A physics-based universal indicator for vertical decoupling and mixing across canopies architectures and dynamic stabilities

Olli Peltola[1], Karl E. Lapo[2], and Christoph K Thomas[2]

[1]Finnish Meteorological Institute
[2]University of Bayreuth

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Abstract

Air flows may be decoupled from the underlying surface either due to strong stratification of air or due to canopy drag suppressing cross-canopy mixing. During decoupling, turbulent fluxes vary with height and hence identification of decoupled periods is crucial for the estimation of surface fluxes with the eddy-covariance (EC) technique and computation of ecosystem-scale carbon, heat, and water budgets. A new indicator for identifying the decoupled periods is derived using forces (buoyancy and canopy drag) hindering movement of a downward propagating air parcel. This approach improves over the existing methods since 1) changes in forces hindering the coupling are accounted for and 2) it is based on first principles and not on ad-hoc empirical correlations. The applicability of the method is demonstrated at two contrasting EC sites (flat open terrain, boreal forest) and should be applicable also at other EC sites above diverse ecosystems (from grasslands to dense forests).

A physics-based universal indicator for vertical decoupling and mixing across canopies architectures and dynamic stabilities

O. Peltola[1], K. Lapo[2,3], C. K. Thomas[2,3]

[1]Climate Research Programme, Finnish Meteorological Institute, P.O. Box 503, 00101 Helsinki, Finland
[2]Micrometeorology Group, University of Bayreuth, Bayreuth, Germany
[3]Bayreuth Center for Ecology and Environmental Research, Bayceer, University of Bayreuth, Bayreuth, Germany

Key Points:

- a universal indicator for air flow vertical decoupling is derived
- the indicator enables analytical estimation of flow decoupling dependency on height, stratification and leaf area index
- the indicator should be applicable at most flux measurement sites, since it relies only on basic micrometeorological measurements

Corresponding author: O. Peltola, `olli.peltola@fmi.fi`

Abstract

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Plain Language Summary

Air flows may be disconnected (i.e. decoupled) from the surface below, meaning that the properties of the flow (e.g. wind speed, temperature, concentrations of gases, pollutants or particles) do not react to changes at the surface. During these periods, air temperatures near the ground decrease and concentrations of gases, pollutants and particles increase significantly since they are not transported upwards, but rather stay close to the ground. These decoupling periods can take place when the air is strongly stratified (e.g. clear-sky, weak wind nights) or thick forest canopies inhibit air mixing. Controls on flow decoupling are poorly understood, yet the phenomenon has significance for scientific monitoring networks and also for the general public due to its connection e.g. to air quality and frost formation. In this study, we derive a new indicator for flow decoupling, demonstrate its applicability at two measurement sites and discuss variables controlling decoupling in the light of this new indicator.

1 Introduction

Understanding of air flows and mixing in the very stable boundary layer (vSBL) often observed e.g. during clear-sky, weak wind nights persists to be incomplete (Mahrt, 2014). This issue poses problems for all scientific studies enquiring into surface-atmosphere interactions including mass and energy budgets, since they rely on turbulence observations or boundary-layer theories, both of which tend to fail under strong stratification.

The stable stratification, resulting from surface cooling via radiative heat loss, suppresses vertical turbulent mixing. Under strong enough stratification and weak turbulence production via wind shear, the turbulent eddies become detached from the surface, i.e. they are not coupled to the surface. This results in so-called ”z-less” scaling of turbulence statistics (Nieuwstadt, 1984), meaning that distance from the surface is no longer a governing length scale (Sorbjan, 2006; Sorbjan & Balsley, 2008; Grachev et al., 2013; Li et al., 2016). As eddies detach from the surface, they lose their immediate connection to the exchange of momentum, heat and gases at the surface resulting in vertical variability of turbulent flux of these constituents with height (Mahrt et al., 2018).

Vertical variability of turbulent flux in this decoupled flow regime poses a severe problem for the global eddy covariance (EC) flux measurement network (FLUXNET) (Baldocchi, 2014) and a clear solution for the problem is lacking (Aubinet et al., 2010). FLUXNET is the main observational tool to study global terrestrial carbon and water cycles and the accuracy of the network largely hinges upon proper identification of decoupled and coupled flow regimes. Only in latter case EC observations integrate over all sinks and sources and thus can provide biophysically meaningful estimates of carbon, water and heat budgets. Accurate estimates of terrestrial carbon cycle are sorely needed for constraining the global carbon budget (Friedlingstein et al., 2019).

Commonly the friction velocity (u_*) is used to identify decoupled periods from continuous flux time series, albeit this approach is known to be flawed, in particular at sites with dense canopies (Acevedo et al., 2009; Thomas & Foken, 2007a; Thomas et al., 2013; Jocher et al., 2018; Freundorfer et al., 2019). Various other metrics have been used to identify the weakly stable from the very stable flow regime (Mahrt et al., 1998; Sun et al., 2012; Williams et al., 2013). However, they all rely on uncertain site specific threshold values and were developed for open areas and hence their applicability to forested regions remains unclear (Freundorfer et al., 2019). Canopy flows differ markedly from the air flows above short vegetation, due to prevalence of coherent flow structures (Raupach et al., 1996; Finnigan, 2000; Thomas & Foken, 2007b; Finnigan et al., 2009) and the momentum sink for the air flow caused by canopy drag. The latter can cause the air flows above forests to be decoupled from the forest floor also during daytime (Kruijt et al., 2000; Thomas et al., 2013; Jocher et al., 2017, 2018; Santana et al., 2018).

In this study we aim to advance the mechanistic understanding of flow coupling to the surface, in particular in the presence of emergent vegetation and/or strong stratification. Here we define a 'weakly stable regime' to be governed by eddies which communicate with the surface (z-scaling applies), whereas in the 'strongly stable regime' the large wall-attached eddies are not prevalent. A simple air parcel technique is used to evaluate the flow coupling to the surface. A novel metric is proposed to identify the flow regime and variables controlling the decoupling are discussed. The metric may be applied across the entire gradients from short canopies (e.g. grass, crop, snow) to dense tall forests and hence applicable at most flux sites monitoring ecosystem-atmosphere interactions.

2 Theory

Coupled air layers are defined in this study as follows: air parcels travel between the coupled air layers and facilitate the exchange of heat, mass and momentum between the layers. Therefore there is a direct interaction between the layers. In contrast, air parcels do not travel between decoupled air layers and hence there is no direct interaction between the layers (albeit waves can still transport momentum). When considering coupling of air layer at height z above ground with the surface, based on this definition there need to be air parcels that can traverse the vertical distance of z. This concurs with the notion that in coupled situations large wall-attached eddies that scale with z dominate the flow (Sun et al., 2012; Lan et al., 2018; Sun et al., 2020). Note that the concept proposed below is based on first principles and does not assume e.g. the surface layer similarity theories to be valid. Similar air parcel approaches have been used (e.g. Mahrt, 1979; Sorbjan, 2006; Sorbjan & Balsley, 2008; Mahrt et al., 2012; Zeeman et al., 2013) to derive e.g. relevant length scales in the stable boundary layer, here it is used in canopy flows to examine the coupled air layer.

Movement of downward moving air parcels at the canopy height (h) is hindered by any opposing forces which include canopy drag caused by the foliage (e.g. Poggi, Katul, & Albertson, 2004; Cescatti & Marcolla, 2004; Watanabe, 2004) and buoyancy force inflicted by stably stratified air layers. In order to reach the ground, an air parcels kinetic energy must match or exceed the work performed against the hindering forces. Based on this a critical speed ($w_{e,crit}$) for the air parcel can be derived (see supporting information):

$$w_{e,crit} = -\gamma \hat{c_d} \mathrm{LAI} U_h - \sqrt{\gamma^2 \hat{c_d}^2 \mathrm{LAI}^2 U_h^2 + 2gh \frac{\theta_e - \hat{\theta}}{\hat{\theta}}}, \tag{1}$$

where γ is a constant (=0.277) depending on the horizontal wind and downward penetrating air parcel speed profiles below canopy height h (e.g. Inoue, 1963; Amiro, 1990a; Poggi, Porporato, et al., 2004; Yi, 2008), $\hat{c_d}$ is the mean drag coefficient below h (equal

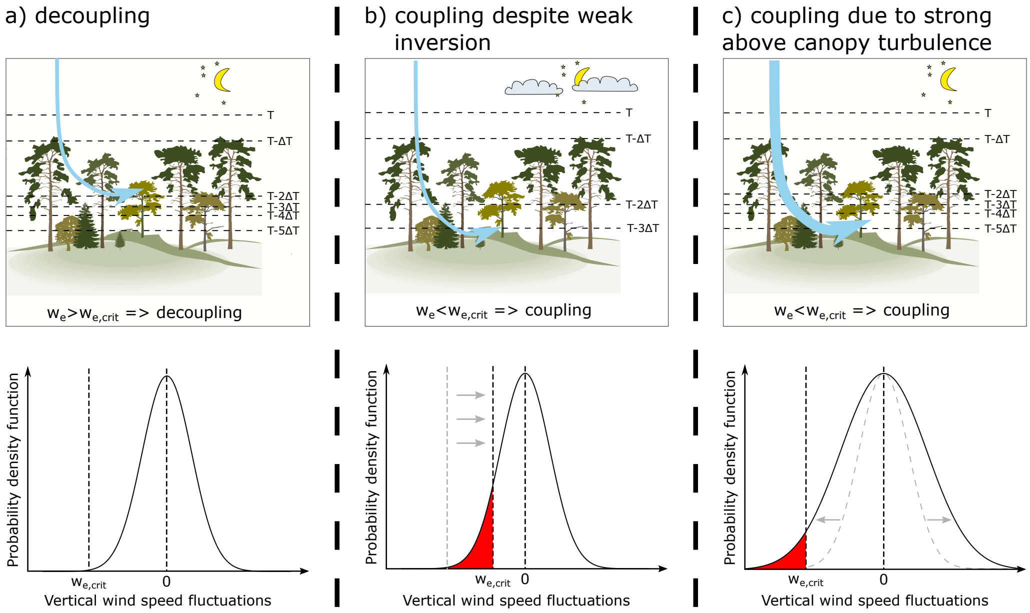

Figure 1. Schematic illustration of different decoupling situations. a): the above canopy flow is decoupled from the surface since the negative vertical wind speed fluctuations are not strong enough to counterbalance the movement hindering forces. b): coupling with the surface due to weaker stratification compared to a). c): coupling due to stronger turbulent fluctuations when compared to a). Bottom: fraction of w' data below $w_{e,crit}$ is shown with red.

to 0.15 for this study), LAI is leaf area index, U_h is horizontal wind speed at the canopy height (m s^{-1}), g is the acceleration due to gravity (m s^{-2}), $\hat{\theta}$ is the mean potential temperature below h and θ_e is the potential temperature of the downward moving air parcel. If the speed of the air parcel is equal to $w_{e,crit}$, then its kinetic energy is sufficient to counterbalance the work performed against the hindering forces. However, if it is less than this critical speed, then its downward movement stops before it reaches the ground and as a result interaction with the surface does not occur (see Fig. 1).

In order to couple above canopy flow with the forest floor a large enough fraction of negative vertical wind speed fluctuations (w') needs to be below $w_{e,crit}$. Considering Taylor's frozen turbulence hypothesis, this coincides with the definition that in coupled flow large enough cross-sectional area of the flow at height z needs to be governed by strong downward gusts which interact with the surface. Here we defined the flow to be coupled with the surface when more than 5% of the w' data were below $w_{e,crit}$, weakly coupled when between 1% and 5% of w' data were below $w_{e,crit}$ and decoupled when less than 1% were below $w_{e,crit}$. Future work is needed to validate the general applicability of these thresholds, yet their applicability at two contrasting sites are demonstrated below (see also Sect. 4.4). Assuming Gaussian distribution for w', these criteria can be described using the standard deviation of w:

$$
\begin{aligned}
&\Lambda \geq 0.61 \rightarrow \text{coupled} \\
0.43 \leq\ &\Lambda < 0.61 \rightarrow \text{weakly coupled} \\
&\Lambda < 0.43 \rightarrow \text{decoupled}
\end{aligned}
\qquad (2)
$$

where the decoupling metric Λ is defined as $\frac{\sigma_w}{|w_{e,crit}|}$. Therefore the flow can couple with the ground if σ_w increases (turbulent mixing increases), U_h or LAI decrease (canopy drag decreases) or $(\theta_e - \hat{\theta})$ or h decreases (influence of buoyancy and vertical distance decrease).

Atmospheric observations are typically made at some distance above the canopy during which the speed of downward propagating air parcel may be already slowed down due to stratification. The change in the speed of the air parcel when it traverses between heights z and h can be calculated as

$$w_e(z) = -\sqrt{w_e(h)^2 + 2g(z-h)\frac{\theta_e - \tilde{\theta}}{\tilde{\theta}}}, \tag{3}$$

where $w_e(z)$ and $w_e(h)$ are the air parcel speed at heights z and h and $\tilde{\theta}$ is the mean air potential temperature between z and h. Hence in order to evaluate the coupling of air at height z with the ground, Eq. 1 should be used to calculate $w_{e,crit}$ at the canopy height (h) and then use Eq. 3 to translate this value from h to z prior to comparing to σ_w values at height z.

In the case of neutral stratification, $w_{e,crit}$ reduces to

$$w_{e,crit} = -2\gamma\hat{c_d}\mathrm{LAI}U_h, \tag{4}$$

indicating that the limiting vertical wind speed needed to couple with the forest floor increases linearly with LAI and U_h. On the other hand, in the case of flat surfaces without emergent vegetation (i.e. LAI≈0), $w_{e,crit}$ reduces to

$$w_{e,crit} = -\sqrt{2gz\frac{\theta_e - \hat{\theta}}{\hat{\theta}}} = -\sqrt{2}zN, \tag{5}$$

where N is the Brunt-Väisälä frequency estimated using the bulk θ gradient ($N = \sqrt{\frac{g(\theta_e - \hat{\theta})}{\hat{\theta}z}}$). Using the definition for buoyancy length scale ($L_B = \frac{\sigma_w}{N}$) (Mahrt, 1979; Moum, 1996; Sorbjan, 2006; Sorbjan & Balsley, 2008; Mahrt et al., 2012) we can write

$$\Lambda = \frac{L_B}{\sqrt{2}z}. \tag{6}$$

Hence in the case of LAI≈0, the criterium for the flow to couple with the surface (Eq. 3) can be described with the ratio between L_B and height z.

3 Data and instrumentation

Measurements were collected at two contrasting locations: observations at Hyytiälä boreal pine forest (61.845°N, 24.289°E, 181 m a.s.l) and during "Fluxes over snow-covered surfaces II" (FLOSS-II) campaign above snow-covered rangeland (40.659°N, 106.324°W, 2477 m a.s.l). Hyytiälä is part of the ICOS measurement network (Franz et al., 2018) and has contributed to the global measurement network FLUXNET since the initiation of the site in 1996. The forest is governed by Scots pines (*Pinus Sylvestris*) with approximate tree height of 17 m. One-sided LAI of the forest is 4 m^2 m^{-2} and the canopy layer is between 10 and 17 m. Turbulence profiles within the forest have been studied in Launiainen et al. (2007). In this study observations made during summer 2019 (25 May to 30 Sep) were utilised. The measurement configuration consisted of vertical fiber-optic based distributed temperature sensing (DTS) observations (until 10 July), EC flux measurements (27 m height) (Rebmann et al., 2018) and temperature and CO_2 concentration profiles (Montagnani et al., 2018). For details, see Peltola, Lapo, Martinkauppi, et al. (2020), however there were four notable differences: 1) 10-min averaging period was used, 2) single-ended data (25 May to 3 June) were also included, 3) both directions in the double-ended configuration were utilised and 4) the DTS temperature observations were denoised using singular value decomposition prior to analysis (Epps & Krivitzky, 2019). Note that denoising has an effect only on Fig. 2 since otherwise mean profiles were used. When calculating $w_{e,crit}$, DTS measurements were utilised when available. All the data analyses were restricted to night time periods (global radiation < 5 W m^{-2}).

The observations made during the FLOSS-II measurement campaign (from Dec 2002 to end of March 2003) have been widely utilised in the analysis of vSBL (e.g. Mahrt & Vickers, 2005, 2006; Mahrt, 2010; Sun et al., 2020). A 30 m tall tower located in a flat terrain with grass and partial snow-coverage was instrumented with 3D sonic anemometers at seven levels and slow-response thermometers at eight levels. Quality-controlled and 5 min averaged data were retrieved from https://doi.org/10.5065/D6QC01XR (UCAR/NCAR - Earth Observing Laboratory, 2017).

4 Results and discussion

4.1 Examples of contrasting flow regimes

Figure 2 shows three 10-min examples of observations in the 125 m tall mast in Hyytiälä pine forest during contrasting flow regimes: a) quiescent flow which was decoupled from the ground, b) turbulent flox above canopy which was decoupled from the forest floor and c) strongly turbulent flow coupled to the ground. Coherent eddies consisting of sweep-ejection cycle (Thomas & Foken, 2007b; Finnigan et al., 2009) were observed in b) and c), but not in a). The downward moving sweep phases of the coherent motions can be identified as the warm tongues penetrating into the cold below-canopy air space, whereas the ejections bring relatively cold below-canopy air to upper levels above the forest canopy (due to downward directed heat flux). Note that the sweeping phases in b) did not reach the forest floor and as a result the flow was decoupled from the ground. This was identified also with the decoupling metric Λ (see subplot title).

CO_2 concentration profiles showed clear differences between the three examples, as a result from the different mixing regimes. The overall concentration difference between the highest (27 m) and lowest level (0.5 m) were 67, 26, 9 ppm, respectively. Note that in case b) this concentration difference resulted almost entirely from the CO_2 pooled below 8.8 m height, since the CO_2 above this height was effectively flushed out from the ecosystem by the coherent eddies.

4.2 Decoupling in relation to TKE production and transport

Above open terrain, Sun et al. (2012) argued that in stably stratified coupled flow regime turbulent kinetic energy (TKE) should be driven bulk shear (U/z) due to large eddies and shear production dominating the TKE budget. Hence, they analysed V_{TKE} ($V_{\mathrm{TKE}} = \sqrt{\mathrm{TKE}} = \sqrt{\sigma_u^2 + 0.5\sigma_v^2 + 0.5\sigma_w^2}$) dependency on U and found a threshold value for U above which V_{TKE} dependency on U was linear. Observations falling in this strong wind regime have been considered to relate to coupled flow regime (Mahrt et al., 2015; Acevedo et al., 2016; Sun et al., 2016; Lan et al., 2018; Freundorfer et al., 2019). Figures 3a and 3b show V_{TKE} dependency on U for two heights in FLOSS-II dataset, with data differentiated to separate flow regimes (based on Eq. 3) prior to analysis. In contrast to Sun et al. (2012), in the coupled regime no U threshold was observed and V_{TKE} followed the same linear dependence on U regardless of wind speed value. This suggests that in the stable coupled regime TKE was driven by bulk shear as proposed by Sun et al. (2012), however, this holds regardless of U not confirming the interpretation in Sun et al. (2012). Similar results were found for the forest site (Hyytiälä) using above-canopy U and V_{TKE} (not shown). Hence, we argue that flow decoupling cannot be judged based on U alone.

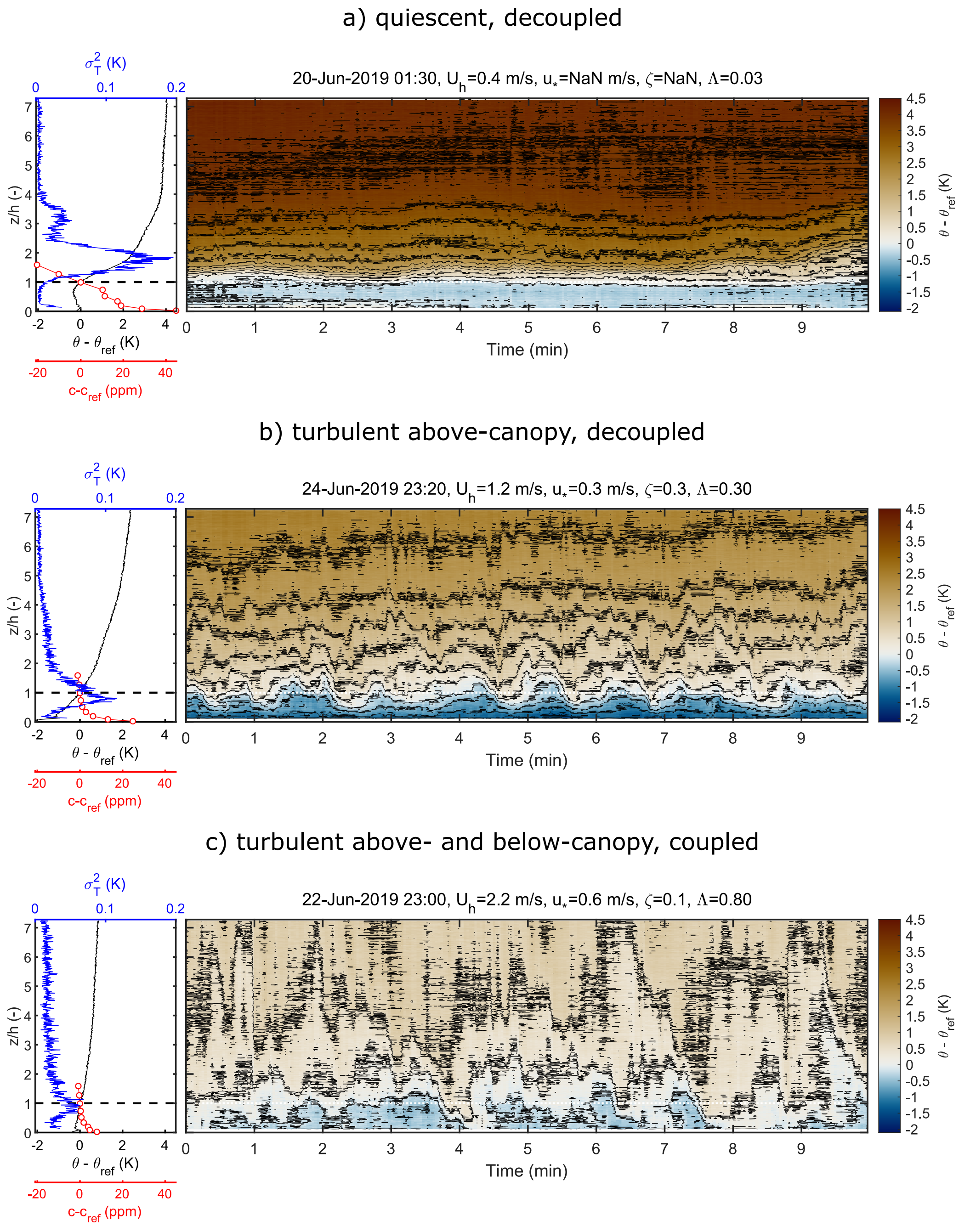

Figure 2. Right: examples of DTS-data during contrasting flow regimes (black lines=θ isolines). Left: corresponding temperature variance (blue), mean potential temperature (θ, black) and CO_2 concentration (c, red dots) profiles. c_{ref} and θ_{ref} equal mean c and θ values at canopy height (h).

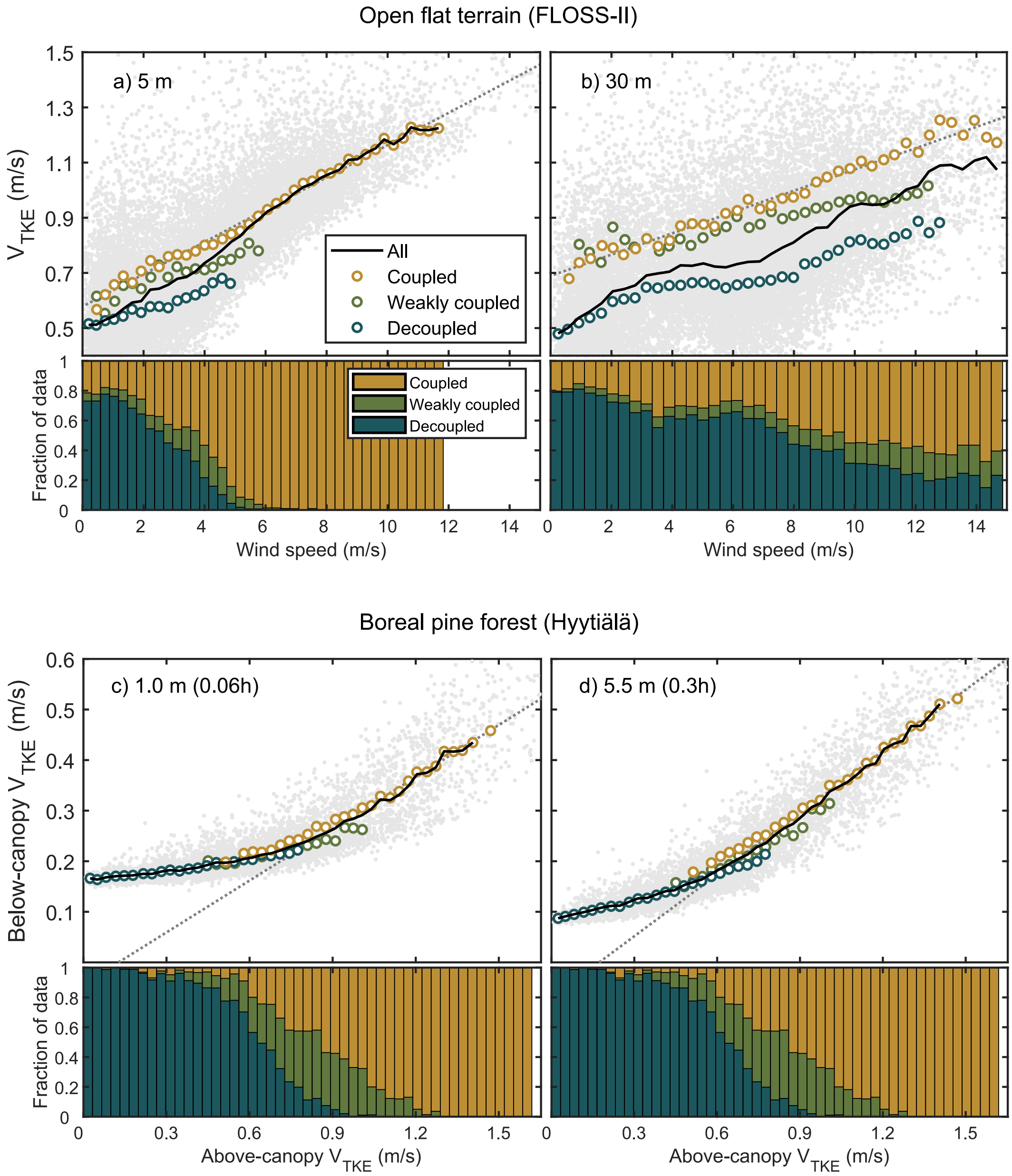

Figure 3. (a and b) V_{TKE} dependency on wind speed (U) at FLOSS-II following Sun et al. (2012). Additionally, data were divided into different coupling regimes (see Eq. 3) prior to analysis. Note that threshold wind speed (Sun et al., 2012) was not observed in the coupled regime. (c and d) comparison of above- and below-canopy V_{TKE} at Hyytiälä following Thomas et al. (2013). Gray dots = all the night-time data, circles and black lines = bin-averages for bins with more than 20 data points. Bottom: fraction of data in the three flow regimes (Eq. 3).

In prior studies cross-canopy coupling have been analysed by comparing concurrent measurements of σ_w below- and above-canopies (Thomas et al., 2013; Jocher et al., 2017, 2018; Freundorfer et al., 2019). Linear dependence between the two observations of σ_w were thought to signal coupling, since downward penetrating canopy-scale sweeps dominate the below-canopy TKE in coupled flow (Vickers & Thomas, 2013; Russell et al., 2017; Freundorfer et al., 2019). In accordance with these studies, the coupled flow regime was typically related to periods with high above-canopy V_{TKE} with a linear dependence between above- and below-canopy V_{TKE} (Figs. 3c and 3d). In contrast, low above-canopy V_{TKE} was related to decoupled regime. In this regime, below-canopy TKE was dominated by Kármán vortex streets created behind trees and hence independent of above-canopy TKE (Cava et al., 2008; Russell et al., 2017) since downward propagating sweeps did not reach the below-canopy air space (see also Fig. 2b). In our study, the wake-production generated a clear secondary peak in turbulence spectra (especially in 1 m height data) at the vortex shedding frequency based on constant Strouhal number, U and tree trunk diameter (not shown). At intermediate above-canopy V_{TKE} levels (0.5...0.8 m/s) the observations related to coupled flow regime departed from the linear dependence observed at higher V_{TKE} values. This might be due to importance of both, wake-production and sweeps, on below-canopy TKE at these above-canopy TKE levels and further analyses are warranted.

4.3 Turbulent fluxes in the coupled and decoupled layer

The sensible heat flux (H) profiles in the FLOSS-II dataset were analysed in the view of flow decoupling dependency on height (Eq. 6, Sect. 4.4.1). Nocturnal flux profiles were calculated so that each of the seven measurement heights was used as the highest observational level identified to be coupled with the surface (denoted with z_{co}). Hence observations below and above z_{co} correspond to coupled and decoupled layers, respectively. The fluxes were normalised with the H values at height z_{co} (H_{co}). Below z_{co} nearly constant H was observed, whereas above z_{co} the flux H decreased with height, since the flow above z_{co} was not connected to the surface (Fig. 4a). In the coupled air layer (i.e. below z_{co}), bin-averaged H was between $0.95H_{co}$ and $1.18H_{co}$ in agreement with the typical notion for constant-flux layer flows where the vertical turbulent fluxes are expected to vary by $\pm 10\%$. Note that discrepancies between flux footprints at different heights and biases stemming from instrument calibrations may have also influenced the observed H profiles.

CO_2 fluxes measured above the Hyytiälä forest during night depended on the degree of coupling (i.e. Λ) when $\Lambda < 0.61$, whereas in the coupled regime the fluxes were independent of Λ due to direct coupling of flux measurement height with the ground with turbulent mixing being no longer limiting. Figs. 4a and 4b shows physically the same phenomenon, but for different sites. Fluxes above z_{co} (Fig. 4a) and during periods with $\Lambda < 0.61$ (Fig. 4b) correspond to decoupled flow, whereas on the contrary above z_{co} and during periods with $\Lambda \geq 0.61$ correspond to coupled flow.

These results suggest that the method proposed in Sect. 2 can be used to estimate the depth of the layer that was coupled with the surface and hence e.g. to assess whether the observed turbulent fluxes related to the exchange of heat (FLOSS-II) or CO_2 (Hyytiälä) on the surface. Note that these results were obtained at two contrasting measurement sites without site-specific thresholds. This is due to using a ratio of variables related to kinetic energy (σ_w) and the energy required to couple with the ground ($w_{e,crit}$) in the analysis, instead of using σ_w (Acevedo et al., 2009; Thomas et al., 2013; Jocher et al., 2018) or related variables (u_*, U) (e.g. Gu et al., 2005; Sun et al., 2012) alone. Furthermore, this ratio does not depend on the source for the turbulent mixing in any way, it merely compares the existing kinetic energy to the energy needed to couple with the ground. Hence the decoupling metric should be applicable also in situations when the source does

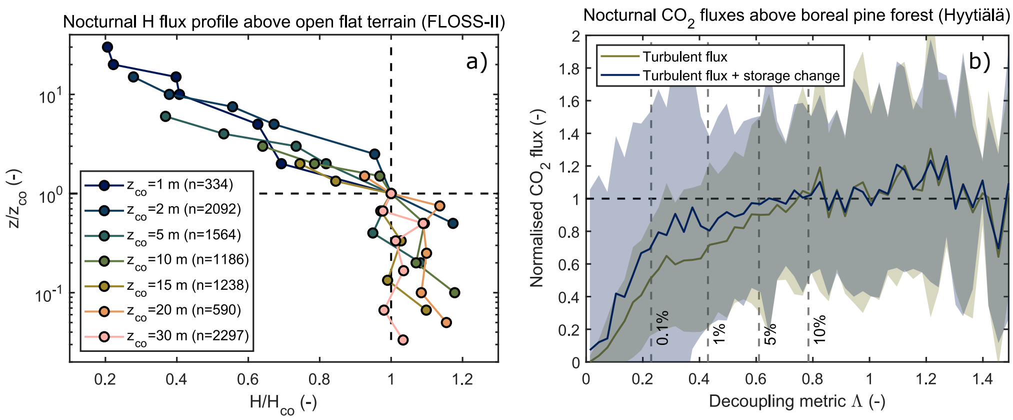

Figure 4. The physical interpretation of the coupling metric for (a) sensible heat profiles at FLOSS-II and (b) cross-canopy coupling of carbon dioxide at Hyytiälä. a): Normalised sensible heat flux (H) profiles (bin medians) observed at FLOSS-II. Profiles were calculated from periods when height z_{co} was coupled with the surface (cf. Eq. 3), but heights above z_{co} were not. Fluxes were normalised with H observed at z_{co} (H_{co}) b): Normalised nocturnal CO_2 fluxes measured at Hyytiälä plotted against Λ (lines=bin means, areas=$\pm\sigma$). Data were filtered based on stationarity criteria (Foken & Wichura, 1996). The storage change term (Finnigan, 2006) was also included. Fluxes were normalised with 2-week running means of nocturnal CO_2 fluxes during coupled regime. Vertical dashed lines = fraction of w' data below $w_{e,crit}$.

not conform with the traditional boundary layer (e.g. upside down boundary layer) (Mahrt et al., 2013; Mahrt, 2014).

4.4 Controls on flow decoupling

4.4.1 Flows above short vegetation

Above short vegetation (i.e. $LAI \approx 0$), $w_{e,crit}$ depends linearly on z and N (Eq. 5) and the definition for coupling (Eq. 3) can be written as

$$\sigma_w \geq 0.61\sqrt{2}zN. \tag{7}$$

Hence at a given value for N, the σ_w needed to couple the flow with the surface increases linearly with height. This is in line with prior experimental findings (Mahrt et al., 2013; Acevedo et al., 2016). The increase reflects the fact that the kinetic energy of downward moving air parcel needs to be higher when the height increases since there is a thicker air column below the air parcel within which the buoyancy force opposes its movement, i.e. the potential energy of the air parcel increases with height. In the FLOSS-II dataset rarely the upper level was identified to be coupled with the surface when the observation level below was not (less than 1% of observations). In general the lower levels were observed to be coupled with the surface more frequently than the upper levels, for instance 5 m height was coupled with the surface 64 % of time, whereas 20 m height only 39 % of time.

4.4.2 Flows above tall vegetation

In the case of neutral stratification below canopy height, using Eq. 4 the definition for coupling (Eq. 3) can be written as

$$I_w \geq 1.22\gamma\hat{c_d}\text{LAI}, \tag{8}$$

where I_w is the vertical turbulence intensity at the canopy height ($I_w = \frac{\sigma_w}{U_h}$). Note the similarity between the right hand side of Eq. 8 and the ratio between canopy height and coherent eddy penetration depth in Nepf et al. (2007), Cava et al. (2008) and Ghisalberti (2009) (i.e. $\propto \hat{c_d}$LAI) which describes whether the coherent canopy eddies are interacting with the surface or not. At the Hyytiälä site in near neutral conditions above the forest I_w was on average 0.26, whereas the limit for decoupling calculated using Eq. 8 was 0.20, indicating coupling at this site in near-neutral conditions. In contrast, Thomas et al. (2013) observed frequent decoupling above their dense forest (PAI=9.4 m^2m^{-2}) even during daytime despite similar I_w levels (0.25-0.30) and the decoupling could have been predicted with Eq. 8. It should be noted however that γ and $\hat{c_d}$ depend on canopy architecture (Amiro, 1990a, 1990b) and the influence of these parameters should be investigated. Clearly this method should be tested across range of sites with contrasting canopies, albeit similarities to the studies of Nepf et al. (2007), Cava et al. (2008) and Ghisalberti (2009) do suggest of a more general applicability.

5 Conclusions

Poor understanding of the very stable boundary layer is an obstacle for all scientific studies investigating surface-atmosphere interactions, in particular in the case of canopy flows. Here, we propose a novel simple first-principle based scheme to identify periods when the air flow is not in interaction with the underlying surface (i.e. it is decoupled). It was shown to correctly identify periods when the measured turbulent fluxes were not representative of the fluxes at the surface. The metric for flow decoupling based on this concept enabled analytical derivation of flow decoupling dependency on height, stratification and leaf area index. The approach is an improvement to the commonly used methods based on e.g. friction velocity filtering since 1) the proposed approach takes into account also changes in forces hindering the coupling (canopy drag, stable stratification) unlike traditional methods which utilise metrics for turbulent mixing or production alone and 2) it is based on first principles and not on ad-hoc empirical correlations. From a practical point-of-view, the approach requires only basic micrometeorological measurements (turbulence measurements at one height and temperature profile below it) in addition to knowledge of canopy density and hence should be applicable at most flux sites through the complete gradient from locations with short canopies to dense tall forests.

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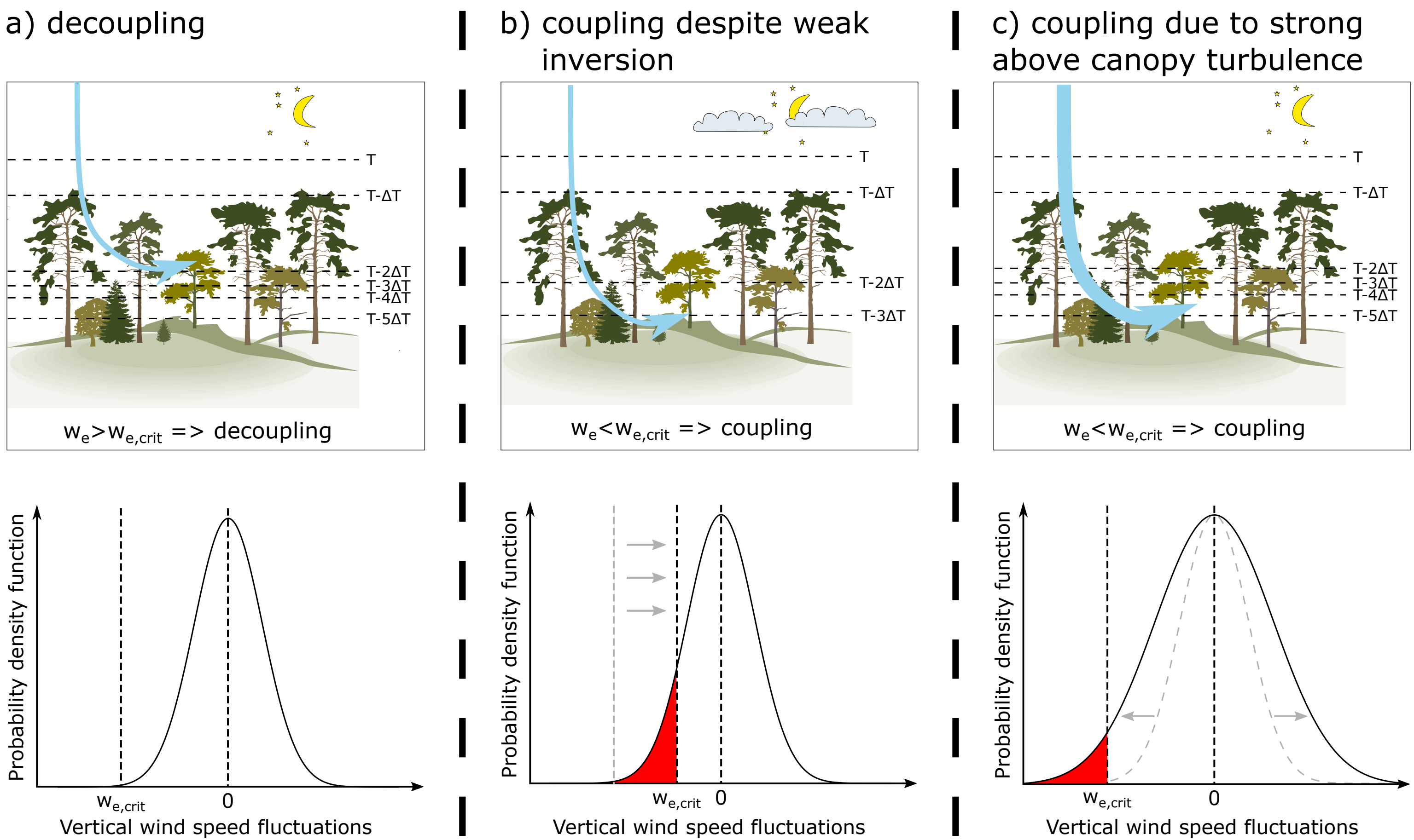
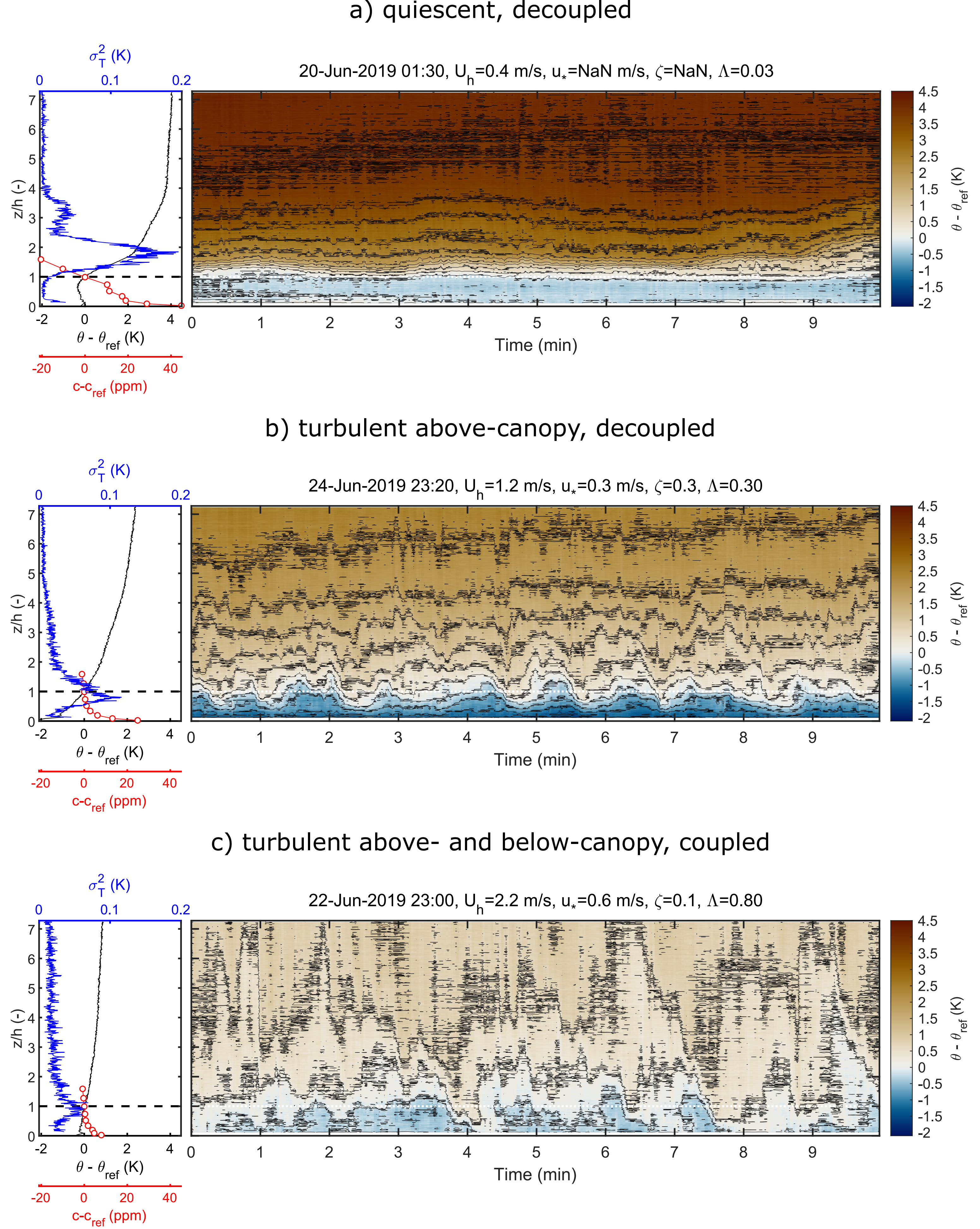
Figure 1.



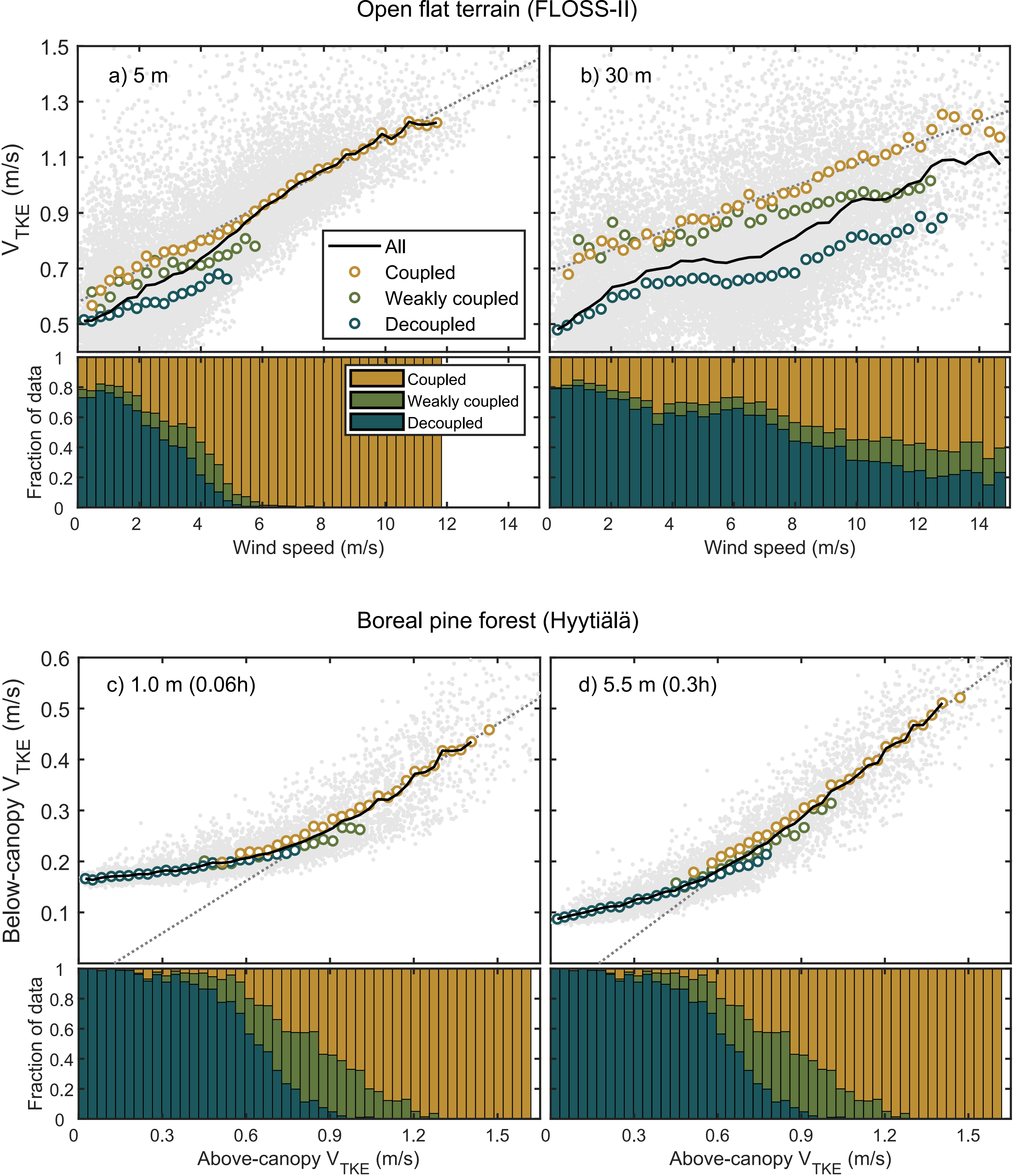
a) decoupling
b) coupling despite weak inversion
c) coupling due to strong above canopy turbulence
T
T-ΔT
T-2ΔT
T-3ΔT
T-4ΔT
T-5ΔT
$w_e > w_{e,crit}$ => decoupling
$w_e < w_{e,crit}$ => coupling
$w_e < w_{e,crit}$ => coupling
Probability density function
$w_{e,crit}$
0
Vertical wind speed fluctuations

Figure 2.



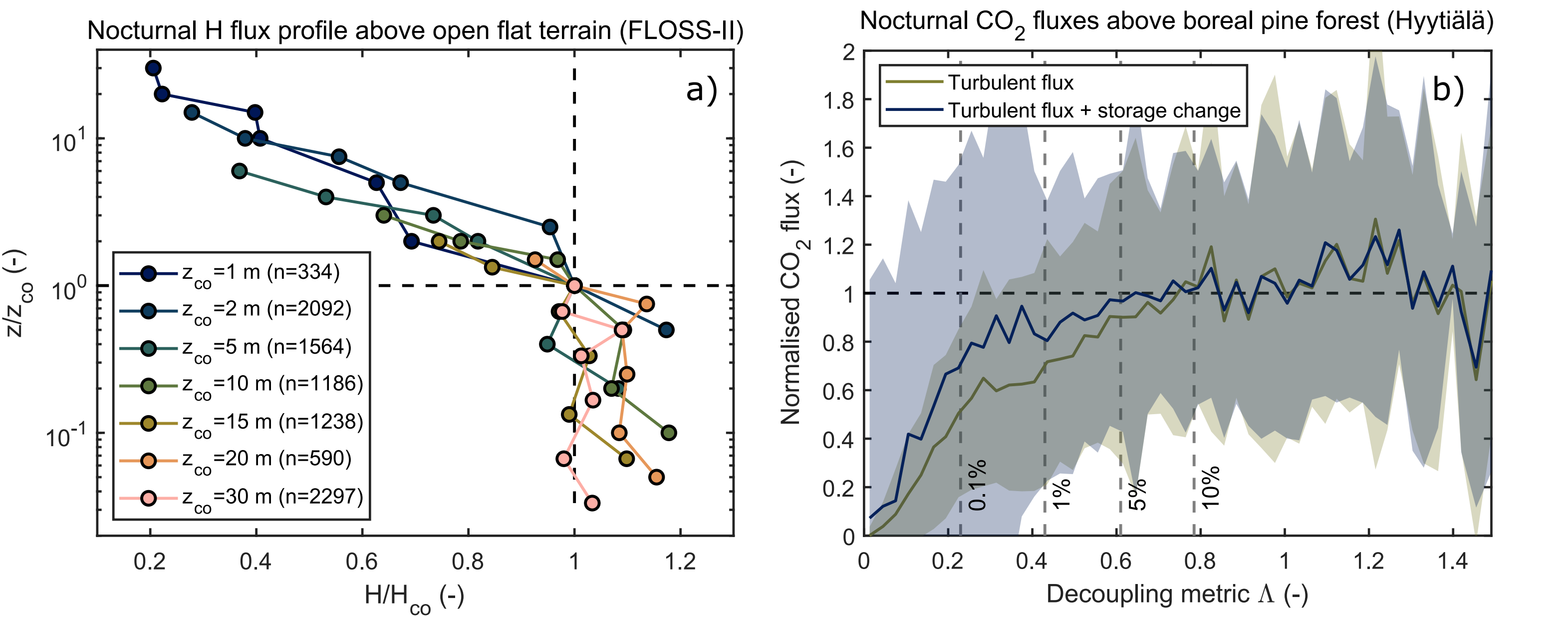
a) quiescent, decoupled
20-Jun-2019 01:30, U_h=0.4 m/s, u_*=NaN m/s, ζ=NaN, Λ=0.03
b) turbulent above-canopy, decoupled
24-Jun-2019 23:20, U_h=1.2 m/s, u_*=0.3 m/s, ζ=0.3, Λ=0.30
c) turbulent above- and below-canopy, coupled
22-Jun-2019 23:00, U_h=2.2 m/s, u_*=0.6 m/s, ζ=0.1, Λ=0.80
σ_T^2 (K)
z/h (-)
θ - θ_ref (K)
c-c_ref (ppm)
Time (min)

Figure 3.



Open flat terrain (FLOSS-II)
a) 5 m
b) 30 m
V_TKE (m/s)
All
Coupled
Weakly coupled
Decoupled
Fraction of data
Wind speed (m/s)
Boreal pine forest (Hyytiälä)
c) 1.0 m (0.06h)
d) 5.5 m (0.3h)
Below-canopy V_TKE (m/s)
Above-canopy V_TKE (m/s)

Figure 4.



Nocturnal H flux profile above open flat terrain (FLOSS-II)
a)
z/z_co (-)
H/H_co (-)
z_co=1 m (n=334)
z_co=2 m (n=2092)
z_co=5 m (n=1564)
z_co=10 m (n=1186)
z_co=15 m (n=1238)
z_co=20 m (n=590)
z_co=30 m (n=2297)
Nocturnal CO2 fluxes above boreal pine forest (Hyytiälä)
b)
Turbulent flux
Turbulent flux + storage change
Normalised CO2 flux (-)
Decoupling metric Λ (-)
0.1%
1%
5%
10%

Supporting Information for "A physics-based universal indicator for vertical decoupling and mixing across canopies architectures and dynamic stabilities"

O. Peltola[1], K. Lapo[2,3], C. K. Thomas[2,3]

[1]Climate Research Programme, Finnish Meteorological Institute, P.O. Box 503, 00101 Helsinki, Finland

[2]Micrometeorology Group, University of Bayreuth, Bayreuth, Germany

[3]Bayreuth Center for Ecology and Environmental Research, Bayceer, University of Bayreuth, Bayreuth, Germany

Contents of this file

Corresponding author: O. Peltola, Climate Research Programme, Finnish Meteorological Institute, P.O. Box 503, 00101 Helsinki, Finland. (olli.peltola@fmi.fi)

Text S1: Derivation of $w_{e,crit}$

Derivation of $w_{e,crit}$ relies on the assumption that in order for a downward moving air parcel to reach the ground its kinetic energy must match the work needed to counterbalance the forces hindering the downward movement. Under stable stratification downward movement is hindered by buoyancy force F_B:

$$F_B = g\frac{\rho - \rho_e}{\rho_e}, \tag{1}$$

where g is acceleration due to gravity (m s^{-2}), ρ_e is density of the downward moving air parcel (kg m^{-3}) and ρ is the air density of air surrounding the air parcel. Note that F_B is relative to unit mass and both ρ and F_B depend on height z. Also canopy drag hinders air movement through the canopy. The drag force (F_D) per unit mass can be approximated with (e.g. Poggi, Katul, & Albertson, 2004; Cescatti & Marcolla, 2004; Watanabe, 2004):

$$F_D = -c_d a U w_e, \tag{2}$$

where c_d is drag coefficient (unitless), a is leaf area density (m^2 m^{-3}), U is horizontal wind speed (m s^{-1}) and w_e is the speed of the air parcel (m s^{-1}). All these four variables vary with height z. The work (W) needed to offset these two forces can be calculated as line integral from height h to the surface ($z = 0$ m):

$$\begin{aligned} W &= -\int_h^0 (F_B + F_D)\, dz && (3) \\ &= -\int_h^0 \left(g\frac{(\rho - \rho_e)}{\rho_e} - c_d a U w_e \right) dz && (4) \\ &= gh\frac{\hat{\rho} - \rho_e}{\rho_e} + \int_h^0 c_d a U w_e dz && (5) \end{aligned}$$

where $\hat{\rho}$ is the average air density in the air column below h. Following prior studies (Inoue, 1963; Amiro, 1990; Poggi, Porporato, et al., 2004; Yi, 2008) U and w_e profiles below canopy height were parameterized as $U(z) = U(h)e^{\beta(z/h-1)}$ and $w_e(z) = w_e(h)e^{\alpha(z/h-1)}$. The coefficients α and β were obtained by fitting to observations ($\beta = 2.0, R^2 = 0.98$ and $\alpha = 1.5, R^2 = 0.96$). σ_w profiles measured at the same site in a prior study were used (Launiainen et al., 2007) for determining α. This approach assumes that σ_w below canopy is governed by downward penetrating sweeps. Now if we assume that c_d and a are constant with height ($\hat{c_d}$ and $\hat{a}$, respectively), after integration we find

$$W \approx gh\frac{\hat{\rho} - \rho_e}{\rho_e} + \hat{c_d}\hat{a}U(h)w_e(h)\frac{h}{\beta + \alpha}\left(e^{-\beta-\alpha} - 1\right), \tag{6}$$

which can be further reduced to

$$W = gh\frac{\hat{\rho} - \rho_e}{\rho_e} - \gamma\hat{c_d}\mathrm{LAI}U_h w_e(h), \tag{7}$$

where LAI is leaf area index (LAI $= h\hat{a}$), $U_h = U(h)$ and γ is a constant depending on the horizontal wind and downward penetrating air parcel speed profiles below h ($\gamma = \frac{1-e^{-\beta-\alpha}}{\beta+\alpha}$). Note that since $\alpha > 0$ and $\beta > 0$, therefore also $\gamma > 0$.

Now since kinetic energy of downward moving air parcel ($\frac{1}{2}w_e(h)^2$) must match the work, we can equate

$$\frac{1}{2}w_e(h)^2 = gh\frac{\hat{\rho} - \rho_e}{\rho_e} - \gamma\hat{c_d}\mathrm{LAI}U_h w_e(h), \tag{8}$$

which can be solved for $w_e(h)$ to get $w_{e,crit}$:

$$w_{e,crit} = -\gamma\hat{c_d}\mathrm{LAI}U_h - \sqrt{\gamma^2\hat{c_d}^2\mathrm{LAI}^2U_h^2 + 2gh\frac{\hat{\rho} - \rho_e}{\rho_e}}. \tag{9}$$

Here only the negative root was selected as physically meaningful. Assuming that air density changes only due to temperature and that the air parcel heats up adiabatically during its descent, then $w_{e,crit}$ can be written using potential temperature (θ)

$$w_{e,crit} = -\gamma\hat{c_d}\mathrm{LAI}U_h - \sqrt{\gamma^2\hat{c_d}^2\mathrm{LAI}^2U_h^2 + 2gh\frac{\theta_e - \hat{\theta}}{\hat{\theta}}}, \tag{10}$$

which equals Eq. (1) in the main text.

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