyzed, this work demonstrates the potential of ERA5 for investigating local convective patterns and supporting future assessments of forest blowdown events in the Amazon.

Keywords: Amazon Forest, Extreme wind, Forest disturbance, Windthrow, Convective gusts.

4.1 Introduction

Convection plays a fundamental role in determining the climate and weather patterns in the Amazon region (Fisch et al., 1998). Convective systems transport heat and are an important mechanism for redistributing excess energy to regions with energy deficits. (Machado, 2000). This process also influences the regional hydrological cycle and global atmospheric circulation, playing a key role in regulating the Earth's climate.

Storms associated with deep convection can produce substantial precipitation, lightning, and strong wind gusts (Albrecht et al., 2011, Feng et al., 2023). In the Amazon, several studies have shown the occurrence of convective systems, including an aligned cluster of deep convective cells known as squall lines (Cohen et al., 1995; Garstang et al., 1998, Gonçalves et al., 2024, among others). These systems are often accompanied by intense downdrafts, which develop due to evaporative cooling of precipitation. When downdrafts reach the surface, the descending air currents generate a horizontal outflow of cold air (or cold pools), which can cause significant damage to vegetation and urban infrastructure (Fujita, 1985; Wakimoto, 2015).

Intense convective gusts can result in significant ecological impacts, such as tree mortality, commonly referred to as blowdowns or windthrow, and have been widely documented in the Amazon region (Nelson et al., 1994; Marra et al., 2014). Negrón-Juárez et al. (2010) provided a study linking convective systems to forest disturbances. Using Landsat imagery and ground-based forest inventories, they analyzed a squall line that propagated from southwest to northeast across the Amazon Basin between 16 and 18 January 2005. The authors found that changes in non-photosynthetic vegetation (ΔNPV), which served as an indicator of tree mortality, were strongly correlated with field-measured tree mortality, demonstrating that the observed blowdown resulted from this specific convective event. Negrón-Juárez et al. (2010) established a clear connection between squall lines, downdrafts, and forest damage, highlighting the vulnerability of Amazon trees to wind-driven mortality.

Subsequently, Urquiza-Muñoz et al. (2024) analyzed large-scale windthrows (≥ 30 ha) across the Amazon from 1985 to 2020 using satellite data. They found a significant increase in both the number of events (from 50 in 1985 to 205 in 2020) and the total affected area. They also observed an increase in the 95th percentile of afternoon convective available potential energy (CAPE) over the 35-year period,

suggesting that environments favorable to severe convective systems have intensified regionally.

Downdrafts have been associated with characteristic deviations in meteorological variables measured at the surface. In the Central Amazon, Garstang et al. (1998) observed simultaneous increase in wind speed and atmospheric pressure, accompanied by temperature drops during the occurrence of gusts propagated by convective storms. Furthermore, Dias-Júnior et al. (2017), analyzed observational data and showed that downdrafts caused a reduction in CAPE and equivalent potential temperature. However, observational meteorological data are limited in remote areas such as the Amazon, making it difficult to study the spatial and temporal characteristics of extreme wind events. Reanalysis products like ERA5 provide a practical alternative, allowing the investigation of mesoscale atmospheric patterns associated with convective systems, despite their limited spatial and temporal resolution.

In this study, we investigated whether ERA5 can reproduce the thermal and dynamical patterns of downdrafts associated with a well-documented blowdown event in the Central Amazon. We first analyze the 16–18 January 2005 event (Negrón-Juárez et al., 2010) as a case study to assess if ERA5 reproduces characteristic patterns of wind, temperature, and CAPE during the passage of a squall line. Additionally, we extend our analysis to examine long-term trends in extreme 10-m wind speeds over the same region from 1985 to 2020, aiming to provide insights into potential links between changes in wind speed and increasing windthrow occurrence.

4.2 Study site and data acquisition

4.2.1 Reanalysis data

We used data from the fifth-generation atmospheric reanalysis ERA5, produced by the European Centre for Medium-Range Weather Forecasts (ECMWF) (Hersbach et al., 2023). ERA5 provides hourly estimates of atmospheric variables from 1940 to the present by combining numerical model output with a wide range of observations through a data assimilation system. Previous evaluations indicate that ERA5 surface wind fields exhibit higher agreement with in situ observations than those from other reanalysis products (Ramon et al., 2019).

For this study, ERA5 data were obtained on a regular latitude-longitude grid with a horizontal resolution of 0.25° (~ 31 km) and temporal resolution of 1 hour. Two groups of variables were considered:

- i) Single-level variables: including the eastward (u) and northward (v) components of the wind (ms^{-1}) at a 10 m and 100 m height, 2-m air temperature (T_2 , kelvin), mean sea level pressure (P_s , hPa), and convective available potential energy (CAPE, Jkg^{-1});
- ii) Pressure-level variables: including u and v wind components, and temperature (T , kelvin) at five pressure levels (1000, 975, 950, 925, and 900 hPa).

The horizontal wind speed and direction were derived from the u and v components as $V = (u^2 + v^2)^{0.5}$ where V_{10} and V_{100} represent the wind speed at 10 m and 100m, respectively. It is important to note that, like most global reanalysis products, ERA5 may have limitations in representing near-surface winds over complex terrain or forests due to model topography, boundary-layer parameterizations, and its horizontal resolution (~ 31 km) (Zeng et al., 2019). Therefore, our analysis focuses on the behavior and temporal evolution of the variables rather than their absolute magnitudes.

4.2.2 Site

Negrón-Juárez et al. (2010) analyzed Landsat imagery (scene P231 R062, covering approximately $3.4 \times 10^4 \text{ km}^2$) to quantify changes in ΔNPV , which served as an indicator of tree mortality. Based on the identified blowdown area from ΔNPV , forest inventory plots were established along the disturbance gradient to directly quantify tree mortality in the field. A strong relationship observed between Landsat ΔNPV and field-measured tree mortality data, confirming that the imagery accurately represented areas affected by the blowdown.

Figure 4-1a shows the location of the Landsat scene within the Amazon Basin, while a closer view of the blowdown area is provided in Figure 4-1b, including the ERA5 grid spacing ($\sim 0.25^\circ$, yellow lines). The central point of the blowdown area (centered at 2.6°S and 60.3°W , red point in Figure 4-1b), reported by Negrón-Juárez et al. (2010), falls near the edge of the ERA5 grid cell. To account for the spatial

resolution of the reanalysis, we selected the four grid points surrounding the reported location (latitude = 2.5 °S, 2.25 °S, and longitude = 60.25 °W, 60.5 °W) to represent the atmospheric conditions associated with the event. These selected grid points were used to extract ERA5 data for the case study period (16-18 January 2005) and the long-term analysis of extreme 10-m wind speeds from 1985 to 2020. This approach ensures consistency between the spatial domain of the reanalysis and the location of the observed blowdown.

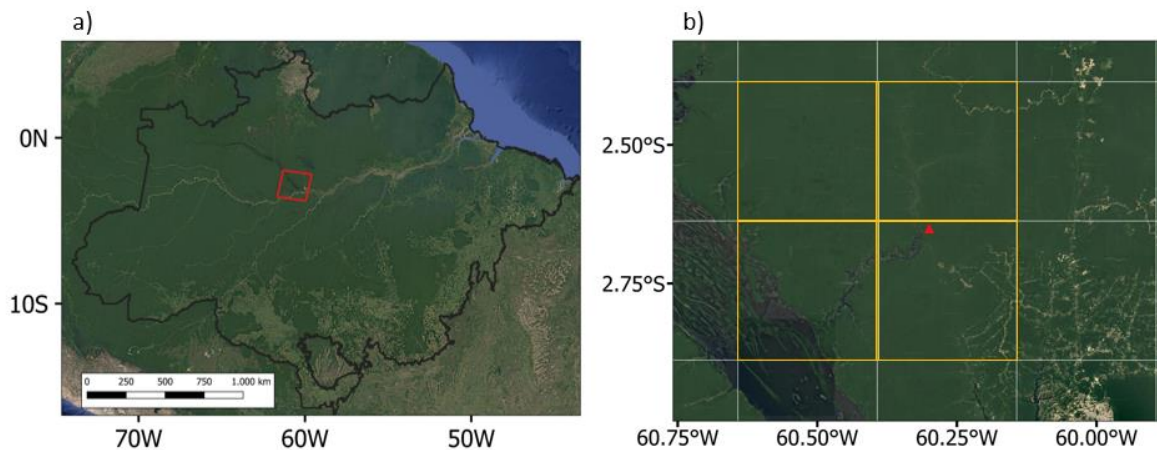


Figure 4-1. Study site and ERA5 grid selection. (a) The location of the Landsat image is illustrated by the solid line square, (b) Blowdown region (within Landsat scene), showing the central coordinates of the observed tree mortality (red point, 2.6°S, 60.3°W) and the surrounding ERA5 grid points (0.25° resolution) used to extract atmospheric data.

4.3 Methodology

4.3.1 Atmospheric conditions

To assess whether ERA5 captures atmospheric patterns associated with the convective system responsible for the documented forest damage, hourly ERA5 data were analyzed, including wind speed and direction, air temperature, and CAPE. Particular attention was given to simultaneous changes in these variables indicative of the occurrence of downdrafts, similar to those described by Garstang et al. (1998) and Betts et al. (2002).

4.3.2 Trend analysis

Extreme wind events were defined using the 99th percentile (P99) of the 10 m wind speed distribution, which provides a robust characterization of the upper tail while retaining a sufficient sample size for statistical analysis. The long hourly record (1985-2020) ensures a robust estimation of the 99th percentile. Subsequently, trend analyses were performed at both annual and seasonal scales. The annual P99 of the 10-m wind speed (hereafter P99V10) were calculated for each year from 1985 to 2020. Seasonal P99V10 values were also computed for December, January and February (DJF), March, April and May (MAM), June, July and August (JJA), and September, October and November (SON). Seasons were defined following the climatological classification commonly adopted for the Amazon Basin (e.g., Espinoza et al., 2009). DJF represents the wet season, JJA the dry season, and MAM and SON the transition seasons.

Trends in annual and seasonal P99V10 time series were assessed using the nonparametric Mann-Kendall (MK) test and Sen's slope estimator. The MK test is suitable for non-normally distributed and for testing trends in non-linear time series, such as wind speed, while Sen's slope quantifies trend magnitude and uncertainty. Trend significance was evaluated at the 95% confidence level ($p < 0.05$). Segmented regression (R segmented package) was used to identify potential breakpoints in trend behavior.

4.4 Results and Discussion

4.4.1 Atmospheric conditions associated with LI

Figure 4-2 shows the horizontal wind speed (V) and wind direction (WD) at 10 m and 100 m for the three analyzed days (16-18 January 2005) in UTC (local time = UTC-4). On 17-Jan at 15 UTC, V at both levels were close to 1 ms^{-1} , increasing within the following two hours to approximately 3.0 ms^{-1} at 10 m and 4.3 ms^{-1} at 100 m (Figure 4-2b). The amplitudes of V10 and V100 were 2.5 ms^{-1} and 3.8 ms^{-1} , respectively. During the same period, WD shifted from easterly–northeasterly ($50\text{--}100^\circ$) to northwesterly ($\sim 300^\circ$) at both heights, indicating a rapid reorganization of the low-level flow associated with convective activity.

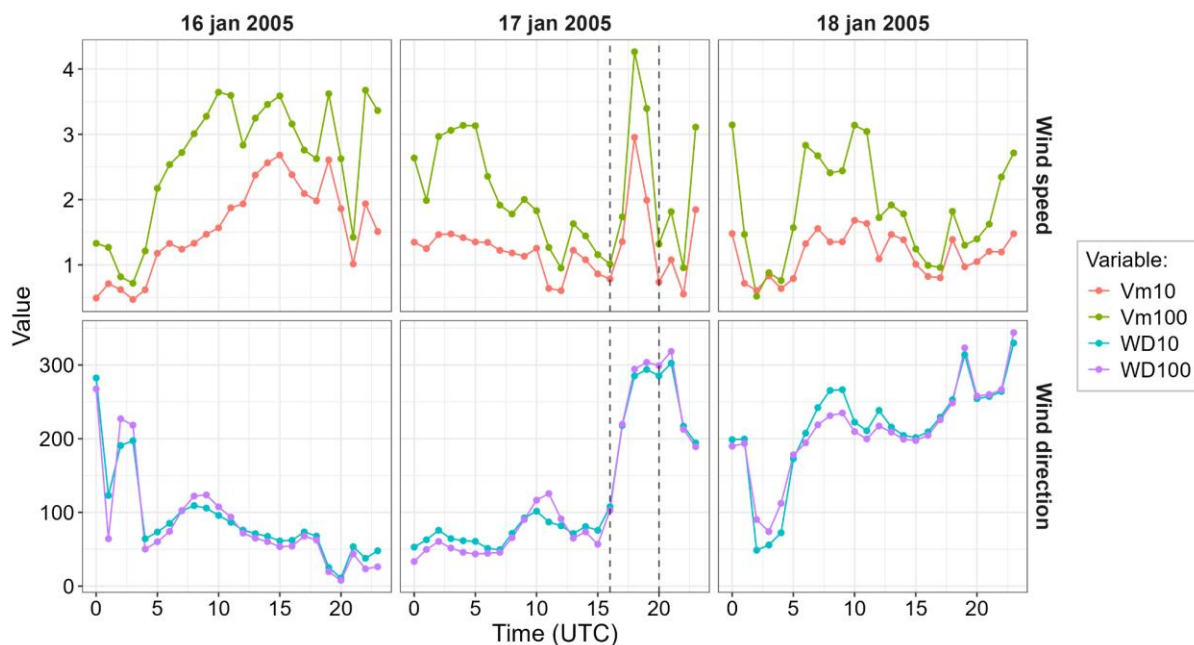


Figure 4-2. Hourly wind speed (V_m , ms^{-1}) and wind direction (WD , $^\circ$) from ERA5 at 10 m and 100 m for 16-18 January 2005. Colors indicate different heights. The dashed vertical line marks 16:00 UTC and 20:00 UTC intervals.

A similar temporal evolution was identified across five pressure levels (1000-900 hPa; Figure 4-3). While the timing of the wind intensification and directional shift was consistent throughout the lower troposphere, wind speed amplitudes increased with height, ranging from 2.4 ms^{-1} at 1000 hPa to 9.0 ms^{-1} at 900 hPa. This vertical gradient indicates the presence of enhanced low- to mid-level vertical wind shear, a key dynamical ingredient for the organization and maintenance of convective systems.

The 2-m air temperature (T_2) values exhibited a marked cooling during the afternoon hours (16-20 UTC; Figure 4-4a), particularly on 17-Jan, when T_2 decreased by approximately 6 K. On 16-Jan, the temperature drop reached about 3 K, whereas on 18-Jan the decrease was limited to about 1 K. The pronounced cooling on 17-Jan is consistent with the passage of a convective system and associated cold pool and downdrafts, which transport cooler and denser air from the mid-troposphere toward the surface. In addition, T_2 on 16 and 17 January was 2-4 K higher than on 18-Jan, suggesting that the convective activity on 17-Jan contributed to a cooler atmosphere during the subsequent day.

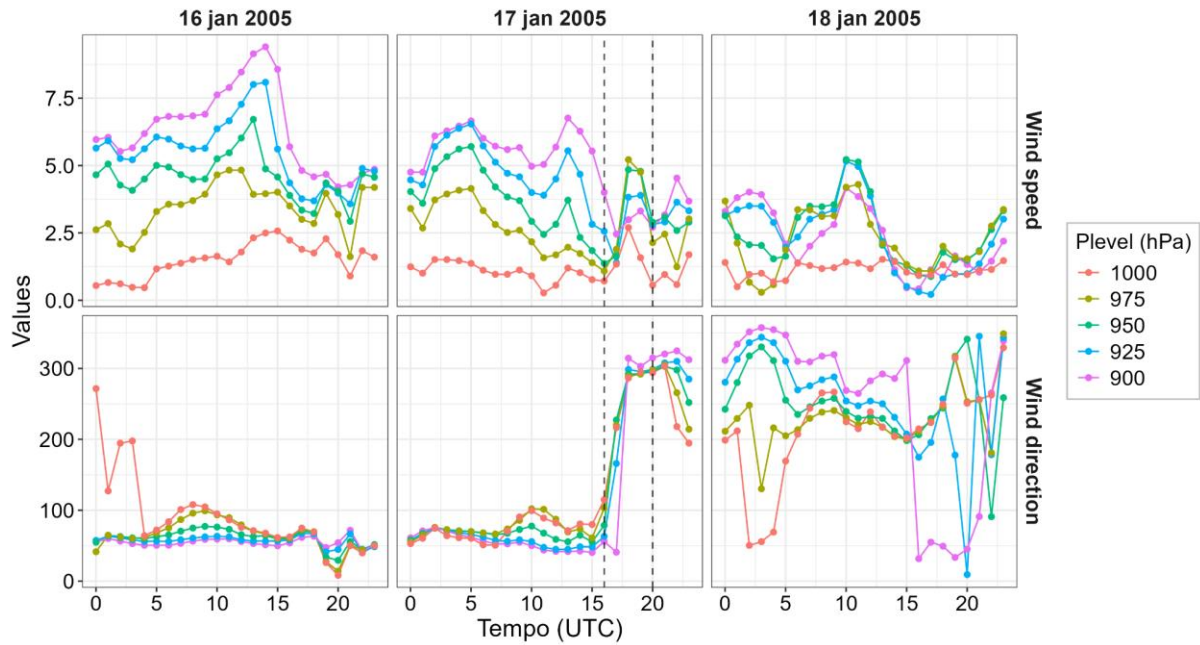


Figure 4-3. Hourly wind speed (ms^{-1}) and wind direction (degrees) from ERA5 at multiple pressure levels (1000, 975, 950, 925, and 900 hPa) for 16–18 January 2005. Colors indicate different pressure levels. The dashed vertical line marks 16:00 UTC and 20:00 UTC intervals.

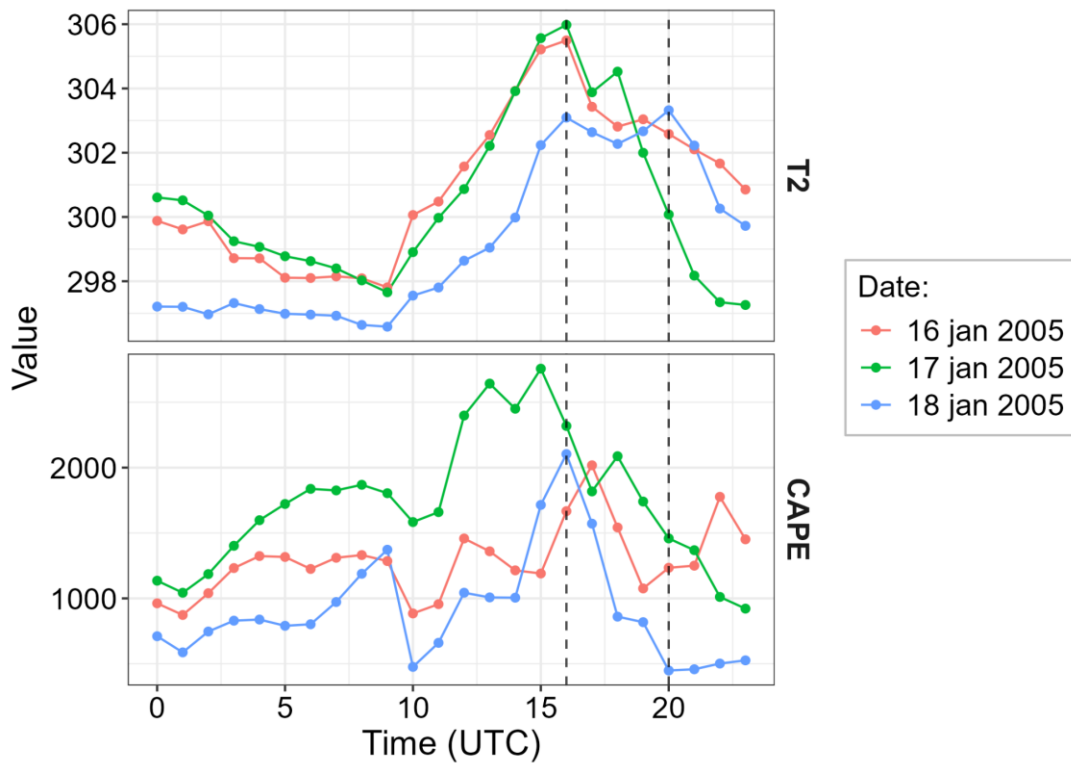


Figure 4-4. Hourly convective available potential energy (CAPE, Jkg^{-1}) and 2-m air temperature (T2, K) from ERA5 for 16–18 January 2005. The dashed vertical line marks 16:00 and 20:00 UTC intervals.

Thermodynamic conditions further support this interpretation. CAPE reached its highest value on 17-jan, with a maximum of 2757 Jkg^{-1} at 15 UTC (Figure 4-4b), compared to lower maxima on 16-jan (2018 Jkg^{-1} at 17 UTC) and 18-jan (2104 Jkg^{-1} at 16 UTC). Elevated CAPE prior to the event indicates a strongly unstable environment favorable for deep convection and intense downdrafts, consistent with the observational findings of Dias-Júnior et al. (2017) over the Amazon region.

Although the absolute magnitudes differ from those reported in observational studies, the combined occurrence of wind intensification, wind direction shift, high pre-convective CAPE, and surface cooling, suggests that ERA5 captures the convective thermodynamic evolution previously reported in observational studies (Dias-Júnior et al., 2017, Silva et al., 2026).

4.4.2 Temporal trends in wind speed

The annual 99th percentile (P99) of 10-m wind speed (P99V10) showed a change in trend over the study period (Figure 4-5). From 1985 to approximately 2005, P99V10 decreased at a rate of $-0.10 \text{ ms}^{-1} \text{ decade}^{-1}$, although this slope was not statistically significant ($p > 0.1$). Segmented regression identified a breakpoint in 2005, after which P99V10 increased significantly at a rate of $0.30 \text{ ms}^{-1} \text{ decade}^{-1}$ ($p < 0.05$). This increase was approximately three times larger than the magnitude of the decline observed prior to 2005. The Mann-Kendall test applied to the period after the breakpoint indicated a significant positive trend ($\tau = 0.43$, $p = 0.02$).

The decline in wind speed, known as the period of “stilling”, has been replaced by a reversal at the global scale, as reported by Zeng et al. (2019). Using direct in situ observations of near-surface wind speed between 1978 and 2017, they found a wind speed reversal around 2010, followed by an increase about three times larger than the preceding decline. Although the present analysis examines extreme winds (P99) over the Amazon and Zeng et al. (2019) analyzed data from mid-latitude regions of the Northern Hemisphere, the similarity in the timing and relative magnitude of the reversal suggests a common large-scale control.

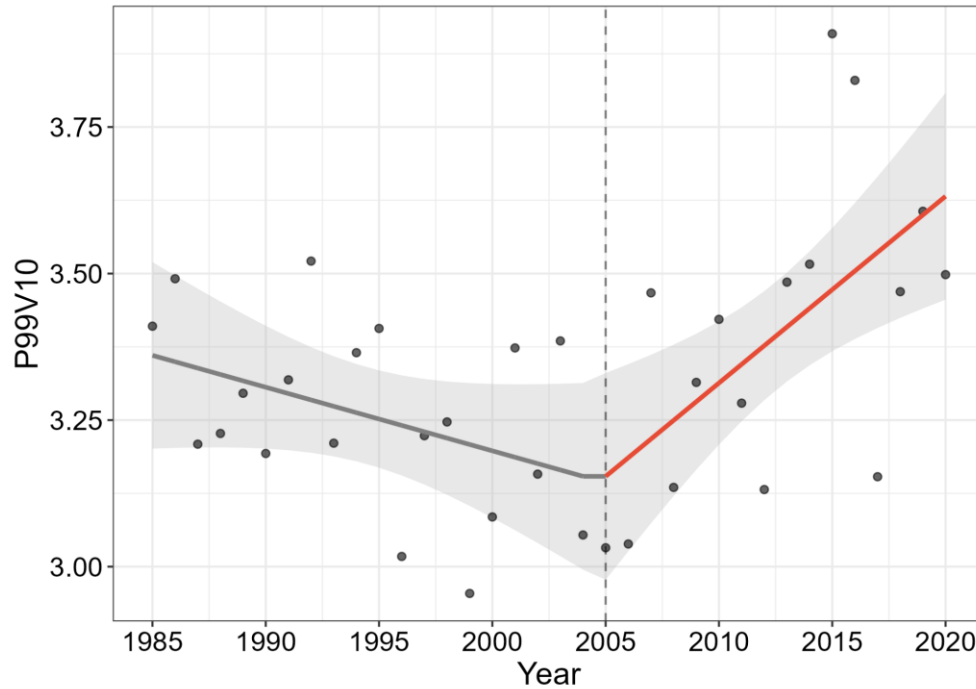


Figure 4-5. Annual 99th percentile (P99) of 10-m wind speed (P99V10) during 1985-2020 (black dots). The slope of the segmented regression is shown in red if statistically significant ($p < 0.05$) and in grey otherwise. Shaded areas denote the 95% confidence intervals, and the vertical dashed line marks the 2005 breakpoint.

The seasonal evolution of the annual P99V10 and the corresponding segmented regression for DJF, MAM, JJA, and SON are shown in Figure 4-6. In DJF and MAM, a breakpoint was detected around 1999. After this point, both seasons exhibited a significant positive trend according to the Mann-Kendall test ($\tau = 0.36$, $p = 0.02$ and $\tau = 0.34$, $p = 0.02$, respectively), shown as red lines in Figure 4-6a and Figure 4-6b.

In JJA, the breakpoint occurred around 2006, followed by a significant positive trend ($\tau = 0.47$, $p = 0.02$, Figure 4-6c). However, the corresponding regression slope in JJA was not statistically significant, suggesting a consistent increase over time, but with an uncertain magnitude. In SON, although a breakpoint was identified around 2006, no significant trend was detected after this transition ($\tau = 0.23$, $p = 0.23$, Figure 4-6d). Before each breakpoint, all seasons showed decreasing trends, but none were statistically significant ($\tau = -0.01$ for DJF, -0.25 for MAM, -0.18 for JJA, and -0.04 for SON). Linear slopes ($\text{ms}^{-1} \text{ decade}^{-1}$) before and after the breakpoint are reported in Table 4-1, where only DJF and MAM were statistically significant after the breakpoint.

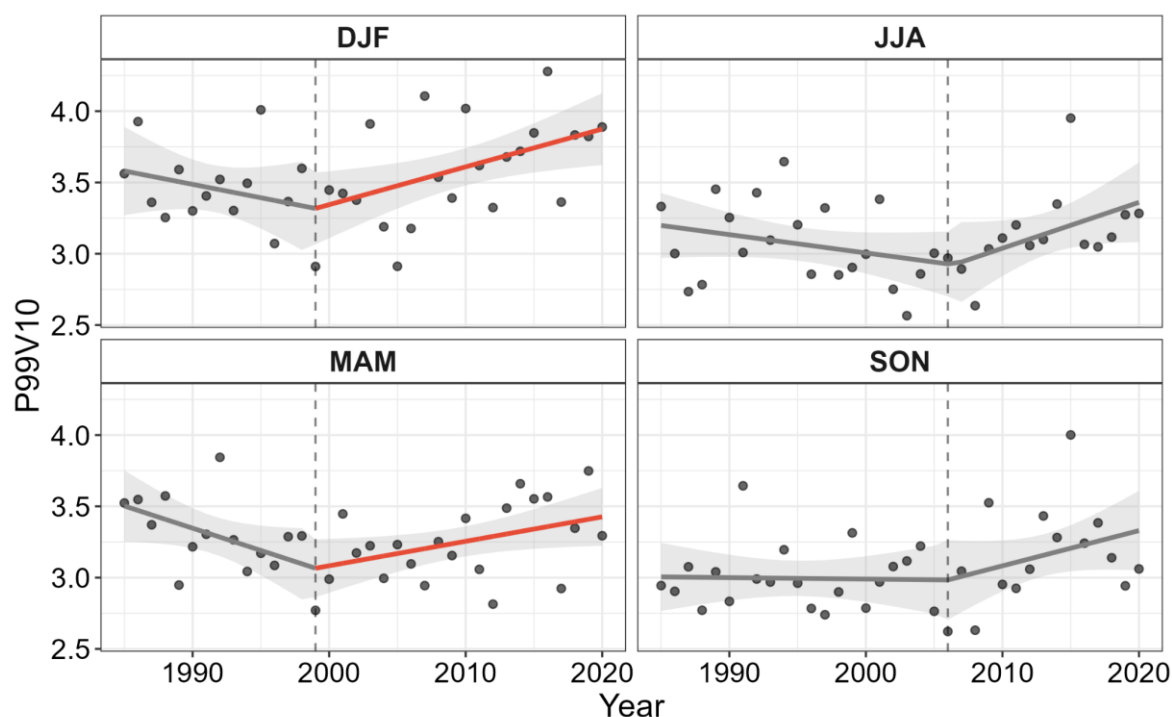


Figure 4-6. Same as Figure 4-5, but for seasonal P99V10: (a) DJF (December-February), (b) MAM (March-May), (c) JJA (June-August), and (d) SON (September-November).

Interestingly, the seasonal breakpoints in DJF and MAM occurred around 1999, earlier than the annual breakpoint in 2005, while JJA and SON exhibited breakpoints closer to the annual value. This suggests that the increase in extreme wind events first became more pronounced during the wet season (DJF) and the transition from wet to dry season (MAM). This period coincides with the seasonal southward shift of the Intertropical Convergence Zone (ITCZ), when convective storms and associated gust fronts are more frequent over the Amazon (Cohen et al., 1995). These seasonal differences highlight the role of convective processes in modulating extreme winds in the Amazon.

Overall, the seasonal and annual analyses of P99V10 highlight the occurrence of a reversal in extreme wind trends in the Amazon, with earlier breakpoints during the wet (DJF) and transitional (MAM) seasons compared to the annual signal. While these results are consistent with the global-scale reversal of near-surface wind trends reported by Zeng et al. (2019), it is important to note that ERA5, like most reanalysis products, may have limitations in representing near-surface winds over land due to model topography and boundary-layer parameterizations.

Table 4-1. Annual and seasonal trends (1985-2020) in the 99th percentile (P99) of 10-m wind speed derived from ERA5. The table reports the mean P99 $\pm$ standard deviation (SD), minimum (Min) and maximum (Max) P99 values, the estimated breakpoint (Bp) from the segmented regression, and the linear slope before and after the breakpoint. DJF: December–February; MAM: March–May; JJA: June–August; SON: September–November. ** denotes statistical significance of the slope at the 5% level ($p < 0.05$).

	Mean P99 $\pm$ Sd (ms ⁻¹)	Min P99 (ms ⁻¹)	Max P99 (ms ⁻¹)	Breakpoint (Bp)	Slope before Bp (ms ⁻¹ decade ⁻¹)	Slope after Bp (ms ⁻¹ decade ⁻¹)
Annual	3.31 $\pm$ 0.21	2.95	3.90	2005	- 0.10	0.31**
DJF	3.54 $\pm$ 0.32	2.91	4.27	1999	-0.18	0.26**
MAM	3.26 $\pm$ 0.26	2.77	3.84	1999	-0.31	0.17**
JJA	3.09 $\pm$ 0.28	2.56	3.95	2006	-0.12	0.32
SON	3.06 $\pm$ 0.28	2.62	4.00	2006	-0.01	0.24

4.5 Conclusions

Using a case study, this work shows that ERA5 was able to capture the main thermal and dynamical patterns associated with the 16-18 January 2005 blowdown in the Central Amazon, including wind intensification, shifts in wind direction, surface cooling, and high CAPE. It is worth noting that the results focus on behavior rather than absolute values. Analysis of extreme 10-m wind speeds (99th percentile) from 1985 to 2020 revealed a reversal in trends around 2005, with earlier changes observed during the wet (DJF) and transitional (MAM) seasons. These findings indicate that ERA5 can provide valuable information on convective patterns at the local scale and support the potential link between changes in atmospheric conditions and increasing forest blowdown occurrence in the Amazon. Together, the case study and long-term trends highlight how environments favorable to downdrafts and strong gusts may be becoming more frequent.

We recognize the limitations of the reanalysis data, particularly the spatial (0.25°, ~31 km) and temporal resolution of ERA5. In addition, analyses were performed

for a limited set of grid points corresponding to the blowdown area, which may restrict the generalization of the findings to the wider Amazon region. These limitations highlight the need for complementary in situ observations to further evaluate and validate ERA5 in capturing convective patterns.

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CONSIDERAÇÕES FINAIS

A presente tese demonstra que a organização da Camada Limite Noturna sobre a floresta Amazônica é influenciada pela estabilidade atmosférica, pela rugosidade da superfície e pela atuação de forçantes atmosféricas de maior escala. Esses fatores condicionam a intensidade da turbulência, o estado de acoplamento entre o dossel e a atmosfera e, conseqüentemente, a variabilidade da CLN. Em conjunto, os resultados obtidos ampliam a compreensão dos processos físicos responsáveis pela organização da atmosfera noturna amazônica.

A partir dessa perspectiva, foi possível demonstrar que a variabilidade da altura da CLN está associada aos diferentes regimes turbulentos e às características da superfície. A utilização de medições turbulentas obtidas no sítio ATTO permitiu estimar a altura da CLN (h_N) em função desses regimes, evidenciando que h_N varia em função da estabilidade atmosférica e da rugosidade da superfície. Da mesma forma, verificou-se que a altura do nariz dos Jatos de Baixos Níveis não representa h_N , de forma universal, uma vez que essa relação depende do estado de acoplamento entre a superfície e a atmosfera. Em condições fracamente estáveis, o jato acompanha a profundidade da CLN. Entretanto, sob forte estratificação térmica, o jato se estabelece acima da CLN, levando à superestimativa de h_N . Esses resultados demonstram que a estrutura da atmosfera noturna amazônica não pode ser interpretada a partir de indicadores isolados, mas deve ser compreendida como consequência da organização dos diferentes regimes turbulentos.

Os *downdrafts* associados à passagem de sistemas convectivos promoveram mudanças abruptas na velocidade e na direção do vento, redução da temperatura e alterações na estrutura turbulenta, favorecendo o acoplamento entre o interior do dossel e a atmosfera acima da floresta. Adicionalmente, verificou-se que os sistemas convectivos capazes de reorganizar a turbulência acima do dossel também estão associados aos ventos extremos responsáveis por eventos de mortalidade de árvores em larga escala. A análise com dados de reanálise mostrou que o ERA5 reproduziu os principais padrões térmicos e dinâmicos associados à passagem da tempestade, embora sua resolução não permita representar com precisão as magnitudes locais do vento. A análise dos valores extremos de vento revelou uma intensificação, especialmente na estação chuvosa, indicando que ambientes atmosféricos favoráveis à ocorrência de *downdrafts* podem estar se tornando mais frequentes.

Embora os resultados obtidos representem avanços importantes na compreensão da atmosfera noturna amazônica, diversas questões permanecem abertas. A integração entre estimativas diretas da altura da CLN e observações de gases-traço, aerossóis e fluxos de carbono poderá ampliar a compreensão do papel da turbulência no funcionamento biogeoquímico da floresta. Da mesma forma, simulações numéricas de alta resolução e a ampliação da análise de eventos extremos utilizando observações e reanálises poderão contribuir para o desenvolvimento de modelos mais robustos dos mecanismos físicos associados à turbulência, aos processos convectivos e à mortalidade de árvores na Amazônia.

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APÊNDICE A: MATERIAL SUPLEMENTAR DO CAPÍTULO 2

Text A1-A2. A factorial ANOVA was conducted to compare measurements taken at 81 m (INSTANT Tower) and 83 m (TALL Tower). The analysis revealed no statistically significant differences between the data ($p > 0.05$).

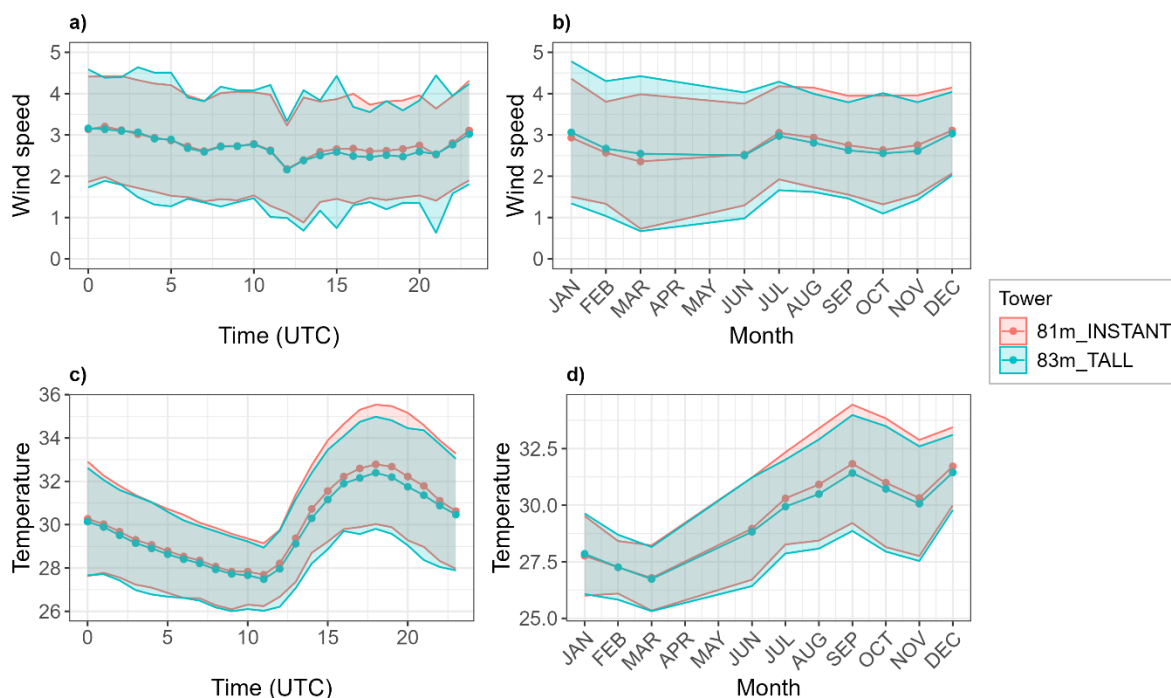


Figure A1. Mean and standard deviation of the hourly (left panels) and monthly (right panels) time series of (a,b) wind speed (ms⁻¹) and (c,d) temperature (°C) for the year 2022 at the ATTO site.

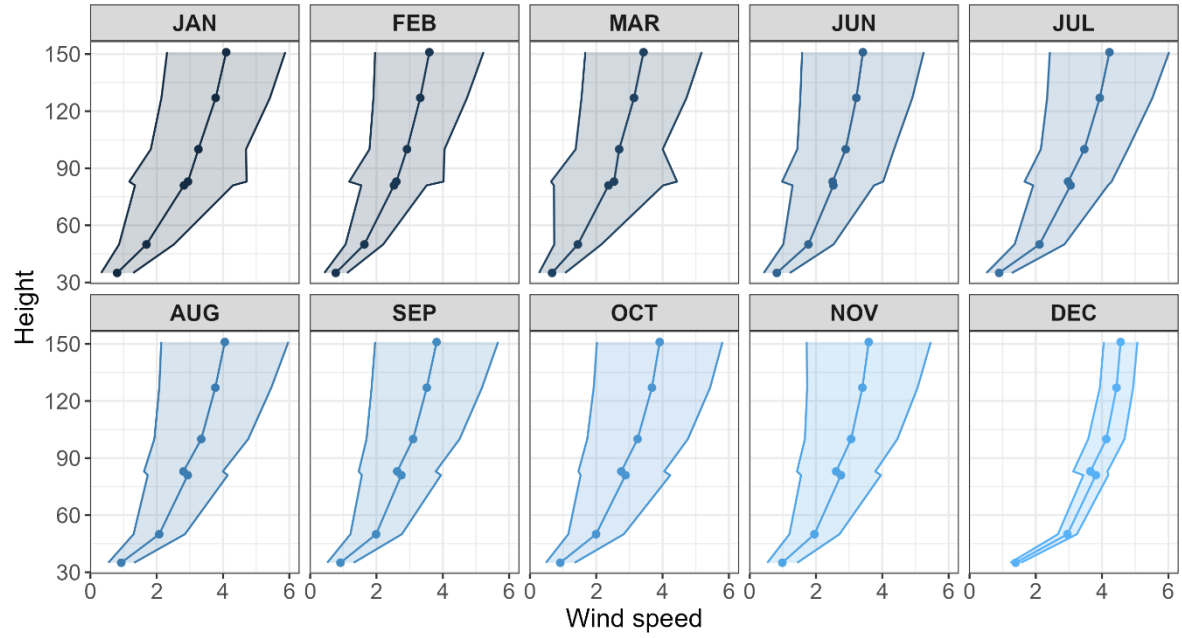


Figure A2. Average monthly profile of wind speed (ms^{-1}) for the year 2022 at the ATTO site

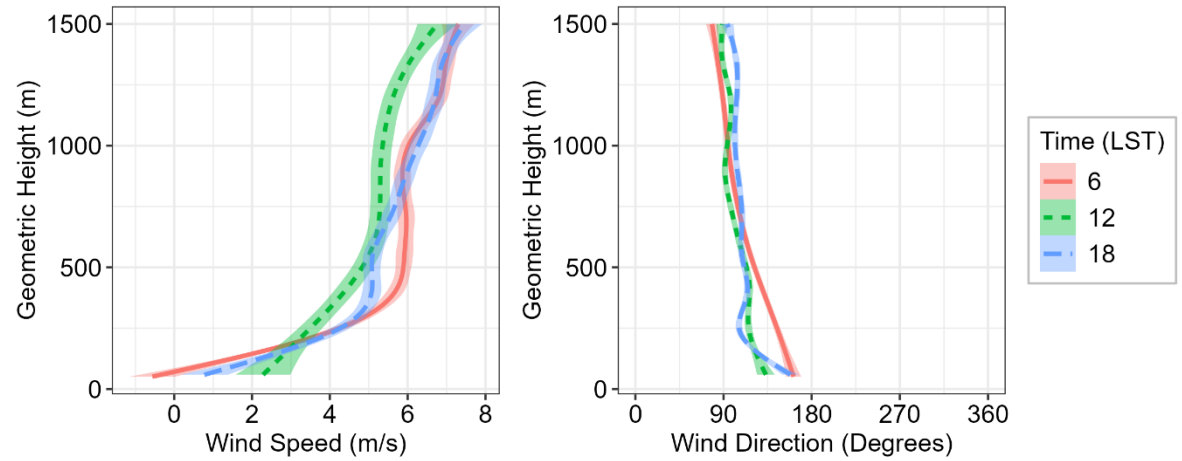


Figure A3. Average wind speed (left panel) and wind direction (right panel) profiles derived from radiosonde measurements taken at 0600, 12:00 and 18:00 LT (Local Time = UTC-4h) during August 9 to 18, 2022. Shaded regions represent the 95% confidence intervals.