# 1 Reviewer 4 (R4):

I believe that the authors benefit from reviewers' comments and address them satisfactorily.
Definitely, inclusion of variables in Table 1, derivation of equations and more detailed

4 information about Omega Theory improved readability and are useful for readers of HESS. I

5 found some minor points to improve the readability of the paper.

6 Response: We thank R4 for the detailed reading and comments to improve the manuscript.

This is a humble recommendation to the authors, they may be benefit from Zuecco et al.
(2016) about the magnitude of hysteresis (L667) for their future study. I agree with the
authors about their comparison underlining area or size of the hysteresis (L494-5). However,
if they want to parameterize the magnitude of hysteresis, it may be good to check this
aforementioned study for their future study.

Response: We have included the reference of Zuecco et al. (2016) in the revisedmanuscript.

14 <u>Minor Comments:</u>

L99. Delete 's' after 'represent'. Insert comma after 'supply'....net radiation and soil moisture represent the supply, and ....

- 17 Response: 's' is deleted now.
- 18 L111. Insert comma after 'cover'.
- 19 Response: A comma is inserted.
- L154. Replace 'are' with 'is'. My understanding you are expressing about the investigation.
- 21 Response: 'Are' is replaced with 'is'.
- 22 L205. I recommend plural because of more than one retrieval. The retrievals...
- 23 Response: Done, as suggested.

L221. I recommend re-write the parenthesis. My suggestion:....(compared to the sensitivity of TR to soil moisture and  $\lambda E$ ).

- 26 Response: Response: Necessary changes are made.
- 27 L223. Insert a dash after 'water'. ...water-stress controls...
- 28 Response: a dash is inserted.
- L268. Ensure unity. Write numbers in words as you did in L393, L399 etc. '4' and '8' should
  be replaced by four and eight, respectively.
- 31 Response: Necessary changes are made.
- 32 L304. Delete 's'. Plural verb. ...consist of...
- 33 Response: 's' deleted.

- 34 L367. I recommend 'was' instead of 'is'. It depends on authors.
- 35 Response: 'is' replaced by 'was'.
- 36 L375. Word choice. Replace 'from' with 'during'. ...during 1995-2005....

37 Response: 'from' is replaced with 'during'.

L383. I recommend taking TR outside the radiation properties. Pay attention to L550, you use TR as a thermal variable. So, it will be good to take it out from the radiation variables. It depends on authors. My recommendation:

- 41 ...radiation (RN, shortwave and longwave), thermal (TR), meteorological...
- 42 Response: Necessary changes are made.
- 43 L401. I think 'over' should be replaced with 'for' as seen in L403.
- 44 Response: 'over' is replaced with 'for'.
- 45 L404. Delete 'that'. ... less than half of those....
- 46 Response: 'that' is deleted.
- 47 L424. Insert 'respectively' after 'gA-BM13' ......gA-BM13, respectively.
- 48 Response: 'respectively' is inserted in appropriate place.
- 49 L439. Insert 'respectively' after 'fluxes' ......fluxes, respectively (Fig.4).
- 50 Response: 'respectively' is inserted in appropriate place.
- 51 L468. ...relativeLY higher....
- 52 Response: Corrected now.
- L469. I may be confused here due to 'of'. My understanding which defines 'reasons', so isshould be are. Please check the sentence.
- 55 Response: The sentence is modified as...

'Interestingly, coupling was relatively higher in pasture during the dry seasons and thereasons are detailed in the following section and discussion.'

- 58 L555. I think 'is' should be 'are'. ... the accuracies....ARE limited....
- 59 Response: 'is' replaced with 'are'.
- 60 L588. 'results' should be 'result'. ....changes .... RESULT in....
- 61 Response: The sentence is corrected as,.....
- 62 'Here, fractional change in  $g_C$  results in an equivalent fractional change in  $\lambda E_T$ .'

- L622. I think 'is' should be 'are'. My understanding, 'which' refers to forests, gA of forests are
   higher than that of pastures.
- Response: Here 'is' seems to be appropriate. Here, 'which' refers to the coupling.
- L626. I recommend plural form of 'forest' due to using plural form of 'pasture'. ...forestS thanin the pastures.
- 68 Response: Necessary correction is made.
- 69 L679. Insert comma after 'model'.
- 70 Response: A comma is inserted.
- 71 L776. Replace 'is' with 'are'. .....differences.....ARE....
- 72 Response: 'is' replaced by 'are'.
- <sup>73</sup> L778. I recommend using (SW) instead of SW85. You used throughout the paper PM-SW.
- 74 Response: 'SW85' replace by 'SW'.
- 75 L791. Singular verb. ...tendS to....
- 76 Response: Necessary correction is made.
- 77 L1130. Table 2. Please insert degree sign (°) beneath latitude and longitude. Please define
- in the caption, (-) refers to (S) and (W) for latitude and longitude, respectively.
- 79 Response: Necessary correction is made.
- 80 S1P2L3. Introduced by Mallick et al (2015).
- 81 Response: Necessary correction is made.
- 82 S1P2L5. 'was' should be 'were'. ....effects.....WERE...
- 83 Response: 'was' replaced by 'were'.
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# 93 Canopy-scale biophysical controls of transpiration and 94 evaporation in the Amazon Basin

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## 126 Abstract:

Canopy and aerodynamic conductances ( $g_C$  and  $g_A$ ) are two of the key land surface 127 biophysical variables that control the land surface response of land surface schemes in 128 climate models. Their representation is crucial for predicting transpiration ( $\lambda E_T$ ) and 129 evaporation  $(\lambda E_E)$  flux components of the terrestrial latent heat flux  $(\lambda E)$ , which has 130 important implications for global climate change and water resource management. By 131 132 physical integration of radiometric surface temperature  $(T_R)$  into an integrated framework of the Penman-Monteith and Shuttleworth-Wallace model, we present a novel approach to 133 directly quantify the canopy-scale biophysical controls on  $\lambda E_T$  and  $\lambda E_E$  over multiple plant 134 functional types (PFTs) in the Amazon Basin. Combining data from six LBA (Large-scale 135 Biosphere-Atmosphere Experiment in Amazonia) eddy covariance tower sites and a  $T_{R}$ -136 driven physically-based modeling approach, we identified the canopy-scale feedback-137 response mechanism between  $g_C$ ,  $\lambda E_T$ , and atmospheric vapor pressure deficit ( $D_A$ ), without 138 using any leaf-scale empirical parameterizations for the modelling. The  $T_R$ -based model 139 shows minor biophysical control on  $\lambda E_T$  during the wet (rainy) seasons where  $\lambda E_T$  becomes 140 141 predominantly radiation driven and net radiation  $(R_N)$  determines 75% to 80% of the variances of  $\lambda E_T$ . However, biophysical control on  $\lambda E_T$  is dramatically increased during the 142 dry seasons, and particularly the 2005 drought year, explaining 50% to 65% of the variances 143 of  $\lambda E_T$  and indicates  $\lambda E_T$  to be substantially soil moisture driven during rainfall deficit phase. 144 Despite substantial differences in  $g_A$  between forests and pastures, very similar canopy-145 atmosphere 'coupling' was found in these two biomes due to soil moisture induced decrease 146 in  $g_C$  in the pasture. This revealed the pragmatic aspect of the  $T_R$ -driven model behavior 147 which exhibits a high sensitivity of  $g_C$  to per unit change in wetness as opposed to  $g_A$  that is 148 not sensitive to surface wetness variability. Our results reveal the occurrence of a significant 149 hysteresis effect between  $\lambda E_T$  and  $g_C$  during the dry season for the pasture sites, which is 150

151	attributed to relatively low soil water availability as compared to the rainforests, likely due to						
152	differences in rooting depth between the two systems. Evaporation was significantly						
153	influenced by $g_A$ for all the PFTs and across all wetness conditions. Our analytical framework						
154	accurately captures the responses of $g_C$ and $g_A$ to changes in radiation forcings, $D_A$ , and						
155	surface radiometric temperature, and thus appears to be promising for the improvement of						
156	existing land-surface-atmosphere exchange parameterisations across a range of spatial scales.						
157	Keywords: Canopy conductance, aerodynamic conductance, transpiration, evaporation,						
158	Penman-Monteith, Shuttleworth-Wallace, coupling, Amazon, LBA						
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## 173 **1 Introduction**

The Amazon rainforest is one of the world's most extensive natural ecosystems influencing 174 the Earth's water, energy, and carbon cycles (Malhi et al., 2012), and also a major source of 175 global terrestrial evapotranspiration (E) or latent heat flux ( $\lambda E$ ) (Costa et al., 2010; Harper et 176 al., 2014). An intensification of the Amazon hydrological cycle was observed in the past two 177 decades (Cox et al., 2000; Huntingford et al., 2008; Gloor et al., 2013). Recent Amazonian 178 droughts have gained particular attention due to the sensitivity of the tropical forest  $\lambda E$  to 179 climate change (Hilker et al., 2014). If persistent precipitation extremes become more 180 181 prevalent (Hilker et al., 2014); the Amazon rainforest may increasingly become a net source of carbon as a result of both the suppression of net biome exchange by drought and carbon 182 emissions from fires (Gatti et al., 2014). Changes in land cover due to conversion of tropical 183 forest to pastures significantly alters the energy partitioning of the region by decreasing  $\lambda E$ 184 and increasing sensible heat fluxes (*H*) over pasture sites (e.g. Priante-Filho et al., 2004). This 185 will ultimately lead to severe consequences for the water balance in the region, with 186 modifications to river discharge already observed in some parts of the Basin (Davidson et al., 187 2012). Evaluating the  $\lambda E$  response to changing climate and land use in the Amazon basin is 188 189 critical to understand the stability of the tropics within the Earth system (Lawrence and Vandecar, 2015). The control of  $\lambda E$  can be viewed as complex supply-demand interactions, 190 where net radiation and soil moisture represents the supply and the atmospheric vapor 191 pressure deficit represents the demand. This supply-demand interaction accelerates the 192 biophysical feedbacks in  $\lambda E$  and understanding these biophysical feedbacks is necessary to 193 assess the terrestrial biosphere response to water availability. Therefore, quantifying the 194 critical role of biophysical variables on  $\lambda E$  will add substantial insight to assessments of the 195 196 resilience of the Amazon basin under global change.

The aerodynamic and canopy conductances ( $g_A$  and  $g_C$ , hereafter) (unit m s<sup>-1</sup>) are the two 197 most important biophysical variables regulating the evaporation ( $\lambda E_E$ ) and transpiration ( $\lambda E_T$ ) 198 flux components of  $\lambda E$  (Monteith and Unsworth, 2008; Dolman et al., 2014; Raupach, 1995; 199 200 Colaizzi et al., 2012; Bonan et al., 2014). While  $g_A$  controls the bulk aerodynamic transfer of energy and water through the near-surface boundary layer,  $g_C$  represents the restriction on 201 water vapour flow through the aggregated conductance from stomata of the leaves, in case of 202 203 a vegetated land surface. In case of partial vegetation cover,  $g_C$  also includes soil surface conductance for evaporation. At small  $g_C/g_A$  ratio, the vapor pressure deficit close to the 204 205 canopy source/sink height  $(D_0)$  approximates the atmospheric vapor pressure deficit  $(D_A)$  due to aerodynamic mixing and/or low transpiration. These results in a strong canopy-atmosphere 206 coupling and such conditions are prevalent under soil moisture deficits. On the contrary, large 207  $g_C/g_A$  ratio influences the gradients of vapor pressure deficit just above the canopy, such that 208  $D_0$  tend towards zero and thus remains different from  $D_A$  (Jarvis and McNaughton, 1986). 209 This situation reflects a weak canopy-atmosphere coupling and such situations prevail under 210 211 predominantly wet conditions and/or poor aerodynamic mixing due to wetness induced low 212 aerodynamic roughness. The Penman-Monteith (PM) equation is a physically-based scheme 213 for quantifying such biophysical controls on canopy-scale  $\lambda E_E$  and  $\lambda E_T$  from terrestrial ecosystems, treating the vegetation canopy as a 'big-leaf' (Monteith, 1965; 1981). Despite its 214 development based on biophysical principles controlling water vapour exchange, quantifying 215 the  $g_A$  and  $g_C$  controls on  $\lambda E$  through the PM equation suffers from the continued 216 longstanding uncertainty over the aggregated stomatal and aerodynamic behaviour within the 217 soil-plant-atmosphere-continuum (Matheny et al., 2014; Prihodko et al., 2008). 218

One of the major sources of uncertainties in modeling  $g_A$  is associated with the empirical (and uncertain) parameterizations of near-surface boundary layer dynamics, which is invariably confounded by space-time variability in atmospheric stability (van der Tol et al., 2009;

Shuttleworth, 1989; Gibson et al., 2011). For example, Monin-Obukhov Similarity Theory 222 (MOST) used for  $g_A$  modeling appears to be only valid over uniform, extensive, and flat 223 surfaces (Monteith and Unsworth, 2008; van der Tol et al., 2009; Holwerda et al., 2012), and 224 its application to complex 'real' canopy systems is problematic due to chaotic interactions 225 between turbulence, canopy roughness and topography (Raupach and Finnigan, 1995; 226 Shuttleworth, 2007; Holwerda et al., 2012). Similarly,  $g_C$  varies in space and time due to 227 variations in plant species, photosynthetic capacity, soil moisture variability and 228 environmental drivers (Monteith and Unsworth, 2008; van der Tol et al., 2009). Despite the 229 230 existence of several semi-mechanistic and empirical parameterisations for  $g_C$  (e.g. Ball et al., 1987; Leuning, 1995; Tuzet et al., 2003; Medlyn et al., 2011), the adaptive tendencies of 231 plant canopies severely compromises the efficacy of such approaches (Matheny et al., 2014), 232 limiting their applicability over most landscapes. Thus, debate over the most appropriate 233 model of canopy conductance has endured for decades. 234

Previous studies in the Amazon Basin focused on developing an observational understanding 235 of the biogeochemical cycling of energy, water, carbon, trace gases, and aerosols in 236 Amazonia (Andreae et al., 2002; Malhi et al., 2002; da Rocha et al., 2009), model-based 237 understanding of surface ecophysiological behaviour and seasonality of  $\lambda E$  (Baker et al., 238 2013; Christoffersen et al., 2014), modelling the environmental controls on  $\lambda E$  (Hasler and 239 Avissar, 2007; Costa et al., 2010), understanding the seasonality of photosynthesis and of  $\lambda E$ 240 (da Rocha et al., 2004; Restrepo-Coupe et al., 2013), and the impact of land use on 241 hydrometeorology (Roy and Avissar, 2002; von Randow et al., 2012). However, the 242 243 combination of climatic and ecohydrological disturbances will significantly affect stomatal functioning, the partitioning of  $\lambda E_E - \lambda E_T$  and carbon-water-climate interactions of tropical 244 vegetation (Cox et al., 2000; Mercado et al., 2009). Hence, investigation of the effects of 245 drought and land cover changes on conductances,  $\lambda E_E$ , and  $\lambda E_T \frac{\text{are is}}{\text{are is}}$  topics requiring urgent 246

attention (Blyth et al., 2010) both because of the cursory way it is handled in current 247 generation of parametric models (Matheny et al., 2014) and because of the centrality of  $g_A$ 248 and  $g_C$  in controlling modelled flux behaviours (Villagarcía et al., 2010). The persistent risk 249 of deforestation is likely to alter the radiation interception, surface temperature, surface 250 moisture, associated meteorological conditions, and vegetation biophysical states of different 251 plant functional types (PFTs). Conversion from forest to pasture is expected to change the 252 253  $g_C/g_A$  ratio of these ecosystems and impact the evapotranspiration components. Besides inverting the PM equation using field measurements of  $\lambda E$ , till date either photosynthesis-254 dependent modeling or leaf-scale experiments were performed to directly quantify  $g_C$  (Ball et 255 al., 1987; Meinzer et al., 1993, 1997; Monteith, 1995; Jones, 1998; Motzer et al., 2005). 256 However, an analytical or physical retrieval for  $g_A$  and  $g_C$  is required not only to better 257 understand the role of the canopy in regulating evaporation and transpiration, but to enable a 258 capability to characterize the conductances using remote observations, across large spatial 259 domains where in-situ observations are not available. This paper aims to leverage this 260 emerging opportunity by exploring data from the Large-scale Biosphere-Atmosphere 261 262 Experiment in Amazonia (LBA) eddy covariance (EC) observations (e.g., de Gonçalves et al., 2013; Restrepo-Coupe et al., 2013) using a novel analytical modeling technique, the 263 Surface Temperature Initiated Closure (STIC) (STIC1.0 and STIC1.1) (Mallick et al., 2014, 264 2015) in order to quantify the biophysical control on  $\lambda E_E$  and  $\lambda E_T$  over several representative 265 PFTs of the Amazon Basin. 266

STIC provides a unique framework for simultaneously estimating  $g_A$  and  $g_C$ , surface energy balance fluxes,  $\lambda E_E$  and  $\lambda E_T$ . It is based on finding analytical solutions for  $g_A$  and  $g_C$  by physically integrating radiometric surface temperature ( $T_R$ ) information (along with radiative fluxes, meteorological variables) into the PM model (Mallick et al., 2014, 2015). The direct estimates of canopy-scale conductances and  $\lambda E$  obtained through STIC are independent of

any land surface parameterisation. This contrasts with the multi-layer canopy models that 272 explicitly parameterize the leaf-scale conductances and perform bottom-up scaling to derive 273 the canopy-scale conductances (Baldocchi et al., 2002; Drewry et al., 2010). A primary 274 advantage of the approach on which STIC is based is the ability to directly utilize remotely 275 sensed  $T_R$  to estimate E, thereby providing a capability to estimate E over large spatial scales 276 using a remotely sensed variable that is central to many ongoing and upcoming missions. 277 This study presents a detailed examination of the performance of STIC to better understand 278 land-atmosphere interactions in one of the most critical global ecosystems and addresses the 279 280 following science questions and objectives:

(1) How realistic are canopy-scale conductances when estimated analytically (or non-parametrically) without involving any empirical leaf-scale parameterization?

283 (2) What are the controls of canopy-scale  $g_A$  and  $g_C$  on evaporation and transpiration in the 284 Amazon basin, as evaluated using STIC?

(3) How do the STIC-based canopy-scale conductances compare with known environmentalconstraints?

287 (4) Is the biophysical response of  $g_C$  consistent with the leaf-scale theory (Jarvis and 288 McNaughton, 1986; McNaughton and Jarvis, 1991; Monteith, 1995)?

The following section describes a brief methodology to retrieve  $g_C$ ,  $g_A$ ,  $\lambda E_E$ , and  $\lambda E_T$ . The data sources used for the analysis are described after the methodology and will be followed by a comparison of the results with fluxes derived from EC measurements. A detailed discussion of the results and potential applicability of the method with implications for global change research are elaborated at the end. A list of symbols and variables used in the present study is given in Table 1.

## 295 **2 Methodology**

#### 296 **2.1 Theory**

The retrievals of  $g_A$ ,  $g_C$ , and  $\lambda E$  are based on finding a 'closure' of the PM equation (eqn. 1 297 below) using the STIC framework (Fig. A1 in Appendix) (Mallick et al., 2015). STIC is a 298 physically-based single-source surface energy balance scheme which includes internally 299 consistent estimation of  $g_A$  and  $g_C$  (Mallick et al., 2014, 2015). Originally designed for 300 301 application to thermal remote sensing data from Earth observation sensors, the STIC framework exploits observations of radiative  $(T_R)$ , and environmental variables including net 302 303 radiation  $(R_N)$ , ground heat flux (G), air temperature  $(T_A)$ , relative humidity  $(R_H)$  or vapor pressure  $(e_A)$  at a reference level above the surface. 304

The foundation of the development of STIC is based on the goal of finding an analytical 305 solution of the two unobserved 'state variables' ( $g_A$  and  $g_C$ ) in the PM equation while 306 307 exploiting the radiative ( $R_N$  and G), meteorological ( $T_A$ ,  $R_H$ ), and radiometric surface temperature  $(T_R)$  as external inputs. The fundamental assumption in STIC is the first order 308 dependence of  $g_A$  and  $g_C$  on the aerodynamic temperature ( $T_0$ ) and soil moisture (through  $T_R$ ). 309 This assumption allows a direct integration of  $T_R$  into the PM equation while simultaneously 310 constraining the conductances through  $T_R$ . Although the  $T_R$  signal is implicit in  $R_N$ , which 311 appears in the numerator of the PM equation (eqn. 1), it may be noted that  $R_N$  has a relatively 312 weak dependence on  $T_R$  (compared to <u>the sensitivity of</u>  $T_R$  sensitivities to <u>of</u> soil moisture and 313  $\lambda E$ ). Given  $T_R$  is the direct signature of the soil moisture availability, inclusion of  $T_R$  in the 314 315 PM equation also works to add water\_stress controls in  $g_C$ . Until now the explicit use of  $T_R$  in the PM model was hindered due to the unavailability of any direct method to integrate  $T_R$  into 316 317 this model, and, furthermore, due to the lack of physical models expressing biophysical states of vegetation as a function of  $T_R$ . Therefore, the majority of the PM-based  $\lambda E$  modeling 318 approaches strongly rely on surface reflectance and meteorology while exploiting the 319

- empirical leaf-scale parameterisations of the biophysical conductances (Prihodko et al., 2008;
  Bonan et al., 2014; Ershadi et al., 2015).
- 322 The PM equation is commonly expressed as,

$$\lambda E = \frac{s\phi + \rho c_P g_A D_A}{s + \gamma \left(1 + \frac{g_A}{g_C}\right)} \tag{1}$$

where  $\rho$  is the air density (kg m<sup>-3</sup>),  $c_P$  is the specific heat of air (J kg<sup>-1</sup> K<sup>-1</sup>),  $\gamma$  is the 323 psychrometric constant (hPa  $K^{-1}$ ), s is the slope of the saturation vapor pressure versus air 324 temperature (hPa  $K^{-1}$ ),  $D_A$  is the saturation deficit of the air (hPa) or vapor pressure deficit at 325 the reference level, and  $\phi$  is the net available energy (W m<sup>-2</sup>) (the difference between  $R_N$  and 326 G). The units of all the surface fluxes and conductances are in W m<sup>-2</sup> and m s<sup>-1</sup>, respectively. 327 For a dense canopy,  $g_C$  in the PM equation represents the canopy surface conductance. 328 Although it is not equal to the canopy stomatal conductance, it contains integrated 329 information of the stomata. For a heterogeneous landscape,  $g_C$  in the PM equation is an 330 aggregated surface conductance containing information on both canopy and soil. 331 Traditionally, the two unknown 'state variables' in eqn. (1) are  $g_A$  and  $g_C$ , and the STIC 332 methodology is based on formulating 'state equations' for these conductances that satisfy the 333 334 PM model (Mallick et al., 2014, 2015). The PM equation is 'closed' upon the availability of canopy-scale measurements of the two unobserved biophysical conductances, and if we 335 assume the empirical models of  $g_A$  and  $g_C$  to be reliable. However, neither  $g_A$  nor  $g_C$  can be 336 337 measured at the canopy-scale or at larger spatial scales. Furthermore, as shown by some recent studies (Matheny et al., 2014; van Dijk et al., 2015), a more appropriate  $g_A$  and  $g_C$ 338 model is currently not available. This implies that a true 'closure' of the PM equation is only 339 possible through an analytical estimation of the conductances. 340

#### 341 2.2 State equations

By integrating  $T_R$  with standard surface energy balance (SEB) theory and vegetation biophysical principles, STIC formulates multiple 'state equations' that eliminate the need for exogenous parametric submodels for  $g_A$  and  $g_C$ , associated aerodynamic variables, and landatmosphere coupling. The state equations of STIC are as follows and their detailed derivations are described Appendix (A1).

$$g_A = \frac{\phi}{\rho c_P \left[ (T_o - T_A) + \left(\frac{e_0 - e_A}{\gamma}\right) \right]}$$
(2)

$$g_{C} = g_{A} \frac{(e_{0} - e_{A})}{(e_{0}^{*} - e_{0})}$$
(3)

$$T_o = T_A + \left(\frac{e_0 - e_A}{\gamma}\right) \left(\frac{1 - \Lambda}{\Lambda}\right) \tag{4}$$

$$\Lambda = \frac{2\alpha s}{2s + 2\gamma + \gamma \frac{g_A}{g_C}(1+M)}$$
(5)

347 Here,  $T_0$  is the temperature (°C) at the source/sink height (or at the roughness length ( $z_0$ ) or in-canopy air stream),  $e_0$  is the atmospheric vapor pressure (hPa) at the source/sink height,  $e_0^*$ 348 is the saturation vapor pressure (hPa) at the source/sink height,  $\Lambda$  is the evaporative fraction 349 (the ratio of  $\lambda E$  and  $\phi$ ),  $\alpha$  is the Priestley-Taylor parameter (unitless) (Priestley and Taylor, 350 1972), and M is a unitless quantity which describes the relative wetness (or moisture 351 availability) of the surface. M controls the transition from potential to actual evaporation and 352 hence is critical for providing constraint against which the conductances can be estimated (M 353 estimation is explained in Appendix A2). Given values of  $R_N$ , G,  $T_A$ , and  $R_H$  or  $e_A$ , the four 354 state equations (eqn. 2 to 5) can be solved simultaneously to derive analytical solutions for 355 the four state variables. This also produces a 'closure' of the PM model, which is independent 356 of empirical parameterizations for both  $g_A$  and  $g_C$ . However, the analytical solution to the 357 above state equations have four accompanying unknowns; M (surface moisture availability), 358  $e_0$  (vapor pressure at the source/sink height),  $e_0^*$  (saturation vapor pressure at the source/sink 359

height), and Priestley-Taylor coefficient ( $\alpha$ ), and as a result there are 4-four equations with 8 360 eight unknowns. Consequently an iterative solution is needed to determine the four unknown 361 variables (as described in Appendix A2), which is a further modification of the STIC1.1 362 framework (Mallick et al., 2015). The present version of STIC is designated as STIC1.2 and 363 its uniqueness is the physical integration of  $T_R$  into a combined structure of the PM and 364 Shuttleworth-Wallace (SW, hereafter) (Shuttleworth and Wallace, 1985) model to estimate 365 the source/sink height vapor pressures (Appendix A2). In addition to physically integrating 366  $T_R$  observations into a combined PM-SW framework, STIC1.2 also establishes a feedback 367 loop describing the relationship between  $T_R$  and  $\lambda E$ , coupled with canopy-atmosphere 368 components relating  $\lambda E$  to  $T_0$  and  $e_0$ . For estimating M, the radiometric surface temperature 369  $(T_R)$  is extensively used in a physical retrieval framework, thus treating  $T_R$  as an external 370 input. In eqn. (5), the Priestley-Taylor coefficient ( $\alpha$ ) appeared due to the use of the 371 Advection-Aridity (AA) hypothesis (Brutsaert and Stricker, 1979) for deriving the state 372 373 equation of  $\Lambda$  (Supplement S1). However, instead of optimising  $\alpha$  as a 'fixed parameter', we 374 have developed a physical equation of  $\alpha$  (eqn. A15 in the Appendix A2) and numerically estimated  $\alpha$  as a 'variable'. The derivation of the equation for  $\alpha$  is described in Appendix A2. 375 The fundamental differences between STIC1.2 and earlier versions are described in Table 376 (A1). 377

In STIC1.2,  $T_0$  is a function of  $T_R$  and they are not assumed equal ( $T_0 \neq T_R$ ). The analytical expression of  $T_0$  is dependent on M and the estimation of M is based on  $T_R$ . To further elaborate this point on the inequality of  $T_0$  and  $T_R$ , we show an intercomparison of retrieved  $T_0$  versus  $T_R$  for forest and pasture (Fig. A2). This indicates the distinct difference of the retrieved  $T_0$  from  $T_R$  for the two different biomes.

#### **383 2.3 Partitioning** λ**E**

The terrestrial latent heat flux is an aggregate of both transpiration ( $\lambda E_T$ ) and evaporation 384  $(\lambda E_E)$  (sum of soil evaporation and interception evaporation from canopy). During rain events 385 the land surface becomes wet and  $\lambda E$  tends to approach the potential evaporation ( $\lambda E^*$ ), while 386 surface drying after rainfall causes  $\lambda E$  to approach the potential transpiration rate ( $\lambda E_T^*$ ) in the 387 presence of vegetation, or zero without any vegetation. Hence,  $\lambda E$  at any time is a mixture of 388 these two end member conditions depending on the degree of surface moisture availability or 389 wetness (M) (Bosveld and Bouten, 2003; Loescher et al., 2005). Considering the general case 390 of evaporation from an unsaturated surface at a rate less than the potential, M is the ratio of 391 the actual to the potential evaporation rate and is considered as an index of evaporation 392 efficiency during a given time interval (Boulet et al., 2015). Partitioning of  $\lambda E$  into  $\lambda E_E$  and 393  $\lambda E_T$  was performed according to Mallick et al. (2014) as follows: 394

395

$$\lambda E = \lambda E_F + \lambda E_T = M \lambda E^* + (1 - M) \lambda E_T^* \tag{6}$$

The estimates of  $\lambda E_E$  in the current method consists of aggregated contribution from both 396 'interception' and 'soil evaporation', and no further attempt is made to separate these two 397 components. In the Amazon forest, 'soil evaporation' has a negligible contribution while the 398 'interception evaporation' contributes substantially to the total evaporative fluxes, and, 399 therefore the partitioning of  $\lambda E$  into  $\lambda E_E$  and  $\lambda E_T$  is crucial. After estimating  $g_A$ ,  $\lambda E^*$  was 400 estimated according to the Penman equation and  $\lambda E_T$  was estimated as the residual in eqn. (6). 401 In this study, we use the term 'canopy conductance' instead of 'stomatal conductance' given 402 the term 'stomata' is applicable at the leaf-scale only. As stated earlier, for a heterogeneous 403 surface  $g_C$  should principally be a mixture of the canopy surface (integrated stomatal 404 information) and soil conductances. However, given the high vegetation density of the 405 Amazon Basin, the soil surface exposure is negligible, and, hence we assume  $g_C$  to be the 406

407 canopy-scale aggregate of the stomatal conductance. Similarly, different  $g_A$  exists for soil-408 canopy, sun-shade, and dry-wet conditions (Leuning, 1995); which is currently integrated 409 into a lumped  $g_A$  (given the big-leaf nature of STIC). From the big-leaf perspective, it is 410 generally assumed that the aerodynamic conductance of water vapor and heat are equal 411 (Raupach, 1998). However, for obtaining partitioned aerodynamic conductances, explicit 412 partitioning of  $\lambda E$  is needed, which is beyond the scope of the current manuscript.

# 413 2.4 Evaluating g<sub>A</sub> and g<sub>C</sub>

Due to the lack of direct canopy-scale  $g_A$  measurements, a rigorous evaluation of  $g_A$  cannot be performed. To evaluate the STIC retrievals of  $g_A$  ( $g_{A-STIC}$ ) we adopted three different methods: (a) By using the measured friction velocity ( $u^*$ ) and wind speed (u) at the EC towers and using the equation of Baldocchi and Ma (2013) ( $g_{A-BMI3}$ ) in which  $g_A$  was expressed as sum of turbulent conductance and canopy (quasi-laminar) boundary layer conductance as,

419 
$$g_{A-BM13} = [(u/u^{*2}) + (2/ku^{*2})(S_c/P_r)^{0.67}]^{-1}$$
(7)

420 where *k* is von Karman's constant, 0.4;  $S_c$  is the Schmidt Number;  $P_r$  is the Prandtl Number 421 and their ratio is generally considered to be unity. Here the conductances of momentum, 422 sensible and latent heat fluxes are assumed to be identical (Raupach, 1998).

423 (b) By inverting  $\lambda E$  observations for wet conditions hence assuming  $\lambda E \cong \lambda E^*$  and estimating 424  $g_A (g_{A-INV})$  as,

425 
$$g_{A-INV} = \gamma \lambda E / \rho c_P D_A \tag{8}$$

426 (c) By inverting the aerodynamic equation of *H* and estimating a hybrid  $g_A$  ( $g_{A-HYB}$ ) from 427 observed *H* and STIC  $T_0$  as ( $T_{0-STIC}$ ),

428 
$$g_{A-HYB} = H/\rho c_P (T_{0-STIC} - T_A)$$
(9)

Like  $g_{A-STIC}$ , direct verification of STIC  $g_C (g_{C-STIC})$  could not be performed as canopy-scale  $g_C$  observations are not possible with current measurement techniques. Although leaf-scale  $g_C$ measurements are relatively straightforward, these values are not comparable to values retrieved at the canopy-scale. However, assuming  $u^*$ -based  $g_A$  as baseline aerodynamic conductance, we have estimated canopy-scale  $g_C$  by inverting the PM equation ( $g_{C-INV}$ ) (Monteith, 1995) to evaluate  $g_{C-STIC}$  by exploiting  $g_{A-BM13}$  in conjunction with the available  $\phi$ ,  $\lambda E$ ,  $T_A$ , and  $D_A$  measurements from the EC towers.

## 436 **2.5 Decoupling coefficient and biophysical controls**

437 The decoupling coefficient or 'Omega' ( $\Omega$ ) is a dimensionless coefficient ranging from 0.0 to 438 1.0 (Jarvis and McNaughton, 1986) and considered as an index of the degree of stomatal 439 control on transpiration relative to the environment. The equation of  $\Omega$  is as follows:

$$\Omega = \frac{\frac{s}{\gamma} + 1}{\frac{s}{\gamma} + 1 + \frac{g_A}{g_C}}$$
(10)

## 440 Introducing $\Omega$ in the Penman-Monteith (PM) equation for $\lambda E$ results in:

$$\lambda E = \Omega \lambda E_{eq} + (1 - \Omega) \lambda E_{imp} \tag{11}$$

$$\lambda E_{eq} = \frac{s\phi}{s+\gamma} \tag{12}$$

$$\lambda E_{imp} = \frac{\rho c_P}{\gamma} g_C D_A \tag{13}$$

441 Where,  $\lambda E_{eq}$  is the equilibrium latent heat flux, which depends only on  $\phi$  and would be 442 obtained over an extensive surface of uniform moisture availability (Jarvis and McNaughton, 443 1986; Kumagai et al., 2004).  $\lambda E_{imp}$  is the imposed latent heat flux, which is 'imposed' by the 444 atmosphere on the vegetation surface through the effects of vapor pressure deficit (triggered 445 under limited soil moisture availability) and  $\lambda E$  becomes proportional to  $g_C$ . When the  $g_C/g_A$  ratio is very small (i.e., water stress conditions), stomata principally control the water loss and a change in  $g_C$  will result in a nearly proportional change in transpiration. Such conditions trigger strong biophysical control on transpiration. In this case the  $\Omega$  value approaches zero and vegetation is believed to be fully coupled to the atmosphere. In contrast, for a high  $g_C/g_A$  ratio (i.e., high water availability), changes in  $g_C$  will have little effects on the transpiration rate, and transpiration is predominantly controlled by  $\phi$ . In this case the  $\Omega$  value approaches unity, and vegetation is considered to be poorly coupled to the atmosphere.

Given both  $g_A$  and  $g_C$  are the independent estimates in STIC1.2, the concept of  $\Omega$  was used to 453 understand the degree of biophysical control on  $\lambda E_T$ , which indicates the extent to which the 454 transpiration fluxes are approaching the equilibrium limit. However, the biophysical 455 characterisation of  $\lambda E_T$  and  $\lambda E_E$  through STIC1.2 significantly differs from previous 456 approaches (Ma et al., 2015; Chen et al., 2011; Kumagai et al., 2004), and the fundamental 457 differences are centered on the specifications of  $g_A$  and  $g_C$  (as described in Table A2). While 458 the estimation of  $g_A$  in previous approaches <u>is-was</u> based on u and  $u^*$ , the estimation of  $g_C$  was 459 based on inversion of observed  $\lambda E$  based on the PM equation (e.g. Stella et al., 2013). 460 However, none of these approaches allow independent quantification of biophysical controls 461 of  $\lambda E$  as  $g_C$  is constrained by  $\lambda E$  itself. 462

## 463 **3 Datasets**

#### 464 **3.1 Eddy covariance and meteorological quantities**

We used the LBA (Large-Scale Biosphere-Atmosphere Experiment in Amazonia) data for quantifying the biophysical controls on the evaporative flux components. LBA was an international research initiative conducted <u>from-during</u> 1995-2005 to study how Amazonia functions as a regional entity within the larger Earth system, and how changes in land use and 469 climate will affect the hydrological and biogeochemical functioning of the Amazon470 ecosystem (Andreae et al., 2002).

A network of eddy covariance (EC) towers was operational during the LBA experiment, such 471 that data from nine EC towers were obtained from the ORNL Distributed Archive Active 472 473 Centre (ftp://daac.ornl.gov/data/lba/carbon\_dynamics/CD32\_Brazil\_Flux\_Network/). These 474 are the quality controlled and harmonized surface flux and meteorological data from the 475 Brazilian Amazon flux network. Time series of surface fluxes ( $\lambda E$ , H, G), radiation ( $T_{R_{T}}R_{N}$ , shortwave and longwave), thermal  $(T_R)$ , meteorological quantities  $(T_A, R_H, wind speed)$  as 476 477 well as soil moisture and rainfall were available from six (out of nine) EC towers. Three of the EC towers had numerous missing data and were not included in the analysis. The surface 478 energy balance was closed by applying the Bowen ratio (Bowen, 1926) closure as described 479 480 in Chavez et al. (2005) and later adopted by Anderson et al. (2007) and Mallick et al. (2015). In the absence of G measurements,  $\phi$  was assumed to be equal to the sum of  $\lambda E$  and H with 481 the assumption that a dense vegetation canopy restricts the energy incident on the soil 482 483 surface, thereby allowing us to assume negligible ground heat flux. For the present analysis, 484 data from six selected EC towers (Table 2) represent two different biomes (forest and pasture) covering four different PFTs, namely, tropical rainforest (TRF), tropical moist forest 485 (TMF), tropical dry forest (TDF), and pasture (PAS), respectively. A general description of 486 the datasets can be found in Saleska et al. (2013). For all sites, monthly averages of the 487 diurnal cycle (hourly time resolution) were chosen for the present analysis. 488

# 489 **4 Results**

# 490 **4.1 Evaluating** $g_A$ , $g_C$ , and surface energy balance fluxes

491 Examples of monthly averages of the diurnal cycles of the four different  $g_A$  estimates and 492 their corresponding  $g_C$  estimates over two different PFTs (K34 for forest and FNS for 493 pasture) reveal that  $g_{A-STIC}$  and  $g_{C-STIC}$  tend to be generally higher over-for the forest than their 494 counterparts, varying from 0 to 0.06 m s<sup>-1</sup> and 0 to 0.04 m s<sup>-1</sup> respectively (Fig. 1a and 1b). 495 The magnitude of  $g_{A-STTC}$  varied between 0 to 0.025 m s<sup>-1</sup> for the pasture (Fig. 1a), while  $g_C$ . 496  $|_{STTC}$  values were less than half that of those estimated over the forest (0 – 0.01 m s<sup>-1</sup>) (Fig. 1b). 497 The conductances showed a marked diurnal variation expressing their overall dependence on 498 net radiation, vapor pressure deficit, and surface temperature. Despite the absolute differences 499 between the conductances from the different retrieval methods, their diurnal patterns were 500 comparable.

501 The canopy-scale evaluation of  $g_{A-STIC}$  is illustrated in Fig. 2a (and Table 3) combining data from the four PFTs. Estimated values range between zero and 0.1 m s<sup>-1</sup> and show modest 502 correlation ( $\mathbb{R}^2 = 0.44$ ) ( $\mathbb{R}^2$  range between 0.22 [±0.18] to 0.55 [±0.12]) between  $g_{A-BM13}$  and 503  $g_{A-STIC}$  with regression parameters ranging between 0.81 (±0.023) and 1.07 (±0.047) for the 504 slope and 0.0019 ( $\pm 0.0006$ ) to 0.0006 ( $\pm 0.0006$ ) m s<sup>-1</sup> for the offset (Table 3). The root mean 505 squared deviation (RMSD) varied between 0.007 (TDF) and 0.013 m s<sup>-1</sup> (TRF). Statistical 506 comparisons between  $g_{A-STIC}$  and  $g_{A-HYB}$  revealed relatively low RMSD and high correlation 507 between them (RMSD = 0.007 m s<sup>-1</sup> and  $R^2 = 0.77$ ) as compared to the error statistics 508 between  $g_{A-STIC}$  and  $g_{A-INV}$  (RMSD = 0.011 m s<sup>-1</sup> and R<sup>2</sup> = 0.50) (Fig. 2b, 2c). The residuals 509 between  $g_{A-STIC}$  and  $g_{A-BM13}$  are plotted as a function of u and  $u^*$  in Fig. (2d) with the aim to 510 ascertain whether significant biases are introduced by ignoring wind and shear information 511 within STIC1.2. As illustrated in Fig. 2d, there appears to be a weak systematic relationship 512 between the residual  $g_A$  difference with either  $u^*$  or u (r = -0.26 and -0.17). However, a 513 considerable relationship was found between wind and shear driven  $g_A$  (i.e.,  $g_{A-BM13}$ ) versus  $\phi$ , 514  $T_R$  and  $D_A$  (r = 0.83, 0.48, and 0.42) (Fig. 2e and 2f), which indicates that these three energy 515 and water constraints can explain 69%, 23%, and 17% variance of  $g_{A-BM13}$ , respectively. 516

517 Canopy-scale evaluation of hourly  $g_C$  is presented in Fig. 3a (and Table 3) combining data 518 from the four PFTs. Estimated values range between zero and 0.06 m s<sup>-1</sup> for  $g_{C-STIC}$  and show

reasonable correlation ( $R^2 = 0.39$ ) ( $R^2$  range between 0.14 [±0.04] to 0.58 [±0.12]) between 519  $g_{C-STIC}$  and  $g_{C-INV}$  with regression parameters ranging between 0.30 (±0.022) and 0.85 520  $(\pm 0.025)$  for the slope and 0.0024  $(\pm 0.0003)$  to 0.0097  $(\pm 0.0007)$  m s<sup>-1</sup> for the offset (Table 521 3). The RMSD varied between 0.007 (PAS) and 0.012 m s<sup>-1</sup> (TRF and TDF). Given  $g_A$ 522 significantly controls  $g_C$ , we also examined whether biases in  $g_C$  are introduced by ignoring 523 wind and shear information within STIC. The scatterplots between residual  $g_C$  difference ( $g_C$ ) 524  $STIC - g_{C-INV}$  versus both u and  $u^*$  (Fig. 3b) showed  $g_C$  residuals to be evenly distributed 525 across the entire range of u and  $u^*$  and no systematic pattern was evident. 526

The reliability of STIC1.2-based  $g_A$  and  $g_C$  retrievals was further verified by evaluating  $\lambda E$ 527 and H estimates (Fig. 4). Both the predicted  $\lambda E$  and H are generally in good agreement with 528 the observations, with substantial correlation (r) ( $\mathbb{R}^2$  from 0.61 to 0.94), reasonable RMSD of 529 33 and 37 W m<sup>-2</sup>, and mean absolute percent deviation (MAPD) of 14% and 32% between 530 the observed and STIC fluxes (Fig. 4), respectively. Regression parameters varied between 531 0.96 (±0.008) to 1.14 (±0.010) for the slope and -16 (±2) to -2 (±2) W m<sup>-2</sup> for the offset for 532  $\lambda E$  (Table 4), whereas for H, these were 0.60 (±0.025) to 0.89 (±0.035) for the slope and 9 533 (±1) to 29 (±2) W m<sup>-2</sup> for the offset (Table 3), respectively. The RMSD in  $\lambda E$  varied from 20 534 to 31 W m<sup>-2</sup> and 23 to 34 W m<sup>-2</sup> for H (Table 3). 535

The evaluation of the conductances and surface energy fluxes indicates some efficacy for the
STIC derived fluxes and conductance estimates which represent a weighted average of these
variables over the source area around EC tower.

#### 539 4.2 Canopy coupling, transpiration and evaporation

From Fig. 5a an overall weak to moderate relationship (r = -0.31 to -0.42) is apparent between the coupling (i.e.,  $1-\Omega$ ) and  $\lambda E_T$ , where  $\lambda E_T$  is negatively related to the coupling for all the PFTs, thus indicating the influence of weak to moderate biophysical controls on  $\lambda E_T$ 

throughout the year in addition to radiative controls. The biophysical control was 543 substantially enhanced in TRF (r increased from -0.36 to -0.53 and -0.60) (47 to 67% 544 increase) and TMF (r increased from -0.31 to -0.53 and -0.58) (70 to 85% increase) during 545 the dry seasons (July-September) (Fig. 5a). A profound increase of biophysical control on 546  $\lambda E_T$  during the dry season was also found in TDF (52% increase) and PAS (37% increase) 547 (Fig. 5a). The negative relationship (r = -0.29 to -0.45) between (1- $\Omega$ ) and  $\lambda E_E$  (Fig. 5b) in all 548 four PFTs indicated the role of aerodynamic control on  $\lambda E_E$ . The aerodynamic control was 549 550 also enhanced during the dry seasons as shown by the increased negative correlation (r = -0.50 to -0.69) (Fig. 5b) between  $(1-\Omega)$  and  $\lambda E_E$ . 551

Illustrative examples of the diurnal variations of  $\lambda E_E$ ,  $\lambda E_T$ , and  $\Omega$  for two different PFTs with 552 different annual rainfall (2329 mm in rainforest, K34 and 1597 mm in pasture, FNS) for three 553 consecutive days during both dry and wet seasons are shown in Fig. 5c to 5f. This shows 554 morning rise of  $\Omega$  and a near-constant afternoon  $\Omega$  in the wet season (Fig. 5c and 5d), thus 555 indicating no biophysical controls on  $\lambda E_E$  and  $\lambda E_T$  during this season. On the contrary, during 556 the dry season, the morning rise in  $\Omega$  is followed by a decrease during noontime (15% to 25%) 557 558 increase in coupling in forest and pasture) (Fig. 5e and 5f) due to dominant biophysical control, which is further accompanied by a transient increase from mid-afternoon till late 559 afternoon and steadily declined thereafter. Interestingly, coupling was relatively higher in 560 pasture during the dry seasons the and the reasons of which is are detailed in the following 561 section and discussion. 562

## 563 **4.3** $g_c$ and $g_A$ versus transpiration and evaporation

Scatter plots between  $\lambda E_T$  and  $\lambda E_E$  versus  $g_C$  and  $g_A$  showed a triangular pattern which became wider with increasing conductances (Fig. 6). To explain this behaviour of  $\lambda E_T$  versus  $g_C$  and  $g_A$ , we further examined the entire mechanism of conductance- $\lambda E_T$  interactions

through two dimensional scatters between  $\lambda E_T$  and conductances for two consecutive diurnal 567 cycles during wet and dry seasons over rainforest and pasture sites with different annual 568 569 rainfall (e.g., K34 as wet and FNS as dry site, annual rainfall 2329 mm and 1597 mm) (Fig. 7). Our results confirm the occurrence of diurnal hysteresis between  $g_C$ - $g_A$  and  $\lambda E_T$  and 570 explain the reason for the shape of the curves obtained in Fig. 6. During the wet season, a 571 distinct environmental control is detectable on  $g_C$  and  $\lambda E_T$  in the morning hours (Fig. 7a and 572 7b) in both PFTs where  $g_C$  and  $\lambda E_T$  increased as a result of increasing  $R_N$ ,  $T_R$ , and  $D_A$ . From 573 the late morning to afternoon, a near-constant (forest) or negligible increase (pasture) of  $\lambda E_T$ 574 is observed despite substantial reduction of both  $g_C$  and  $g_A$  (25 to 50% decrease), after which 575  $\lambda E_T$  starts decreasing. This behaviour of  $\lambda E_T$  was triggered due to the concurrent changes in 576  $R_N$  (15 to 50% change),  $D_A$  (20 to 60% change) and surface temperature ( $T_R$ ) (5% to 14% 577 change), which indicates the absence of any dominant biophysical regulation on  $\lambda E_T$  during 578 the wet season (Fig. 7a and 7b). On the contrary in the dry season, although the morning rise 579 in  $\lambda E_T$  is steadily controlled by the integrated influence of environmental variables, but a 580 modest to strong biophysical control is found for both PFTs during the afternoon where  $\lambda E_T$ 581 582 substantially decreased with decreasing conductances (Fig. 7c and 7d). This decrease in  $\lambda E_T$ is mainly caused by the reduction in  $g_C$  as a result of increasing  $D_A$  and  $T_R$  (as seen later in 583 Fig. 8a and 8c). In the dry season, the area under the hysteretic relationship between  $\lambda E_T$ ,  $g_C$ 584 and environmental variables was substantially wider in pasture (Fig. 7d) than for the 585 rainforest (Fig. 7c), which is attributed to greater hysteresis area between  $R_N$  and  $D_A$  in 586 pasture as a result of reduced water supply. The stronger hysteresis effects in pasture during 587 the dry season (Fig. 7d) ultimately led to the stronger relationship between coupling and  $\lambda E_T$ 588 (as seen in Fig. 5a). 589

#### 590 **4.4 Factors affecting variability of** $g_c$ and $g_A$

The sensitivity of stomatal conductance to vapor pressure deficit is a key governing factor of 591 transpiration (Ocheltree et al., 2014; Monteith, 1995). We examined if the feedback or feed-592 forward response hypothesis (Monteith, 1995; Farquhar, 1987) between  $g_C$ ,  $D_A$ , and  $\lambda E_T$  is 593 reflected in our canopy-scale  $g_C$  retrievals. Combining data of all PFTs, we found an 594 exponential decline of  $g_C$  in response to increasing  $D_A$  regardless of the variations of net 595 radiation (Fig. 8a). High  $g_C$  is consistent with high humidity and low evaporative demand. 596 Five negatively logarithmic scatters fit the data with r values of 0.38 ( $0 < R_N < 150$  W m<sup>-2</sup>), 597 0.63 (150<  $R_N$  <300 W m<sup>-2</sup>), 0.73 (300<  $R_N$  <450 W m<sup>-2</sup>), 0.78 (450<  $R_N$  <600 W m<sup>-2</sup>), and 598 0.87 ( $R_N > 600 \text{ W m}^{-2}$ ). The sensitivity of  $g_C$  to  $D_A$  was at the maximum in the high  $R_N$  range 599 beyond 600 W m<sup>-2</sup> and the sensitivity progressively declined with declining magnitude of  $R_N$ 600  $(0 - 150 \text{ W m}^{-2}).$ 601

602 Scatter plots between  $g_C$  and  $\lambda E_T$  for different levels of  $D_A$  revealed a linear pattern between them for a wide range of  $D_A$  (20>  $D_A$  >0 hPa) (Fig. 8b). Following Monteith (1995), isopleths 603 of  $R_N$  are delineated by the solid lines passing through  $\lambda E_T$  on the x-axis and through  $g_C$  on 604 605 the y-axis. Isobars of  $D_A$  (dotted lines) pass through the origin because  $\lambda E_T$  approaches zero as  $g_C$  approaches zero. Figure (8b) shows substantial reduction of  $g_C$  with increasing  $D_A$ 606 without any increase of  $\lambda E_T$ , like an inverse hyperbolic pattern to  $D_A$  (Monteith 1995; Jones, 607 1998). For all the PFTs, an active biological (i.e., stomatal) regulation maintained almost 608 609 constant  $\lambda E_T$  when  $D_A$  was changed from low to high values (Fig. 8b). At high  $D_A$  (above 10 hPa), after an initial increase of  $\lambda E_T$  with  $g_C$ ,  $g_C$  approached a maximum limit and remained 610 nearly independent of  $\lambda E_T$  (Fig. 8b). Among all the  $D_A$  levels, the maximum control of  $g_C$  on 611 612  $\lambda E_T$  variability (62 to 80%) was found at high atmospheric water demand (i.e., 30 hPa> $D_A$ >20 hPa). The scatter plots between  $g_C$  and  $T_R$  (Fig. 8c) for different levels of  $D_A$ 613 614 revealed an exponential decline in  $g_C$  with increasing  $T_R$  and atmospheric water demand.

615 When retrieved  $g_A$  was plotted against the radiometric surface temperature and air 616 temperature difference  $(T_R - T_A)$ , an exponential decline in  $g_A$  was found in response to 617 increasing  $(T_R - T_A)$  (Fig. 8d). High  $g_A$  is persistent with low  $(T_R - T_A)$  irrespective of the 618 variations in  $R_N$  (with the exception of very low  $R_N$ ). Four negatively logarithmic scatters fit 619  $g_A$  versus  $(T_R - T_A)$  relationship with r values of 0.28 (150<  $R_N$  <300 W m<sup>-2</sup>), 0.55 (3000<  $R_N$ 620 <450 W m<sup>-2</sup>), 0.64 (450<  $R_N$  <600 W m<sup>-2</sup>), and 0.77 ( $R_N$  >600 W m<sup>-2</sup>).

#### 621 **5 Discussion**

#### 622 5.1 Evaluating $g_{A}$ , $g_{C}$ , and surface energy balance fluxes

The aerodynamic conductance retrieved with STIC1.2 showed acceptable correlation and 623 valid estimates of  $g_A$  when compared against an empirical model that uses  $u^*$  and u to derive 624  $g_A$  (Fig. 1 and 2a) and two other inversion/hybrid-based  $g_A$  estimates. The differences 625 between  $g_{A-STIC}$  and  $g_{A-BM13}$  were mainly attributed to the structural differences and empirical 626 nature of the parameterization for the near-surface boundary layer conductance 627  $((2/ku^{*2})(S_c/P_r)^{0.67})$  in  $g_{A-BM13}$ , which results in some discrepancies between  $g_{A-STIC}$  and  $g_{A-BM13}$ 628 particularly in the pasture (Fig. 2a). The extent to which the structural discrepancies between 629  $g_{A-STIC}$  and  $g_{A-BM13}$  relate to actual differences in the conductances for momentum vs. heat is 630 beyond the scope of this manuscript, and a detailed investigation using data on atmospheric 631 profiles of wind speed, temperature etc. are needed to actually quantify such differences. 632 Momentum transfer is associated with pressure forces and not identical to heat and mass 633 transfer (Massman, 1999). In STIC1.2,  $g_A$  is directly estimated and is a robust representative 634 of the conductances to heat/water vapor transfer; whereas  $g_{A-BMI3}$  estimates based on  $u^*$  and u635 636 is more representative for the momentum transfer. Therefore, the difference between the two different  $g_A$  estimates (Fig. 2) can be largely attributed to the actual difference in the 637 conductances for momentum and heat/water vapor. The turbulent conductance equation 638  $(u^{*2}/u)$  in  $g_{A-BM13}$  is also very sensitive to the uncertainties in the sonic anemometer 639

measurement (Contini et al., 2006; Richiardone et al., 2012). However, the evidence of a 640 weak systematic relationship between the  $g_A$  residuals and u (Fig. 2d) and capability of the 641 thermal  $(T_R)$ , radiative  $(\phi)$ , and meteorological  $(T_A, D_A)$  variables in capturing the variability 642 of  $g_{A-BM13}$  (Fig. 2e and 2f) indicates the diagnostic potential  $g_{A-STIC}$  estimates to explain the 643 wind driven  $g_A$  variability. Excluding u might introduce errors in cases where wind is the 644 645 only source of variations in  $g_A$  and surface fluxes (Mallick et al., 2015). In general, the accuracies in commonly used parametric  $g_A$  estimates based on u and surface roughness 646 parameters several meters distant from canopy foliage is are limited due to the uncertainties 647 concerning the attenuation of *u* close to the vegetation surface (Meinzer et al., 1997; Prihodko 648 et al., 2008). The magnitude of u near the foliage can be substantially lower than that 649 measured considerably away at some reference location above or within the canopy (Meinzer 650 et al., 1997). Notwithstanding the inequalities of  $g_A$  estimated with different methods, it is 651 challenging to infer the accuracy of the different estimates. It is imperative to mention that  $g_A$ 652 is one of the main anchors in the PM-SW model because it not only appears in the numerator 653 and denominator of these models,  $g_A$  also provides feedback to  $g_C$ ,  $T_0$ , and  $D_0$  (seminal paper 654 of Jarvis and McNaughton, 1986). Therefore, the estimates of  $\lambda E$  in the PM-SW framework 655 are very sensitive to parameterization of  $g_A$  and stable  $\lambda E$  estimates might be possible if  $g_A$ 656 estimation is unambiguous (Holwerda et al., 2012; van Dijk et al., 2015). Given the lack of 657 consensus in the community on the 'true'  $g_A$  and from the nature of surface flux validation 658 results (Fig. 4) it appears that  $g_{A-STIC}$  tends to be the appropriate aerodynamic conductance 659 that satisfies the PM-SW equation. Discrepancies between  $g_{C-STIC}$  and  $g_{C-INV}$  originated from 660 the differences in  $g_A$  estimates between the two methods. 661

662 Despite the good agreement between the measured and predicted  $\lambda E$  and H (Fig. 4, Table 4), 663 the larger error in H was associated with the higher sensitivity of H to the errors in  $T_R$  (due to 664 poor emissivity correction) (Mallick et al., 2015). Since the difference between  $T_R$  and  $T_A$  is 665

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considered to be the primary driving force of H (van der Tol et al., 2009), the modelled errors in H are expected to arise due to the uncertainties associated with  $T_R$ .

#### 5.2 Canopy coupling, $g_c$ and $g_A$ versus transpiration and evaporation

The correlation analysis between 1- $\Omega$  and  $\lambda E_T$  revealed the extent of biophysical and 668 radiative controls on  $\lambda E_T$  (Fig. 5). The degree of biophysical control is a function of the ratio 669 of  $g_C$  to  $g_A$ . Minor biophysical control on  $\lambda E_T$  was apparent for forest and pasture during the 670 wet seasons (Fig. 5c and 5d) as a result of a high  $g_C/g_A$  ratio along with increasing  $\lambda E_T$ . Such 671 conditions stimulate local humidification of air surrounding the canopy and uncoupling of the 672 in-canopy vapor pressure deficit  $(D_0)$  from that in the air above (i.e.,  $D_0 < D_A$ ) (Meinzer et al., 673 1997; Motzer et al., 2005) (Fig. 9a), which implies that  $\lambda E_T$  becomes largely independent of 674  $g_{C}$ . On the contrary, an enhanced biophysical control on  $\lambda E_{T}$  was apparent during the dry 675 season and drought year 2005 during the period of reduced water supply particularly over 676 PAS (Fig. 5e, 5f, and 7). Such condition leads to a relatively dry canopy surface, and 677 substantially high  $g_A$  compared to  $g_C$ , thus resulting in low  $g_C/g_A$  ratios regardless of their 678 679 absolute values (Meinzer et al., 1993; McNaughton and Jarvis, 1991). Here, fractional 680 changes in  $g_C$  results in an equivalent fractional change in  $\lambda E_T$ . This impedes transpiration from promoting local equilibrium of  $D_0$  and minimizing (or maximizing) the gradient 681 between  $D_0$  and atmospheric vapor pressure deficit  $(D_A)$  (i.e.,  $D_0 \cong D_A$  or  $D_0 > D_A$ ) (eqn. A10) 682 (Fig 9a), thereby resulting in strong coupling between  $D_0$  and  $D_A$  (Meinzer et al., 1993; Jarvis 683 684 and McNaughton, 1986). Besides, a supplemental biophysical control on  $\lambda E_T$  might have been imposed as a consequence of a direct negative feedback of  $D_A$  and  $D_0$  on  $g_C$ 685 (McNaughton and Jarvis, 1991; Jarvis, 1986). Increase in  $D_A$  (or  $D_0$ ) beyond a certain limit 686 decreases  $g_C$  (Fig. 7 and 8), resulting in a low and narrow increase of  $\lambda E_T$ , despite steady 687 increase in  $g_A$  and  $R_N$ . The combination of negative feedback response between  $D_A$  and  $g_C$ 688

689 with the overall radiative-aerodynamic coupling significantly dampens the variation of transpiration in PAS and TDF in the dry season, thus featuring increased biophysical control 690 691 in these PFTs. These results are in agreement with von Randow et al. (2012), who found enhanced biophysical control on  $\lambda E_T$  for the pasture during the dry season. For the wet 692 season, evidence of minor biophysical control indicates the dominance of  $R_N$  driven 693 equilibrium evaporation in these PFTs (Hasler and Avissar, 2007; da Rocha et al., 2009; 694 Costa et al., 2010). In the TRF and TMF, 94% and 99% of the retrieved  $g_C/g_A$  ratios fall 695 above 0.5, and, only 1% and 6% of the retrieved  $g_C/g_A$  ratios fall below the 0.5 range (Fig. 696 9b). In contrast, 90% and 73% of the  $g_C/g_A$  ratios range above 0.5, and 10% to 27% of the 697  $g_C/g_A$  ratios were below 0.5 for TDF and PAS, respectively (Fig. 9b). This shows that, 698 699 although radiation control is prevailing in all the sites, biophysical control is relatively stronger in TDF and PAS as compared to the other sites. For large  $g_C/g_A$  ratios, the conditions 700 within the planetary boundary layer (PBL) become decoupled from the synoptic scale 701 (McNaughton and Jarvis, 1991) and the net radiative energy becomes the important regulator 702 of transpiration. For small  $g_C/g_A$  ratios (e.g., in dry season), the conditions within the PBL are 703 704 strongly coupled to the atmosphere above by rapid entrainment of air from the capping inversion and by some ancillary effects of sensible heat flux on the entrainment 705 (McNaughton and Jarvis, 1991). These findings substantiate the earlier theory of 706 707 McNaughton and Jarvis (1991), who postulated that large  $g_C/g_A$  ratios result in minor biophysical control on canopy transpiration due to the negative feedback on the canopy from 708 709 the PBL. The negative relationship between 1- $\Omega$  and  $\lambda E_E$  (Fig. 5b) over all the PFTs is due to the feedback of  $g_A$  on  $g_C$ . However, over all the PFTs, a combined control of  $g_A$  and 710 environmental variables on  $\lambda E_E$  again highlighted the impact of realistically estimated  $g_A$  on 711  $\lambda E_E$  (Holwerda et al., 2012). 712

713 It is important to mention that forests are generally expected to be better coupled to the atmosphere, which is related to generally higher  $g_A$  (due to high surface roughness) compared 714 to the pastures. This implies that forests exhibit stronger biophysical control on  $\lambda E_T$ . 715 However, due to the broad leaves of the rain forests (larger leaf area index) and higher 716 surface wetness (due to higher rainfall amounts) the wet surface area is much larger in the 717 718 forests than in the pastures. This results in much higher  $g_C$  values for forests than for pastures 719 during the wet season ( $g_C \approx g_A$ ), and  $g_C/g_A \rightarrow 1$ . Consequently, no significant difference in coupling was found between them during the wet season (Fig. 5c and 5d). Despite the 720 absolute differences in  $g_A$  and  $g_C$  between forest and pasture, the high surface wetness is 721 largely offsetting the expected  $\Omega$  difference between them. Although the surface wetness is 722 substantially lower during the dry season, the high water availability in the forests due to the 723 deeper root systems help maintaining a relatively high  $g_C$  compared to the pastures. Hence, 724 725 despite  $g_A$  (forest) >  $g_A$  (pasture) during the dry season, substantially lower  $g_C$  values for the 726 pasture result in lower  $g_C/g_A$  ratio for the pasture compared to the forest, thus causing more 727 biophysical control on  $\lambda E_T$  during the dry season. The relatively better relationship between coupling versus  $\lambda E_T$  in PAS and TDF during the dry season was also attributed to high 728 729 surface air temperature difference  $(T_R - T_A)$  in these PFTs that resulted in low  $g_C/g_A$  ratios 730 (Fig. 9c).

#### 731 **5.3 Factors affecting** $g_c$ and $g_A$ variability

The stomatal feedback-response hypothesis (Monteith, 1995) also became apparent at the canopy-scale (Fig. 8a, 8b), which states that a decrease in  $g_C$  with increasing  $D_A$  is caused by a direct increase in  $\lambda E_T$  (Monteith, 1995; Matzner & Comstock, 2001; Streck, 2003) and  $g_C$ responds to the changes in the air humidity by sensing  $\lambda E_T$ , rather than  $D_A$ . This feedback mechanism is found because of the influence of  $D_A$  on both  $g_C$  and  $\lambda E_T$ , which in turn changes  $D_A$  by influencing the air humidity (Monteith, 1995). The change in  $g_C$  is dominated

by an increase in the net available energy, which is partially offset by an increase in  $\lambda E_T$ . 738 After the net energy input in the canopy exceeds a certain threshold,  $g_C$  starts decreasing even 739 if  $\lambda E_T$  increases. High  $\lambda E_T$  increases the water potential gradient between guard cells and 740 other epidermal cells or reduces the bulk leaf water potential, thus causing stomatal closure 741 (Monteith, 1995; Jones, 1998; Streck, 2003). The control of soil water on transpiration also 742 became evident from the scatter plots between  $g_C$  versus  $\lambda E_T$  and  $T_R$  for different  $D_A$  levels 743 (Fig. 8b, 8c) (also Fig. 7). Denmead and Shaw (1962) hypothesized that reduced  $g_C$  and 744 stomatal closure occurs at moderate to higher levels of soil moisture (high  $\lambda E_T$ ) when the 745 atmospheric demand of water vapor increases (high  $D_A$ ). The water content in the immediate 746 vicinity of the plant root depletes rapidly at high  $D_A$ , which decreases the hydraulic 747 conductivity of soil, and the soil is unable to efficiently supply water under these conditions. 748 749 For a given evaporative demand and available energy, transpiration is determined by the  $g_C/g_A$  ratio, which is further modulated by the soil water availability. These combined effects 750 tend to strengthen the biophysical control on transpiration (Leuzinger and Kirner, 2010; 751 752 Migletta et al., 2011). The complex interaction between  $g_C$ ,  $T_R$ , and  $D_A$  (Fig. 8c) explains why different parametric  $g_C$  models produce divergent results. 753

754 Although  $\lambda E_T$  and  $\lambda E_E$  estimates are interdependent on  $g_C$  and  $g_A$  (as shown in Fig. 6 to Fig. 8); the figures reflect the credibility of the conductances as well as transpiration estimates by 755 realistically capturing the hysteretic behavior between biophysical conductances and water 756 vapor fluxes, which is frequently observed in natural ecosystems (Zhang et al., 2014, Renner 757 et al., 2016) (also Zuecco et al., 2016). These results are also compliant with the theories 758 postulated earlier from observations that the magnitude of hysteresis depends on the 759 760 radiation-vapor pressure deficit time-lag, while the soil moisture availability is a key factor modulating the hysteretic transpiration-vapor pressure deficit relation as soil moisture 761 declines (Zhang et al., 2014; O'Grady et al., 1999; Jarvis and McNaughton, 1986). This 762

shows that despite being independent of any predefined hysteretic function, theinterdependent conductance-transpiration hysteresis is still captured in STIC1.2.

Fig. 8d is in accordance with existing theory that under conditions of extremely high atmospheric turbulence (i.e., high  $g_A$ ), a close coupling exists between the surface and the atmosphere, which causes  $T_R$  and  $T_A$  to converge (i.e.,  $T_R - T_A \rightarrow 0$ ). When  $g_A$  is low, the difference between  $T_R$  and  $T_A$  increases due to poor vertical mixing of the air.

# 769 6 Conclusions

By integrating the radiometric surface temperature ( $T_R$ ) into a combined structure of PM-SW model, we have estimated the canopy-scale biophysical conductances and quantified their control on the terrestrial evapotranspiration components in a simplified SEB modeling perspective that treats the vegetation canopy as 'big-leaf'. The STIC1.2 biophysical modeling scheme is independent of any leaf-scale empirical parameterisation for stomata and associated aerodynamic variables.

Stomata regulate the coupling between terrestrial carbon and water cycles, which implies that 776 their behaviour under global environmental change is critical to predict vegetation 777 functioning (Medlyn et al., 2011). The combination of variability in precipitation (Hilker et 778 779 al., 2014) and land cover change (Davidson et al., 2012) in the Amazon Basin is expected to increase the canopy-atmosphere coupling of pasture or forest systems under drier conditions 780 781 by altering the ratio of the biological and aerodynamic conductances. An increase of 782 biophysical control will most likely be an indicator of shifting the transpiration from an 783 energy-limited to a water-limited regime (due to the impact of  $T_R$ ,  $T_A$ , and  $D_A$  on the  $g_C/g_A$ ratio) with further consequences for the surface water balance and rainfall recycling. At the 784 785 same time, a transition from forest to pasture or agriculture lands will substantially reduce the 786 contribution of interception evaporation in the Amazon, hence, it will affect the regional water cycle. This might change the moisture regime of the Amazonian Basin and affect the 787

moisture transport to other regions. In this context, STIC1.2 provides a new quantitative and internally consistent method for interpreting the biophysical conductances and able to quantify their controls on the water cycle components in response to a range of climatic and ecohydrological conditions (excluding rising atmospheric  $CO_2$ ) across a broad spectrum of PFTs. It could also provide the basis to improve existing land surface parameterisations for simulating vegetation water use at large spatial scales.

It should also be noted that although the case study described here provides general insights into the biophysical controls of  $\lambda E$  and associated feedback between  $g_C$ ,  $D_A$ ,  $T_R$  and  $\lambda E_T$  in the framework of the PM-SW equation, there is a tendency for overestimation of  $g_C$  due to the embedded evaporation information in the current single-source composition of STIC1.2. For accurate characterisation of canopy conductance, explicit partitioning of  $\lambda E$  into transpiration and evaporation (both soil and interception) is one of the further scopes for improving STIC1.2 and this assumption needs to be tested further.

## 801 Acknowledgements

The developed modeling framework contributes to the "Catchments As Organized Systems 802 (CAOS)" Phase-2 research group (FOR 1598) funded by the German Science Foundation 803 (DFG) and to the "HiWET (High-resolution modelling and monitoring of Water and Energy 804 Transfers in wetland ecosystems)" consortium funded by BELSPO and FNR. We sincerely 805 806 thank Dr. Andrew Jarvis (Lancaster University, UK), Dr. Monica Garcia (Technical University of Denmark, Denmark), and Dr. Georg Wohlfahrt (University of Innsbruck, 807 Austria) for very helpful discussions and edits in the manuscript. We are grateful to all 808 809 Brazilian and international collaborators and all the funding agencies that have contributed to the Large-scale Biosphere Atmosphere Experiment in Amazônia (LBA). The authors are 810 indebted to Pavel Kabat, Antônio Ocimar Manzi, David R. Fitzjarrald, Julio Tota, Humberto 811 Ribeiro da Rocha, Michael Goulden, Maarten J. Waterloo and Luiz Martinelli for planning, 812

coordinating, conducting, and evaluating the eddy covariance, meteorological and leaf gas
exchange measurements at the LBA sites. We are particularly grateful to all field technicians
whose hard work were the key ingredients to establish the quality of the datasets used in this
paper. The authors declare no conflict of interest. DTD acknowledges support of the Jet
Propulsion Laboratory, California Institute of Technology, under a contract with the National
Aeronautics and Space Administration.

## 819 Appendix A:

#### 820 A1 Derivation of 'state equations' in STIC 1.2

Neglecting horizontal advection and energy storage, the surface energy balance equation iswritten as follows:

$$\phi = \lambda E + H \tag{A1}$$

Figure (A1) shows that, while *H* is controlled by a single aerodynamic resistance  $(r_A)$  (or 1/g<sub>A</sub>);  $\lambda E$  is controlled by two resistances in series, the surface resistance  $(r_C)$  (or 1/g<sub>C</sub>) and the aerodynamic resistance to vapor transfer  $(r_C + r_A)$ . For simplicity, it is implicitly assumed that the aerodynamic resistance of water vapor and heat are equal (Raupach, 1998), and both the fluxes are transported from the same level from near surface to the atmosphere. The sensible and latent heat flux can be expressed in the form of aerodynamic transfer equations (Boegh et al., 2002; Boegh and Soegaard, 2004) as follows:

$$H = \rho c_P g_A (T_o - T_A) \tag{A2}$$

$$\lambda E = \frac{\rho c_P}{\gamma} g_A(e_0 - e_A) = \frac{\rho c_P}{\gamma} g_C(e_0^* - e_0)$$
(A3)

830 Where  $T_0$  and  $e_0$  are the air temperature and vapor pressure at the source/sink height (i.e., 831 aerodynamic temperature and vapor pressure) or at the so-called roughness length ( $z_0$ ), where 832 wind speed is zero. They represent the vapor pressure and temperature of the quasi-laminar 833 boundary layer in the immediate vicinity of the surface level (Fig. A1), and  $T_0$  can be obtained by extrapolating the logarithmic profile of  $T_A$  down to  $z_0$ .  $e_0^*$  is the saturation vapor pressure at  $T_0$  (hPa).

By combining eqn. (A1), (A2), and (A3) and solving for  $g_A$ , we get the following equation.

$$g_A = \frac{\phi}{\rho c_P \left[ (T_o - T_A) + \left(\frac{e_0 - e_A}{\gamma}\right) \right]}$$
(A4)

837 Combining the aerodynamic expressions of  $\lambda E$  in eqn. (A3) and solving for  $g_C$ , we can 838 express  $g_C$  in terms of  $g_A$ ,  $e_0^*$ ,  $e_0$ , and  $e_A$ .

$$g_C = g_A \frac{(e_0 - e_A)}{(e_0^* - e_0)}$$
(A5)

839 While deriving the expressions for  $g_A$  and  $g_C$ , two more unknown variables are introduced ( $e_0$ 840 and  $T_0$ ), thus there are two equations and four unknowns. Therefore, two more equations are 841 needed to close the system of equations.

842 An expression for  $T_0$  is derived from the Bowen ratio ( $\beta$ ) (Bowen, 1926) and evaporative 843 fraction ( $\Lambda$ ) (Shuttleworth et al., 1989) equation.

$$\beta = \left(\frac{1-\Lambda}{\Lambda}\right) = \frac{\gamma(T_0 - T_A)}{(e_0 - e_A)} \tag{A6}$$

$$T_o = T_A + \left(\frac{e_0 - e_A}{\gamma}\right) \left(\frac{1 - \Lambda}{\Lambda}\right) \tag{A7}$$

844 This expression for  $T_0$  introduces another new variable (A); therefore, one more equation that describes the dependence of  $\Lambda$  on the conductances ( $g_A$  and  $g_C$ ) is needed to close the system 845 846 of equations. In order to express  $\Lambda$  in terms of  $g_A$  and  $g_C$ , we had adopted the advection – aridity (AA) hypothesis (Brutsaert and Stricker, 1979) with a modification introduced by 847 (Mallick et al., 2015). The AA hypothesis is based on a complementary connection between 848 the potential evaporation  $(E^*)$ , sensible heat flux (H), and E; and leads to an assumed link 849 between  $g_A$  and  $T_0$ . However, the effects of surface moisture (or water stress) were not 850 explicit in the AA equation and Mallick et al. (2015) implemented a moisture constraint in 851 the original advection-aridity hypothesis while deriving a 'state equation' of  $\Lambda$  (eqn. A8) 852

below). A detailed derivation of the 'state equation' for  $\Lambda$  is described in the Supplement (S1) (also see Mallick et al., 2014, 2015). Estimation of  $e_0$ ,  $e_0^*$ , M, and  $\alpha$  is described in the Appendix (A2).

$$\Lambda = \frac{2\alpha s}{2s + 2\gamma + \gamma \frac{g_A}{g_C}(1+M)}$$
(A8)

#### 856 <u>A2 Iterative solution of $e_0$ , $e_0^*$ , M, and $\alpha$ in STIC 1.2</u>

In STIC1.0 and 1.1 (Mallick et al., 2014; 2015), no distinction was made between the surface 857 and source/sink height vapor pressures. Therefore,  $e_0^*$  was approximated as the saturation 858 vapor pressure at  $T_R$  and  $e_0$  was empirically estimated from M based on the assumption that 859 the vapor pressure at the source/sink height ranges between extreme wet-dry surface 860 conditions. However, the level of  $e_0$  and  $e_0^*$  should be consistent with the level of the 861 aerodynamic temperature  $(T_0)$  from which the sensible heat flux is transferred (Lhomme and 862 Montes, 2014). The predictive use of the PM model could be hindered due to neglecting the 863 feedbacks between the surface layer evaporative fluxes and source/sink height mixing and 864 coupling (McNaughton and Jarvis, 1984), and their impact on the canopy scale conductances. 865 Therefore, in STIC1.2, we have used physical expressions for estimating  $e_0$  and  $e_0^*$  followed 866 by estimating  $T_{SD}$  and M as described below. The fundamental differences between STIC1.0, 867 1.1 and 1.2 modeling philosophy is are described in Table A1. 868

869 An estimate of  $e_0^*$  is obtained by inverting the aerodynamic transfer equation of  $\lambda E$ .

$$e_0^* = e_A + \left[\frac{\gamma \lambda E(g_A + g_C)}{\rho c_P g_A g_C}\right] \tag{A9}$$

870 Following Shuttleworth and Wallace (1985) (SW85), the vapor pressure deficit ( $D_0$ ) (=  $e_0^*$  -871  $e_0$ ) and vapor pressure ( $e_0$ ) at the source/sink height are expressed as follows.
$$D_0 = D_A + \left[\frac{\{s\phi - (s+\gamma)\lambda E\}}{\rho c_P g_A}\right]$$
(A10)

$$e_0 = e_0^* - D_0 \tag{A11}$$

A physical equation of  $\alpha$  is derived by expressing the evaporative fraction ( $\Lambda$ ) as function of the aerodynamic equations of  $H\left[\rho c_P g_A(T_0 - T_A)\right]$  and  $\lambda E\left[\frac{\rho c_P}{\gamma} \frac{g_A g_C}{g_A + g_C}(e_0^* - e_A)\right]$  as follows.

$$\Lambda = \frac{\lambda E}{H + \lambda E} \tag{A12}$$

$$= \frac{\frac{\rho c_P}{\gamma} \frac{g_A g_C}{g_A + g_C} (e_0^* - e_A)}{\rho c_P g_A (T_0 - T_A) + \frac{\rho c_P}{\gamma} \frac{g_A g_C}{g_A + g_C} (e_0^* - e_A)}$$
(A13)  
$$= \frac{g_C (e_0^* - e_A)}{[\gamma (T_0 - T_A) (g_A + g_C) + g_C (e_0^* - e_A)]}$$
(A14)

## 874 Combining eqn. (A14) and eqn. (A8) (eliminating $\Lambda$ ), we can derive a physical equation of $\alpha$ .

$$\alpha = \frac{g_C(e_0^* - e_A) \left[ 2s + 2\gamma + \gamma \frac{g_A}{g_C} (1+M) \right]}{2s[\gamma(T_0 - T_A)(g_A + g_C) + g_C(e_0^* - e_A)]}$$
(A15)

Following Venturini et al. (2008), *M* can be expressed as the ratio of the vapor pressuredifference to the vapor press deficit between surface to atmosphere as follows.

$$M = \frac{(e_0 - e_A)}{(e_0^* - e_A)} = \frac{(e_0 - e_A)}{\kappa(e_S^* - e_A)} = \frac{s_1(T_{SD} - T_D)}{\kappa s_2(T_R - T_D)}$$
(A16)

Where  $T_{SD}$  is the dewpoint temperature at source/sink height and  $T_D$  is the air dewpoint temperature;  $s_1$  and  $s_2$  are the psychrometric slopes of the saturation vapor pressure and temperature between  $(T_{SD} - T_D)$  versus  $(e_0 - e_A)$  and  $(T_R - T_D)$  versus  $(e_s^* - e_A)$  relationship (Venturini et al., 2008); and  $\kappa$  is the ratio between  $(e_0^* - e_A)$  and  $(e_s^* - e_A)$ . Despite  $T_0$  drives the sensible heat flux, the comprehensive dry-wet signature of underlying surface due to soil moisture variations is directly reflected in  $T_R$  (Kustas and Anderson, 2009). Therefore, using 883  $T_R$  in the denominator of eqn. (A16) tend<u>s</u> to give a direct signature of the surface moisture 884 availability (*M*). In eqn. (A16),  $T_{SD}$  computation is challenging because both  $e_0$  and  $s_1$  are 885 unknown. By decomposing the aerodynamic equation of  $\lambda E$ ,  $T_{SD}$  can be expressed as follows.

$$\lambda E = \frac{\rho c_P}{\gamma} g_A(e_0 - e_A) = \frac{\rho c_P}{\gamma} g_A s_1 (T_{SD} - T_D)$$

$$T_{SD} = T_D + \frac{\gamma \lambda E}{\rho c_P g_A s_1}$$
(A17)

In the earlier STIC versions,  $s_1$  was approximated at  $T_D$ ,  $e_0^*$  was approximated at  $T_R$ ,  $T_{SD}$  was estimated from  $s_1$ ,  $T_D$ ,  $T_R$ , and related saturation vapor pressures (Mallick et al., 2014; 2015), and M was estimated from eqn. (A16) (estimation of  $T_{SD}$  and M was stand-alone earlier). However, since  $T_{SD}$  depends on  $\lambda E$  and  $g_A$ , an iterative procedure is applied to estimate  $T_{SD}$ and M as described below.

In STIC1.2, an initial value of  $\alpha$  is assigned as 1.26 and initial estimates of  $e_0^*$  and  $e_0$  are 891 obtained from  $T_R$  and M as  $e_0^* = 6.13753 e^{\frac{17.27T_R}{(T_R + 237.3)}}$  and  $e_0 = e_A + M(e_0^* - e_A)$ . Initial  $T_{SD}$ 892 and M were estimated as described above. With the initial estimates of these variables; first 893 estimate of the conductances,  $T_0$ ,  $\Lambda$ , and  $\lambda E$  are obtained. The process is then iterated by 894 updating  $e_0^*$  (using eqn. A9),  $D_0$  (using eqn. A10),  $e_0$  (using eqn. A11),  $T_{SD}$  (using eqn. A17) 895 with  $s_1$  estimated at  $T_D$ ), M (using eqn. A16), and  $\alpha$  (using eqn. A15), with the first estimates 896 of  $g_C$ ,  $g_A$ , and  $\lambda E$ , and recomputing  $g_C$ ,  $g_A$ ,  $T_0$ ,  $\Lambda$ , and  $\lambda E$  in the subsequent iterations with the 897 previous estimates of  $e_0^*$ ,  $e_0$ ,  $T_{SD}$ , M, and  $\alpha$  until the convergence  $\lambda E$  is achieved. Stable 898 values of  $\lambda E$ ,  $e_0^*$ ,  $e_0$ ,  $T_{SD}$ , M, and  $\alpha$  are obtained within ~25 iterations. Illustrative examples 899 of the convergence of  $e_0^*$ ,  $e_0$ ,  $T_{SD}$ , M, and  $\alpha$  are shown in Fig. (A3). 900

901 To summarize, the computational steps of the conductances and evaporative fluxes in STIC902 are:

903	Step 1: Analytical solution of the conductances, $T_0$ and $\Lambda$ by solving the 'state equations'
904	(eqn. 2, 3, 4, and 5). Step 2: Initial estimates of the conductances ( $g_c$ and $g_A$ ), $T_0$ , $A$ , $\lambda E$ and
905	<i>H. Step 3: Simultaneous iteration of</i> $\lambda E$ <i>,</i> $e_0^*$ <i>,</i> $e_0$ <i>,</i> $T_{SD}$ <i>,</i> $M$ <i>, and</i> $\alpha$ <i>; and final estimation of the</i>
906	conductances ( $g_C$ and $g_A$ ), $T_0$ , $\Lambda$ , $\lambda E$ and $H$ . Step 4: Partitioning $\lambda E$ into $\lambda E_T$ and $\lambda E_E$ .
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Variables	Description
and	
symbol	
$\lambda E$	Evapotranspiration (evaporation + transpiration) as latent heat flux (W m <sup>-2</sup> )
Н	Sensible heat flux (W m <sup>-2</sup> )
$R_N$	Net radiation (W m <sup>-2</sup> )
G	Ground heat flux (W m <sup>-2</sup> )
φ	Net available energy (W m <sup>-2</sup> )
$T_A$	Air temperature (°C)
$T_D$	Dewpoint temperature (°C)
$T_R$	Radiometric surface temperature (°C)
$R_H$	Relative humidity (%)
$e_A$	Atmospheric vapor pressure at the level of $T_A$ measurement (hPa)
$D_A$	Atmospheric vapor pressure deficit at the level of $T_A$ measurement (hPa)
и	Wind speed (m $s^{-1}$ )
<i>u</i> *	Friction velocity (m s <sup>-1</sup> )
$T_{SD}$	Dew-point temperature at the source/sink height (°C)
$T_0$	Aerodynamic temperature or source/sink height temperature (°C)
$e_S$	'effective' vapor pressure of evaporating front near the surface (hPa)
$e_{S}^{*}$	Saturation vapor pressure of surface (hPa)
$e_0^*$	Saturation vapor pressure at the source/sink height (hPa)
$e_0$	Atmospheric vapor pressure at the source/sink height (hPa)
$D_0$	Atmospheric vapor pressure deficit at the source/sink height (hPa)
$\lambda E_{eq}$	Equilibrium latent heat flux (W m <sup>-2</sup> )
$\lambda E_{imp}$	Imposed latent heat flux (W m <sup>-2</sup> )
$\lambda E_E$	Evaporation as flux (W m <sup>-2</sup> )
$\lambda E_T$	Transpiration flux (W m <sup>-2</sup> )
Ε	Evapotranspiration (evaporation + transpiration) as depth of water (mm)
$\lambda E^{*}$	Potential evaporation as flux (W m <sup>-2</sup> )
$\lambda E_T^*$	Potential transpiration as flux (W m <sup>-2</sup> )
$\lambda E_W$	Wet environment evaporation as flux (W m <sup>-2</sup> )
$\lambda E_P^*$	Potential evaporation as flux according to Penman (W m <sup>-2</sup> )
$\lambda E_{PM}^{*}$	Potential evaporation as flux according to Penman-Monteith (W m <sup>-2</sup> )
$\lambda E_{PT}^{*}$	Potential evaporation as flux according to Priestley-Taylor (W m <sup>-2</sup> )
$E^{*}$	Potential evaporation as depth of water (mm)
$E_P^*$	Potential evaporation as depth of water according to Penman (mm)
$E_{PM}^{*}$	Potential evaporation as depth of water according to Penman-Monteith (mm)
$E_{PT}^{*}$	Potential evaporation as depth of water according to Priestley-Taylor (mm)
$E_W$	Wet environment evaporation as depth of water (mm)
$g_A$	Aerodynamic conductance (m s <sup>-1</sup> )
<i>g</i> <sub>C</sub>	Stomatal / surface conductance (m s <sup>-1</sup> )
$g_M$	Momentum conductance (m s <sup>-1</sup> )
$g_B$	Quasi-laminar boundary layer conductance (m s <sup>-1</sup> )
8 <sub>Cmax</sub>	Maximum stomatal / surface conductance (m s <sup>-1</sup> ) (= $g_C/M$ )

**Table 1:** Variables and symbols and their description used in the present study.

М	Surface moisture availability $(0-1)$
S	Slope of saturation vapor pressure versus temperature curve (hPa $K^{-1}$ ) (estimated at $T_A$ )
<i>S</i> <sub>1</sub>	Slope of the saturation vapor pressure and temperature between $(T_{SD} - T_D)$ versus $(e_0 - e_A)$
	(approximated at $T_D$ ) (hPa K <sup>-1</sup> )
<i>s</i> <sub>2</sub>	Slope of the saturation vapor pressure and temperature between $(T_R - T_D)$ versus $(e_S^* - e_A)$
	(hPa K <sup>-1</sup> ), estimated according to Mallick et al. (2015).
<b>S</b> 3	Slope of the saturation vapor pressure and temperature between $(T_R - T_{SD})$ versus $(e_S^* - e_S)$
	(approximated at $T_R$ ) (hPa K <sup>-1</sup> )
К	Ratio between $(e_0^* - e_A)$ and $(e_s^* - e_A)$
λ	Latent heat of vaporization of water (j kg $^{-1}$ K $^{-1}$ )
$Z_R$	Reference height (m)
$Z_M$	Effective source-sink height of momentum (m)
$Z_0$	Roughness length (m)
d	Displacement height (m)
γ	Psychrometric constant (hPa K <sup>-1</sup> )
ρ	Density of air (kg m <sup>-3</sup> )
$c_p$	Specific heat of dry air (MJ kg <sup>-1</sup> K <sup>-1</sup> )
Λ	Evaporative fraction (unitless)
$\beta$	Bowen ratio (unitless)
α	Priestley-Taylor parameter (unitless)
Ω	Decoupling coefficient (unitless)
$S_c$	Schmidt number (unitless)
$P_r$	Prandtl number (unitless)
k	Von Karman's constant (0.4)

1241	Table 2: Overview of the LBA tower sites. Here, (-) refers to (S) and (W) for latitude and longitude
1242	respectively.

	-							
Biome	PFT	Site	LBA Code	Data availability period	Latitude	Longitude (°)	Tower height (m)	Annual rainfall (mm)
Forest	Tropical rainforest (TRF)	Manaus KM34	K34	06/1999 to 09/2006	-2.609	-60.209	50	2329
Forest	Tropical moist forest (TMF)	Santarem KM67	K67	01/2002 to 01/2006	-2.857	-54.959	63	1597
Forest	Tropical moist forest (TMF)	Santarem KM83	K83	07/2000 to 12/2004	-3.018	-54.971	64	1656
Forest	Tropical dry forest (TDF)	Reserva Biológica Jarú	RJA	03/1999 to 10/2002	-10.083	-61.931	60	2354
Pasture	Pasture (PAS)	Santarem KM77	K77	01/2000 to 12/2001	-3.012	-54.536	18	1597
Pasture	Pasture (PAS)	Fazenda Nossa Senhora	FNS	03/1999 to 10/2002	-10.762	-62.357	8.5	1743

**Table 3**: Comparative statistics for the STIC and tower-derived hourly  $g_A$  and  $g_C$  for a range of PFTs

in the Amazon Basin (LBA tower sites). Values in parenthesis are  $\pm$  one standard deviation (standard

1260 error for correlation).

PFTs		$g_A$	-STIC VS. $g_{A-I}$	3M13		$g_{C-STIC}$ VS. $g_{C-INV}$			
	RMSD	$R^2$	Slope	Offset	Ν	RMSD	$\mathbf{R}^2$	Slope	Offset
	$(m s^{-1})$		•	$(m s^{-1})$		$(m s^{-1})$			$(m s^{-1})$
TRF	0.013	0.41	1.07	0.0031	1159	0.012	0.14	0.39	0.0097
		(±0.03)	$(\pm 0.047)$	$(\pm 0.0008)$			$(\pm 0.04)$	(±0.039)	(±0.0007)
TMF	0.012	0.55	0.81	0.0006	1927	0.009	0.55	0.85	0.0032
		(±0.12)	(±0.023)	(±0.0006)			(±0.12)	(±0.025)	(±0.0005)
TDF	0.007	0.49	0.89	0.0019	787	0.012	0.33	0.30	0.0050
		(±0.15)	(±0.041)	(±0.0006)			(±0.19)	(±0.022)	(±0.0005)
PAS	0.012	0.22	1.03	0.0059	288	0.007	0.58	0.65	0.0024
		(±0.18)	(±0.083)	(±0.0007)			(±0.12)	(±0.025)	(±0.0003)
Mean	0.012	0.44	0.76	0.0047	4161	0.010	0.39	0.63	0.0046
		(±0.10)	(±0.016)	(±0.003)			$(\pm 0.08)$	(±0.016)	(±0.0003)

1261 N = number of data points; RMSD = root mean square deviation between predicted (P) and observed (O) 1262 variables =  $\left[\frac{1}{N}\sum_{i=0}^{N}(P_i - O_i)^2\right]^2$ .

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**Table 4**: Comparative statistics for the STIC and tower-derived hourly  $\lambda E$  and *H* for a range of PFTs in the Amazon Basin (LBA tower sites). Values in parenthesis are ±one standard deviation (standard error for correlation).

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PFIS		ΛI	5		Н				
	RMSD	$\mathbb{R}^2$	Slope	Offset	RMSD	$R^2$	Slope	Offset	N
	$(W m^{-2})$			$(W m^{-2})$	$(W m^{-2})$			$(W m^{-2})$	
TRF	28	0.96	1.10	-16	34	0.52	0.60	29	1159
		(±0.007)	$(\pm 0.008)$	(±2)		(±0.030)	(±0.025)	(±2)	
TMF	20	0.98	1.08	-11	23	0.71	0.61	20	1927
		$(\pm 0.004)$	$(\pm 0.004)$	(±1)		(±0.019)	(±0.014)	(±1)	
TDF	26	0.96	0.96	-7	30	0.66	0.89	20	787
		(±0.009)	$(\pm 0.008)$	(±2)		(±0.032)	(±0.035)	(±3)	
PAS	31	0.96	1.14	-2	33	0.88	0.67	9	288
		(±0.009)	(±0.010)	(±2)		(±0.016)	(±0.011)	(±1)	
Mean	33	0.94	1.04	-1	37	0.61	0.58	24 (±2)	4161
		(±0.005)	(±0.005)	(±1)		(±0.021)	(±0.009)		

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**Figure 1**. Examples of monthly averages of the diurnal time series of canopy-scale (a)  $g_A$  and (b)  $g_C$ estimated for two different biomes (forest and pasture) in the Amazon Basin (LBA sites K34 and FNS). The time series of four different  $g_A$  estimates and their corresponding  $g_C$  estimates are shown here.



(a) Time series  $g_A$ 

**Figure 2.** (a) Comparison between STIC derived  $g_A$  ( $g_{A-STIC}$ ) with an estimated aerodynamic conductance based on friction velocity ( $u^*$ ) and wind speed (u) according to Baldocchi and Ma (2013) ( $g_{A-BMI3}$ ), (b) Comparison between  $g_{A-STIC}$  with an inverted  $g_A$  ( $g_{A-INV}$ ) based on EC observations of  $\lambda E$ and  $D_A$ , (c) Comparison between  $g_{A-STIC}$  with a hybrid  $g_A$  ( $g_{A-INV}$ ) based on EC observations of H and estimated  $T_0$  over the LBA EC sites, (d) Comparison between residual  $g_A$  differences versus u and  $u^*$ , (e) and (f) Relationship between wind and shear derived  $g_A$  versus  $\phi$ ,  $T_R$ , and  $D_A$  over the LBA EC sites.



**Figure 3.** (a) Comparison between STIC derived  $g_C(g_{C-STIC})$  and  $g_C$  computed by inverting the PM model ( $g_{C-INV}$ ) over the LBA EC sites, where  $g_{A-BMI3}$  was used as aerodynamic input in conjunction with tower measurements of  $\lambda E$ , radiation and meteorological variables, (b) Residual  $g_C$  differences versus wind speed (u) and friction velocity ( $u^*$ ) over the LBA EC sites.



**Figure 4.** Comparison between STIC derived (a)  $\lambda E$  and (b) *H* over four different PFTs in the Amazon Basin (LBA tower sites). MAPD is the percent error defined as the mean absolute deviation between predicted and observed variable divided by mean observed variable.



**Figure 5.** Correlation of coupling  $(1-\Omega)$  with (a) transpiration  $(\lambda E_T)$  and (b) evaporation  $(\lambda E_E)$  and over four different PFTs by combining data for all the years, only during dry seasons for all the years, and during drought year 2005. Data for 2005 was not available for TDF and PAS. (c) to (e) Examples of diurnal pattern of  $\Omega$  (black lines),  $\lambda E_E$  (grey dotted lines) and  $\lambda E_T$  (grey solid lines) estimated over two ecohydrologically contrasting biomes (K34 for forest and FNS for pasture) in the Amazon Basin (LBA tower sites) during wet and dry seasons.



**Figure 6**. Scatter plots of transpiration ( $\lambda E_T$ ) and evaporation ( $\lambda E_E$ ) versus  $g_C$  and  $g_A$  over four different PFTs in the Amazon Basin (LBA tower sites).





**Figure 7.** Illustrative examples of the occurrence of diurnal hysteresis of transpiration ( $\lambda E_T$ ) during wet and dry seasons with canopy and environmental controls over two different sites with different annual rainfall (2329 mm and 1597 mm, respectively) in the Amazon Basin (LBA tower sites K34 and FNS).



(a) Rainforest: Wet season

(b) Pasture: Wet season

**Figure 8.** (a) Response of retrieved  $g_C$  to atmospheric vapor pressure deficit  $(D_A)$  for different classes of net radiation  $(R_N)$ , (b) Response of retrieved  $g_C$  to transpiration for different classes of  $D_A$ , (c) Response of retrieved  $g_C$  to radiometric surface temperature  $(T_R)$  for different classes  $D_A$ , (d) Relationship between retrieved  $g_A$  and radiometric surface temperature and air temperature difference  $(T_R - T_A)$  in the Amazon Basin (LBA tower sites).





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**Figure 9**. (a) Scatter plots between source/sink height (or in-canopy) vapor pressure deficit ( $D_0$ ) and atmospheric vapor pressure deficit ( $D_A$ ) for two different classes of  $g_C/g_A$  ratios over four PFTs, which clearly depicts a strong coupling between  $D_0$  and  $D_A$  for low  $g_C/g_A$  ratios. (b) Histogram distribution of  $g_C/g_A$  ratios over the four PFTs in the Amazon Basin (LBA tower sites). (c) Scatter plots between  $g_C/g_A$  ratio versus surface air temperature difference ( $T_R - T_A$ ) for the four PFT during wet season and dry season in the Amazon Basin (LBA tower sites).



(a)  $D_0$  vs.  $D_A$  over four PFTs

(b) Distribution of  $g_C/g_A$  ratio over four PFTs

(c)  $g_C/g_A$  vs.  $T_R$ - $T_A$  over four PFTs



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**Table A1**: Differences in the modeling philosophy of source/sink height vapor pressures  $(e_0, e_0^*)$  and dewpoint temperature  $(T_{SD})$ , surface wetness (M), and  $\alpha$  between STIC1.0, STIC1.1 and STIC1.2.

	Principles							
Variable estimation	STIC1.0 (Mallick et al., 2014)	STIC1.1 (Mallick et al., 2015)	STIC1.2 (This study [Mallick et al., 2016])					
Saturation vapor pressure at source/sink height $(e_0^*)$	$e_0^*$ was approximated as the saturation vapor pressure at $T_R$ .	Same as STIC1.0	$e_0^*$ is estimated through numerical iteration by inverting the aerodynamic equation of $\lambda E$ (as described in appendix A2). $e_0^* = e_A + \left[\frac{\gamma \lambda E(g_A + g_C)}{\rho c_P g_A g_C}\right]$					
Actual vapor pressure at source/sink height $(e_0)$	$e_0$ was empirically estimated from $M$ based on the assumption that the vapor pressure at the source/sink height ranges between extreme wet–dry surface conditions.	Same as STIC1.0	$e_0 \text{ is estimated as } e_0 = e_0^* - D_0, \text{ where } D_0$ was iteratively estimated by combining PM with Shuttleworth-Wallace approximation (as described in appendix A2). $D_0 = D_A + \left[\frac{\{s\phi - (s + \gamma)\lambda E\}}{\rho c_P g_A}\right]$					
Dewpoint temperature at source/sink height ( $T_{SD}$ )	$T_{SD} = \frac{(e_S^* - e_A) - s_3 T_R + s_1 T_D}{(s_1 - s_3)}$ s <sub>1</sub> and s <sub>3</sub> are the slopes of saturation vapor pressures at temperatures, approximated at $T_D$ and $T_R$ , respectively.	Same as STIC1.0	$T_{SD}$ is estimated through numerical iteration by inverting the aerodynamic equation of $\lambda E$ (as described in appendix A2). $T_{SD} = T_D + \frac{\gamma \lambda E}{\rho c_P g_A s_1}$					
Surface moisture availability ( <i>M</i> )	As a stand-alone equation, without any feedback to $\lambda E$ .	Same as STIC1.0	A feedback of $M$ into $\lambda E$ is introduced and $M$ is iteratively estimated after estimating $T_{SD}$ (as described in appendix A2).					
Priestley- Taylor parameter (α)	As fixed parameter (1.26).	A physical equation of $\alpha$ is derived as a function of the conductances and $\alpha$ is numerically estimated as a variable.	A physical equation of $\alpha$ is derived as a function of the conductances and $\alpha$ is numerically estimated as a variable (eqn. A15) (as described in appendix A2).					

**Table A2**: Fundamental differences in the modeling principles between STIC1.2 and previous approaches for characterising the biophysical controls on  $\lambda E$  components.

Biophysical states	Modeling principles					
	Parametric modeling (Ma et al., 2015; Chen et al., 2011; Kumagai et al., 2004)	STIC1.2				
<i>8A</i>	Either $g_A$ is assumed to be the momentum conductance $(g_M)$ or estimated as a sum of $g_M$ and quasilaminar boundary-layer conductance $(g_B)$ . $l/g_A = l/g_M + l/g_B$ $g_M = u^*/u$ $g_B = f{Nusselt number, leaf dimension, thermalconductivity of air in boundary layer, u, kinematicviscosity, Reynolds number}If u^* is available from EC tower, it is directlyused, otherwise u^* is parametrized using Monin-Obukhov Similarity Theory (MOST).Disadvantages: (1) MOST is only valid for anextended, uniform, and flat surface (Foken,2006). MOST tends to fail over rough surfacesdue to breakdown of the similarity relationshipsfor heat and water vapor transfer in the roughnesssub-layer, which results in an underestimation ofthe 'true' g_A by a factor 1-3 (Thom et al., 1975;Chen and Schwerdtfeger, 1988; Simpson et al.,1998; Holwerda et al., 2012). (2) In the state-of-art \lambda E modeling, the parametric g_A sub-modelsare stand alone and empirical, and do not provideany feedback to g_C, aerodynamic temperature(T_0), and aerodynamic vapor pressures (e_0 andD_0). (3) Additional challenges in grid-scale orspatial-scale g_A estimation are the requirementsof numerous site specific parameters (e.g.,vegetation height, measurement height,vegetation roughness, leaf size, soil roughness)and coefficients needed to correct theatmospheric stability conditions (Raupach, 1998).$	Analytically retrieved by solving 'n' state equations and 'n' unknowns, with explicit convective feedback and without any wind speed (u) information. In a hallmark paper by Choudhury and Monteith (1986), it is clearly stated that 'aerodynamic conductance determined by wind speed and roughness is assumed to be unaffected by buoyancy. Strictly, the aerodynamic conductance should be replaced by a term which accounts for radiative as well as convective heat transfer'. The role of $g_A$ is associated with the role of convection (Choudhury and Monteith, 1986) according to the surface energy balance principle as reflected in the derivation of eqn. (A4). Wind is generated as a result of the differences in atmospheric pressure which is a result of uneven surface radiative heating. Therefore, the aerodynamic conductance (and wind as well) is an effect of net radiative heating and there should be a physical relationship between $T_R$ and $\lambda E$ , coupled with canopy-atmosphere components relating $\lambda E$ to $T_0$ and $e_0$ . (2) Supports the findings of Villani et al. (2003) which stated that during unstable surface layer conditions the major source of net available energy is located at the canopy top and drives the convective motion in the layers above.				
<i>8c</i>	<ul> <li>(a) If λE measurements are available from the EC towers, g<sub>C</sub> is estimated by inverting the PM equation. None of these approaches allow independent quantification of biophysical controls of λE as g<sub>C</sub> is constrained by λE itself.</li> <li>(b) Sometimes g<sub>C</sub> is modelled either by coupled leaf-scale photosynthesis models (Ball et al., 1987; Leuning, 1995) or g<sub>C</sub> is estimated from standalone empirical models (Larvis, 1976)</li> </ul>	Analytically retrieved by solving 'n' state equations and 'n' unknowns where physical feedbacks of $g_A$ , soil moisture, and vapor pressure deficit are embedded (as explained in STIC1.2 equations in Appendix).				

1387 Figure A1. Schematic representation of one-dimensional description of STIC1.2. In STIC1.2, a 1388 feedback is established between the surface layer evaporative fluxes and source/sink height mixing and coupling, and the connection is shown in dotted arrows between  $e_0$ ,  $e_0^*$ ,  $g_A$ ,  $g_C$ , and  $\lambda E$ . Here,  $r_A$ 1389 and  $r_c$  are the aerodynamic and canopy (or surface in case of partial vegetation cover) resistances,  $g_A$ 1390 and  $g_c$  are the aerodynamic and canopy conductances (reciprocal of resistances),  $e_s^*$  is the saturation 1391 vapor pressure of the surface,  $e_0^*$  is the saturation vapor pressure at the source/sink height,  $T_0$  is the 1392 1393 source/sink height temperature (i.e. aerodynamic temperature) that is responsible for transferring the 1394 sensible heat (H),  $e_0$  is the source/sink height vapor pressure,  $e_s$  is the vapor pressure at the surface,  $z_0$ 1395 is the roughness length,  $T_R$  is the radiometric surface temperature,  $T_{SD}$  is the source/sink height 1396 dewpoint temperature, M is the surface moisture availability or evaporation coefficient,  $R_N$  and G are 1397 net radiation and ground heat flux,  $T_A$ ,  $e_A$ , and  $D_A$  are temperature, vapor pressure, and vapor pressure 1398 deficit at the reference height  $(z_R)$ ,  $\lambda E$  is the latent heat flux, H is the sensible heat flux, respectively.



Single-source Surface energy balance representation

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**Figure A2.** Aerodynamic temperature obtained from STIC1.2 ( $T_{0-STIC}$ ) versus radiometric surface temperature ( $T_R$ ) over two different biomes in the Amazon basin. The regression equation of line of best fit is  $T_{0-STIC} = 0.67(\pm 0.10)T_R + 10.59 (\pm 2.79)$  with r = 0.65.



1419 **Figure A3.** (a) Convergence of the iteration method for retrieving the source/sink height (or in-1420 canopy) vapor pressures ( $e_0$  and  $D_0$ ) and Priestley-Taylor coefficient ( $\alpha$ ). (b) Convergence of the 1421 iteration method for retrieving the surface wetness (M) and source/sink height dewpoint temperature 1422 ( $T_{SD}$ ). The initial values of  $\lambda E$ ,  $g_A$ ,  $g_C$ , and  $T_0$  were determined with  $\alpha = 1.26$ . The process is then 1423 iterated by updating  $\lambda E$ ,  $e_0$ ,  $D_0$ , M,  $T_{SD}$ , and  $\alpha$  in subsequent iterations with the previous estimates of 1424  $g_A$ ,  $g_C$ , and  $T_0$ .



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1428 Supplement of

## 1429 Canopy-scale biophysical controls of transpiration and evaporation in the 1430 Amazon Basin

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## 1458 S1. Derivations of evaporative fraction ( $\Lambda$ ) 'state equation' in STIC1.2

In order to express  $\Lambda$  in terms of  $g_A$  and  $g_C$ , we had adopted the advection – aridity hypothesis (Brutsaert and Stricker, 1979) with a modification introduced by Mallick et al. (2015). Although the advection–aridity hypothesis leads to an assumed link between  $g_A$  and  $T_0$ , the effects of surface moisture (or water stress) were not explicit in the advection–aridity equation. Mallick et al. (2015) implemented a moisture constraint in the original advection– aridity hypothesis for deriving an expression of  $\Lambda$ . A modified form of the original advection– aridity hypothesis is written as follows.

$$E_{PM}^* = 2E_{PT}^* - E (S1)$$

Here  $E_{PM}^*$  is the potential evapotranspiration according to Penman-Monteith (Monteith, 1965) for any surface, and  $E_{PT}^*$  is the potential evapotranspiration according to Priestley-Taylor (Priestley and Taylor, 1972). Dividing both sides by *E* we get,

$$\frac{E}{E_{PM}^*} = \frac{E}{2E_{PT}^* - E} \tag{S2}$$

1469 and dividing the numerator and denominator of the right hand side of eqn. (S2) by  $E_{PT}^*$  we 1470 get,

$$\frac{E}{E_{PM}^*} = \frac{\frac{E}{E_{PT}^*}}{2 - \frac{E}{E_{PT}^*}}$$
(S3)

1471 Again assuming the Priestley-Taylor equation for any surface is a variant of the PM potential 1472 evapotranspiration equation, we will derive an expression of  $E_{PT}^*$  for any surface.

$$E_{PM}^{*} = \frac{s\phi + \rho c_{P} g_{A} D_{A}}{s + \gamma \left(1 + \frac{g_{A}}{g_{cmax}}\right)}$$
(S4)
$$= \frac{s\phi}{s + \gamma \left(1 + \frac{g_A}{g_{cmax}}\right)} \left(1 + \frac{\rho c_P g_A D_A}{s\phi}\right)$$
$$= \frac{\alpha s\phi}{s + \gamma \left(1 + \frac{g_A}{g_{cmax}}\right)}$$
$$= E_{PT}^*$$
(S5)

Here  $\gamma$  is the psychrometric constant (hPa K<sup>-1</sup>), s is the slope of the saturation vapor pressure 1473 versus air temperature (hPa K<sup>-1</sup>),  $\alpha$  is the Priestley-Taylor parameter ( $\alpha = 1.26$  under non-1474 limiting moisture conditions),  $D_A$  is the vapor pressure deficit of air (hPa).  $g_{Cmax}$  is defined as 1475 the maximum possible  $g_C$  under the prevailing atmospheric conditions whereas  $g_C$  is limited 1476 1477 due to the moisture availability (M) and hence  $g_{Cmax} = g_C/M$  (Monteith, 1995; Raupach, 1998). We assume that M is a significant controlling factor for the ratio of actual and 1478 1479 potential evapotranspiration (or transpiration for a dry canopy), and the interactions between the land and environmental factors are substantially reflected in M. Since, Penman (1948) 1480 derived his equation over the open water surface and  $g_{Cmax}$  over the water surface is very high 1481 (Monteith, 1965; 1981),  $g_A/g_{Cmax}$  was assumed to be negligible. 1482

1483 Expressing  $\phi$  as  $\phi = E/\Lambda$  and expressing  $E_{PT}^*$  according to eqn. (S5) gives the following 1484 expression of  $E/E_{PT}^*$ .

$$\frac{E}{E_{PT}^*} = \frac{\Lambda \left[ s + \gamma \left( 1 + \frac{g_A}{g_{cmax}} \right) \right]}{\alpha s}$$
(S6)

1485 Now substituting  $E/E_{PT}^*$  from eqn. (S6) into eqn. (S3) and after some algebra we obtain the 1486 following expression.

$$\frac{E}{E_{PM}^{*}} = \frac{\Lambda \left[ s + \gamma \left( 1 + \frac{g_{A}}{g_{cmax}} \right) \right]}{2\alpha s - \Lambda \left[ s + \gamma \left( 1 + \frac{g_{A}}{g_{cmax}} \right) \right]}$$
(S7)

## 1487 According to the PM equation (Monteith, 1965) of actual and potential evapotranspiration,

$$\frac{E}{E_{PM}^{*}} = \frac{\frac{s\phi + \rho c_{p}g_{A}D_{A}}{s + \gamma \left(1 + \frac{g_{A}}{g_{C}}\right)}}{\frac{s\phi + \rho c_{p}g_{A}D_{A}}{s + \gamma \left(1 + \frac{g_{A}}{g_{cmax}}\right)}}$$
(S8)

1488 Combining eqn. (S7) and (S8) (eliminating  $E/E_{PM}^*$ ) gives an expression for  $\Lambda$  in terms of the 1489 conductances.

$$\frac{s+\gamma\left(1+\frac{Mg_A}{g_C}\right)}{s+\gamma\left(1+\frac{g_A}{g_C}\right)} = \frac{\Lambda\left[s+\gamma\left(1+\frac{Mg_A}{g_C}\right)\right]}{2\alpha s-\Lambda\left[s+\gamma\left(1+\frac{Mg_A}{g_C}\right)\right]}$$
(S9)

## 1490 After some algebra the final expression of $\Lambda$ is as follows.

$$\Lambda = \frac{2\alpha s}{2s + 2\gamma + \gamma \frac{g_A}{g_C}(1+M)}$$
(S10)

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