Reviewer 2 (R2):

This manuscript describes a study that infers stomatal and aerodynamic conductances from eddy flux observations. I think in general, this study is innovative and presents novel material, so that in principle it should be published. I hesitate recommendation for publication mostly because I am not entirely convinced by the approach and I feel that this needs revision. Hence, I recommend major revisions, although I do not think that it necessarily involves a lot of work to address the points below.

Response: We thank R2 for the encouraging comments and for appreciating the novelty of the aerodynamic and canopy conductance (g_A and g_C) retrieval to assess their controls on evaporation and transpiration. We appreciate the valuable suggestions which will further improve the manuscript.

Major points:

(1) My major problem with the manuscript is that I do not understand the approach, so that it is difficult to assess its plausibility. While the main equations are provided in the manuscript (eqn. 2-5), there is no more description on where these equations come from, except for references to prior papers by the authors. I think it is necessary to at least provide a description at a qualitative level where these equations come from.

Response: We agree to include the derivations of eqns. 2-5 in the Appendix of the revised manuscript. A detailed derivation of the STIC1.2 state equations (eqns. 2-5) is given below.

Formulation of STIC was based on the goal to find an analytical solution of the two unobserved 'state variables' (i.e., aerodynamic and canopy conductances) (g_A and g_C) in the Penman-Monteith (PM) equation while exploiting the radiative (net radiation and ground heat flux), meteorological (air temperature, humidity), and radiometric surface temperature (T_R) as external inputs. **The fundamental assumption in STIC is the first order dependence of g_A and g_C on the aerodynamic temperature (T_0) and soil moisture (through the radiometric surface temperature, T_R).** These assumptions allow direct integration of T_R into the PM equation and simultaneously constrain the conductances. **Given T_R is the direct signature of the soil moisture availability, inclusion of T_R in the PM equation also works to add water stress controls in g_C.**

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

$$\phi = \lambda E + H \qquad (1)$$

Where $\phi \cong R_N - G$, with R_N being net radiation, and G being the conductive surface heat flux or ground heat flux, H is the sensible heat flux and λE is the latent heat flux (or evapotranspiration, E).

According to Figure A1 in the manuscript, while the sensible heat flux is controlled by a single aerodynamic resistance (r_A) (or $1/g_A$); the water vapor flux is controlled by two resistances in series, the surface resistance (r_C) (or $1/g_C$) 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:

$$H = \rho c_P g_A (T_o - T_A) \tag{2}$$

$$\lambda E = \frac{\rho c_P}{\gamma} g_A (e_0 - e_A) = \frac{\rho c_P}{\gamma} g_C (e_0^* - e_0) \tag{3}$$

Where ρ is the density of dry air (kg m^{-3}), c_P is the specific heat of dry air (MJ kg^{-1} K^{-1}), γ is the psychrometric constant (hPa K^{-1}), T_A is the air temperature at the reference height (z_R), e_A is the atmospheric vapor pressure (hPa) at the level at which T_A is measured, e_0 and T_0 are the atmospheric vapor pressure and air temperature at the source/sink height (i.e., aerodynamic temperature), or at the so-called roughness length (z_0), where wind speed is zero. They represent the vapor pressure and temperature of the quasi-laminar boundary layer in the immediate vicinity of the surface level (Figure 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 eq. 1, 2 and 3 and solving for g_A, we get the following equation.

$$g_A = \frac{\phi}{\rho c_P \left[(T_o - T_A) + \left(\frac{e_0 - e_A}{\gamma} \right) \right]} \tag{4}$$

Combining the aerodynamic expressions of λE in eq. 3 and solving for g_C, we can express g_C in terms of g_A, e_0^*, e_0, and e_A.

$$g_C = g_A \frac{(e_0 - e_A)}{(e_0^* - e_0)} \tag{5}$$

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

An expression for T_0 is derived from the Bowen ratio (β) (Bowen, 1926) and evaporative fraction (Λ) (Shuttleworth et al., 1989) equation.

$$\beta = \left(\frac{1 - \Lambda}{\Lambda} \right) = \frac{\gamma (T_0 - T_A)}{(e_0 - e_A)} \tag{6}$$

$$T_o = T_A + \left(\frac{e_0 - e_A}{\gamma} \right) \left(\frac{1 - \Lambda}{\Lambda} \right) \tag{7}$$

This expression for T_0 introduces another new variable (Λ**); therefore, one more equation that describes the dependence of Λ on the conductances (g_A and g_C) is needed to close the system of equations.** The detailed derivation of an expression for Λ is described in Mallick et al. (2014, 2015) and this is briefly described below. Estimation of e_0 is based on numerical iteration as described in the Appendix of the manuscript and is also described in the response (3) below.

In order to express Λ 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) was not explicit in the advection–aridity equation. We implemented a moisture constraint in the original advection-aridity hypothesis for deriving an expression of Λ. The logic of using the advection-aridity hypothesis for finding an expression of Λ is described in Mallick et al. (2014). A modified form of the original advection-aridity hypothesis is written as follows.

$$E_{PM}^{*} = 2E_{PT}^{*} - E \qquad (8)$$

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,

$$\frac{E}{E_{PM}^{*}} = \frac{E}{2E_{PT}^{*} - E} \qquad (9)$$

and dividing the numerator and denominator of the right hand side of eqn. 9 by E_{PT}^{*} we get,

$$\frac{E}{E_{PM}^{*}} = \frac{\frac{E}{E_{PT}^{*}}}{2 - \frac{E}{E_{PT}^{*}}} \qquad (10)$$

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

$$E_{PM}^{*} = \frac{s\phi + \rho c_P g_A D_A}{s + \gamma\left(1 + \frac{g_A}{g_{Cmax}}\right)} \qquad (11)$$

$$= \frac{s\phi}{s + \gamma\left(1 + \frac{g_A}{g_{Cmax}}\right)}\left(1 + \frac{\rho c_P g_A D_A}{s\phi}\right)$$

$$= \frac{\alpha s\phi}{s + \gamma\left(1 + \frac{g_A}{g_{Cmax}}\right)} \qquad (12)$$

$$= E_{PT}^{*}$$

Here γ is the psychrometric constant (hPa K^{-1}), s is the slope of the saturation vapor pressure versus air temperature (hPa K^{-1}), α is the Priestley-Taylor parameter (α =1.26 under non-limiting moisture conditions), D_A is the vapor pressure deficit of air (hPa). g_{Cmax} is defined as the maximum possible g_C under the prevailing atmospheric conditions whereas g_C is limited 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 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) derived his equation over the open water surface and g_{Cmax} over the water surface is very high (Monteith, 1965; 1981), g_A/g_{Cmax} was assumed to be negligible.

Expressing ϕ as ϕ = E/Λ and expressing E_{PT}^{*} according to eqn. 12 gives the following expression of E/E_{PT}^{*}.

$$\frac{E}{E^*_{PT}} = \frac{\Lambda\left[s + \gamma\left(1 + \frac{g_A}{g_{Cmax}}\right)\right]}{\alpha s} \tag{13}$$

Now substituting E/E^*_{PT} from eqn. 13 into eq. 10 and after some algebra we obtain the following expression.

$$\frac{E}{E^*_{PM}} = \frac{\Lambda\left[s + \gamma\left(1 + \frac{g_A}{g_{Cmax}}\right)\right]}{2\alpha s - \Lambda\left[s + \gamma\left(1 + \frac{g_A}{g_{Cmax}}\right)\right]} \tag{14}$$

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

$$\frac{E}{E^*_{PM}} = \frac{\dfrac{s\phi + \rho c_p g_A D_A}{s + \gamma\left(1 + \frac{g_A}{g_C}\right)}}{\dfrac{s\phi + \rho c_p g_A D_A}{s + \gamma\left(1 + \frac{g_A}{g_{Cmax}}\right)}} \tag{15}$$

Combining eqn. 14 and 15 (eliminating E/E^*_{PM}) gives an expression for Λ in terms of the conductances.

$$\frac{s + \gamma\left(1 + \frac{M g_A}{g_C}\right)}{s + \gamma\left(1 + \frac{g_A}{g_C}\right)} = \frac{\Lambda\left[s + \gamma\left(1 + \frac{M g_A}{g_C}\right)\right]}{2\alpha s - \Lambda\left[s + \gamma\left(1 + \frac{M g_A}{g_C}\right)\right]} \tag{16}$$

After some algebra the final expression of Λ is as follows.

$$\Lambda = \frac{2\alpha s}{2s + 2\gamma + \gamma\frac{g_A}{g_C}(1 + M)} \tag{17}$$

Given the information of R_N, G, T_A, and R_H or e_A, **the four state equations (eqns. 4, 5, 7, and 17) can be solved simultaneously to derive analytical solutions for the four state variables. The analytical solutions to the state equations 4, 5, 7, and 17 still have four additional unknowns; M, e_0, e_0^*, and α, and these variables are iteratively estimated as described in the Appendix of the current manuscript. For estimating M, we have extensively used the radiometric surface temperature (T_R) in a physical retrieval framework in STIC1.2, thus treating T_R as an external input.**

(2) The point where I really got confused is that eqn. 5 uses the Priestley-Taylor coefficient, which is an empirical coefficient in an evaporation equation that is rather different from the Penman Monteith equation. Where does this coefficient suddenly come from? I find this quite confusing, and it needs at least a minimum of explanation as it is not obvious.

Response: Good point indeed and we apologise for the confusion.

From the derivation of the equation 17 above (eqn. 5 in the manuscript), it is apparent that the Priestley-Taylor coefficient (α) appeared due to the use of the Advection-Aridity hypothesis for deriving the state equation of the evaporative fraction. However, instead of assuming α as a 'fixed parameter', we have developed a physical equation of α (eqn. A8 in the manuscript) and numerically estimated α as a 'variable'. **The derivation of the equation for α is described is the following response** (also in the Appendix of the manuscript in line 609 to 611).
We will make this description more explicit in the revised version of the manuscript.

(3) What I also do not understand is why an iterative scheme is needed.

Response: **The analytical solution to the above state equations 4, 5, 7, and 17 (eqn. 2 – 5 in the manuscript) have four accompanying unknowns; M (surface moisture availability), e_0 (vapor pressure at the source/sink height), e_0^* (saturation vapor pressure at the source/sink height), and α, and as a result there are 4 equations with 8 unknowns. Consequently an iterative solution is needed to determine the four unknown variables (as described below).**

An estimate of e_0^* is obtained by inverting the aerodynamic transfer equation of λE.

$$e_0^* = e_A + \left[\frac{\gamma \lambda E(g_A + g_C)}{\rho c_P g_A g_C}\right] \tag{18}$$

Following Shuttleworth and Wallace (1985) (SW85, hereafter), the vapor pressure deficit (D_0) ($=e_0^*$ - e_0) and vapor pressure (e_0) at the source/sink height are expressed as follows.

$$D_0 = D_A + \left[\frac{\{s\phi - (s+\gamma)\lambda E\}}{\rho c_P g_A}\right] \tag{19}$$

$$e_0 = e_0^* - D_0 \tag{20}$$

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

$$\Lambda = \frac{\lambda E}{H + \lambda E} \tag{21}$$

$$= \frac{\frac{\rho c_P}{\gamma}\frac{g_A g_C}{g_A + g_C}(e_0^* - e_A)}{\rho c_P g_A (T_0 - T_A) + \frac{\rho c_P}{\gamma}\frac{g_A g_C}{g_A + g_C}(e_0^* - e_A)} \tag{22}$$

$$= \frac{g_C(e_0^* - e_A)}{[\gamma(T_0 - T_A)(g_A + g_C) + g_C(e_0^* - e_A)]} \tag{23}$$

Combining eqn. 23 and eqn. 17 (eliminating Λ), we can derive a physical expression of α.

$$\alpha = \frac{g_C(e_0^* - e_A)\left[2s + 2\gamma + \gamma\frac{g_A}{g_C}(1+M)\right]}{2s[\gamma(T_0 - T_A)(g_A + g_C) + g_C(e_0^* - e_A)]} \tag{24}$$

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

$$M = \frac{(e_0 - e_A)}{(e_0^* - e_A)} = \frac{s_1(T_{SD} - T_D)}{s_2(T_0 - T_D)} \tag{25}$$

Where T_{SD} is the dewpoint temperature of the evaporating front (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_0 – T_D) versus (e_0^* - e_A) relationship (Venturini et al., 2008). Since T_0 is not available and T_R and e_A are available, we compute s_2 as s_2 = (e_S^* - e_A)/(T_R – T_D) with the assumption that errors due to any inequality between T_0 versus T_R and e_0^* versus e_S^* tend to be cancelled out in this ratio. This appears to be a valid assumption due to the close relationship between T_0 and T_R (Huband and Monteith, 1986). 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 T_R in the denominator of eqn. 25 gives a direct signature of the surface moisture availability (M). In eqn. 25, T_{SD} computation is challenging because both e_0 and s_1 are unknown. By decomposing the aerodynamic equation of λE, T_{SD} can be expressed as follows.

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

$$T_{SD} = T_D + \frac{\gamma \lambda E}{\rho c_P g_A s_1} \tag{26}$$

In the earlier STIC versions, s_1 was approximated at T_D, 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. 25 (estimation of T_{SD} and M was stand-alone earlier). However, since T_{SD} depends on λE and g_A, an iterative procedure is applied in STIC1.2 to estimate T_{SD} and M as described below, which is another modification of the STIC1.0 and STIC1.1.

In STIC1.2, an initial value of α is assigned as 1.26 and initial estimates of e_0^* and e_0 are obtained from T_R and M as $e_0^* = 6.13753e^{\frac{17.27T_R}{(T_R+237.3)}}$ and $e_0 = e_A + M(e_0^* - e_A)$. Initial T_{SD} and M were estimated as described above. With the initial estimates of these variables; first estimate of the conductances, T_0, Λ, and λE are derived. The process is then iterated by updating D_0 (using eqn. 19), e_0^* (using eqn. 18), e_0 (using eqn. 20), T_{SD} (using eqn. 26 with s_1 estimated at T_D), M [M = s_1(T_{SD} – T_D)/s_2(T_R – T_D)], and α (using eqn. 24), with the first estimates of g_C, g_A, and λE, and recomputing g_A, g_C, T_0, Λ, and λE in the subsequent iterations with the previous estimates of e_0^*, e_0, T_{SD}, M, and α until the convergence λE is achieved. Stable values of λE, e_0^*, e_0, T_{SD}, M, and α are obtained within ~25 iterations.

The above equations are previously included in the appendix of the current manuscript.

(4) Can't one simply use the observations and use a simple partitioning based on the Bowen ratio?

Response: Here we intended to partition evapotranspiration into component water fluxes. Although the Bowen ratio (Bowen, 1926) is an energy partitioning ratio to understand the relative apportioning between sensible and latent heat flux, it is not relevant for the latent heat flux partitioning into transpiration and evaporation. In this context an aggregated surface moisture availability (or water stress factor) is a better metric for dry-wet latent heat flux partitioning and we used the retrieved surface moisture availability (M) for partitioning of the latent heat flux.

(5) It would be good to describe what the differences and similarities are to previous approaches. As the authors propose a new approach, they should provide a better description that is easier to follow of what is being done.

Response: We assume R2 is intending to the differences of STIC with other approaches that earlier attempted to understand the biophysical controls of evapotranspiration, which is briefly described in the table below.

Biophysical states	Modeling principles	
	Parametric (Ma et al., 2015; Kumagai et al., 2004)	Nonparametric (STIC)
g_A	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). $1/g_A = 1/g_M + 1/g_B$ $g_M = f\{u^*, \textit{wind speed})$ g_B = *f{Nusselt number, leaf dimension, thermal conductivity of air in boundary layer, wind speed, kinematic viscosity, Reynolds number}* If u* is available from EC tower, it is directly used, otherwise u* is estimated using Monin-Obukhov Similarity Theory (MOST). MOST is only valid for an extended, uniform, and flat surface (Foken, 2006)	Analytically retrieved by solving 'n' state equations and 'n' unknowns, with explicit convective feedback.
g_C	(a) If λE measurements are available from the EC towers, g_C is estimated by inverting the PM equation. This leads to circularity. Since λE observations are used to obtain g_C, the same g_C should not be used to assess the biophysical controls of λE. (b) If λE measurements are not available from the EC towers (i.e., at grid-scale or spatial scale), g_C is modelled either by coupling with leaf-scale photosynthesis models (Ball et al., 1987; Leuning, 1995) or g_C is estimates from standalone empirical models (Jarvis, 1976)	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 the STIC1.2 equations).

If R2 is intending the differences between STIC1.2 with other previous versions, we propose to include a table in the appendix to describe the fundamental differences between STIC1.0, STIC1.1, and STIC1.2. The Table is given below.

Variable estimation	Principle		
	STIC1.0 (Mallick et al., 2014)	STIC1.1 (Mallick et al., 2015)	STIC1.2 (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 λE (as described in the appendix of the manuscript). $e_0^* = e_A + \left[\frac{\gamma \lambda E (g_A + g_C)}{\rho c_P g_A g_C}\right]$
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 is estimated as $e_0 = e_0^* - D_0$, where D_0 was iteratively estimated by combining PM with Shuttleworth-Wallace approximation (as described in the appendix of the manuscript). $D_0 = D_A + \left[\frac{\{s\phi - (s+\gamma)\lambda E\}}{\rho c_P g_A}\right]$
Dewpoint temperature at the 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_1 and s_3 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 λE (as described in the appendix of the manuscript). $T_{SD} = T_D + \frac{\gamma \lambda E}{\rho c_P g_A s_1}$
Surface moisture availability (M)	As a stand-alone equation, without any feedback to λE.	Same as STIC1.0	A feedback of M into λE is introduced and M is iteratively estimated after estimating T_{SD} (as described in the appendix of the manuscript).
Priestley-Taylor parameter (α)	As fixed parameter (1.26).	A physical equation of α is derived as a function of the conductances and α is numerically estimated as a variable.	A physical equation of α is derived as a function of the conductances and α is numerically estimated as a variable.

Minor points:

- The authors refer to λE as evaporation, which, technically speaking, is the latent heat flux, not evaporation.

Response: Necessary corrections will be incorporated in the revised manuscript.

- Abstract: dry and wet conditions λE_T, do you mean conditions in which water is not limiting vs. limiting, or precipitation vs. radiation driven conditions?

Response: It is the precipitation vs. radiation driven conditions. We will clarify this in the abstract.

- Biophysical control of λE_T should be briefly explained by what this means.

Response: Aerodynamic (physical) and stomatal (biological) conductances (g_A and g_C) together impose substantial biophysical controls on λE_T. At large g_A and small g_C, the vapor pressure deficit close to the canopy source/sink height (D_0) changes in response to the transpiration rate caused due to changes in the atmospheric vapor pressure deficit (D_A) or g_A. This results in strong canopy-atmosphere coupling and such condition is prevalent under soil moisture deficient conditions. On the other hand large g_C minimizes the gradients of vapor pressure deficit just above the canopy, such that D_0 tend towards zero and remains independent of any change in transpiration rate caused by changes in D_A or g_A. This substantially weakens the canopy-atmosphere coupling and such situation prevails under predominantly wet conditions.
We shall include this description in the introduction of the revised manuscript.

- Line 145: I wonder why approaches that directly link stomatal conductance to photosynthesis are not mentioned, such as Ball-Berry?

Response: We shall include references to photosynthesis-dependent stomatal conductance models in the revised manuscript.

- Line 194: Where do these “state equations” come from? Referring to previously published work is fine for derivations, but the description should still mention what the concepts are that are behind these equations.

Response: As discussed earlier we agree to include a description of the derivation in the revised manuscript.

- A table of variables would help.

Response: A table of variables will be included.

- Line 238: I think the authors assume that the conductances to momentum, sensible and latent heat are identical. If this is the case, it should be mentioned, as there are also approaches to surface exchange that do not treat them as being identical.

Response: Yes, the conductances of momentum for the sensible and latent heat flux are assumed identical. We will mention this in the revised manuscript after equation 7.

- Line 331: As the typical readers of HESS are not micrometeorologists, it would be useful to explain the decoupling coefficient in some more detail. This will help to interpret the following results.

Response: The decoupling coefficient or factor Omega (Ω) is a dimensionless coefficient ranging from 0.0 to 1.0 (Jarvis and McNaughton, 1986) and considered as an index of the degree of stomatal control on transpiration. The equation of Ω is as follows.

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

The Ω form of the Penman-Monteith (PM) equation for evapotranspiration is as follows.

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

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

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

Where, λE_{eq} is the equilibrium evapotranspiration, which depends only on the net available energy and would be obtained over an extensive surface of uniform moisture availability (Jarvis and McNaughton, 1986; Kumagai et al., 2004). λE_{imp} is the imposed evapotranspiration, which is 'imposed' by the atmosphere on the vegetation surface through the effects of vapor pressure deficit (triggered under limited soil moisture availability) and evapotranspiration is proportional to g_C.
When the g_C/g_A ratio is very small (i.e., water stressed conditions), stomata principally control the water loss and a change in g_C will result in a nearly proportional change in transpiration. In this case the Ω 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., water unstressed conditions), changes in g_C will have little effect on the transpiration rate, and transpiration is predominantly controlled by the net available radiative energy. In this case the Ω value approaches unity, and vegetation is considered to be poorly coupled to the atmosphere.
We will add this description in the revised version of the manuscript.

- Line 422: To what extent could these discrepancies between how conductances are derived also relate to actual differences in the conductances for momentum vs. heat?

Response: This is indeed a good point addressed by R2 (although beyond the scope of this manuscript) and will be clarified in the revised version. However, a detailed investigation using data on atmospheric profiles of wind speed, temperature etc. are needed to actually quantify such differences.
Momentum transfer is associated with pressure forces and not identical to heat and mass transfers (Massman, 1999). In principle, the aerodynamic conductances for heat and mass transfers are assumed equal (Monteith, 1965, 1981). In STIC1.2, g_A is directly estimated (as described previously) and is a robust representative of the resistance to heat/water vapor transfer. The parametric g_A estimates based on the friction velocity and wind speed is more representative for momentum transfer. **Therefore, the difference between the two different g_A estimates (Fig. 2) is primarily due to the actual difference in the conductances for momentum and heat/water vapor.**

- Line 498: The authors should stick to the same ratio gA/gC for easier interpretation.

Response: We will correct this.

Reference:

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