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A simple explanation for the sensitivity of the hydrologic cycle to surface temperature and solar radiation and its implications for global climate change

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Abstract. The global hydrologic cycle is likely to increase in strength with global warming, although some studies indicate that warming due to solar absorption may result in a different sensitivity than warming due to an elevated greenhouse effect. Here we show that these sensitivities of the hydrologic cycle can be derived analytically from an extremely simple surface energy balance model that is constrained by the assumption that vertical convective exchange within the atmosphere operates at the thermodynamic limit of maximum power. Using current climatic mean conditions, this model predicts a sensitivity of the hydrologic cycle of 2.2 % K^{-1} to greenhouse-induced surface warming which is the sensitivity reported from climate models. The sensitivity to solar-induced warming includes an additional term, which increases the total sensitivity to 3.2 % K^{-1}. These sensitivities are explained by shifts in the turbulent fluxes in the case of greenhouse-induced warming, which is proportional to the change in slope of the saturation vapor pressure, and in terms of an additional increase in turbulent fluxes in the case of solar radiation-induced warming. We illustrate an implication of this explanation for geoengineering, which aims to undo surface temperature differences by solar radiation management. Our results show that when such an intervention compensates surface warming, it cannot simultaneously compensate the changes in hydrologic cycling because of the differences in sensitivities for solar vs. greenhouse-induced surface warming. We conclude that the sensitivity of the hydrologic cycle to surface temperature can be understood and predicted with very simple physical considerations but this needs to reflect on the different roles that solar and terrestrial radiation play in forcing the hydrologic cycle.

1 Introduction

The hydrologic cycle plays a critical role in the physical functioning of the earth system, as the phase changes of liquid water to vapor require and release substantial amounts of heat. Currently, as climate is changing due to the enhanced greenhouse effect and surface warming, we would expect the hydrologic cycle to change as well. The most direct effect of such surface warming is that the saturation vapor pressure of near-surface air would increase, which should enhance surface evaporation rates if moisture does not limit evaporation. For current surface conditions, the saturation vapor pressure of air would on average increase at a rate of about 6.5 % K^{-1}. However, climate model simulations predict a mean sensitivity of the hydrologic cycle (or, hydrologic sensitivity) to global warming of about 2.2 % K^{-1} (Allen and Ingram, 2002; Held and Soden, 2006; Allan et al., 2013), with some variation among models. This sensitivity is also reported for climate model simulations of the last ice age (Boos, 2012; Li et al., 2013), and is commonly explained in terms of radiative changes in the atmosphere (Mitchell et al., 1987; Takahashi, 2009).

Some studies on the sensitivity of the hydrologic cycle compared the response to elevated concentrations of carbon dioxide (CO_2) with the sensitivity to absorbed solar radiation. For instance, Andrews et al. (2009) report a hydrologic sensitivity from the Hadley Centre climate model of 1.5 % K^{-1} for a doubling of CO_2, while the simulated sensitivity for a temperature increase due to absorbed solar radiation was 2.4 % K^{-1}. The study by Bala et al. (2008) compared the effects of doubled CO_2 to a geoengineering scheme that reduces solar radiation. They

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also found different hydrologic sensitivities for greenhouse-induced and solar-radiation-induced changes in surface temperature. Govindasamy et al. (2003), Lunt et al. (2008) and Tilmes et al. (2013) report similar effects, namely, that the hydrologic cycle reacts differently to surface temperature differences when the warming results from an enhanced greenhouse effect or enhanced absorption of solar radiation at the surface.

Strictly speaking from a viewpoint of saturation vapor pressure, we would not expect such a difference in hydrologic sensitivity to surface temperature that would depend on whether the surface temperature difference was caused by differences in solar or terrestrial radiation. However, when we focus on the surface energy balance rather than the saturation vapor pressure, it is quite plausible to expect such a difference in sensitivity. After all, the primary cause for surface heating is the absorption of solar radiation, while the exchange of terrestrial radiation as well as the turbulent heat fluxes generally cool the surface. When the surface warms because of changes in the atmospheric greenhouse effect, then the rate of surface heating by absorption of solar radiation remains the same, so that the total rate of cooling by terrestrial radiation and turbulent fluxes remains the same as well. In case the warming is caused by an increase in the absorption of solar radiation, then the overall rate of cooling by terrestrial radiation and turbulent fluxes needs to increase. Hence, we should be able to infer such differences in the hydrologic sensitivity by considering the surface energy balance.

In this paper, we show that hydrologic sensitivities can be predicted by simple surface energy balance considerations in connection with the assumption that convective mass exchange within the atmosphere operates at the thermodynamic limit of maximum power (Kleidon and Renner, 2013). This approach will be briefly summarized in the next section, while the detailed thermodynamic derivations of the maximum power limit, a fuller description of the assumptions and limitations as well as the comparison to observations can be found in the appendix and in Kleidon and Renner (2013). The analytic solution of this model will then be used to derive analytical expressions of the hydrologic sensitivity to surface temperature in Sect. 3 for differences in the atmospheric greenhouse effect as well as for differences in absorption of solar radiation. These sensitivities are compared to the sensitivities obtained from numerical climate model studies. We provide a brief explanation of these differences from an energy balance perspective in Sect. 4, discuss the limitations of our approach, and illustrate one implication of our interpretation for geoengineering approaches to global warming. We close with a brief summary, in which we also point out deficiencies in the concept of radiative forcing that is often used in analyses of global warming and possible extensions of our approach to other aspects of global climatic change.

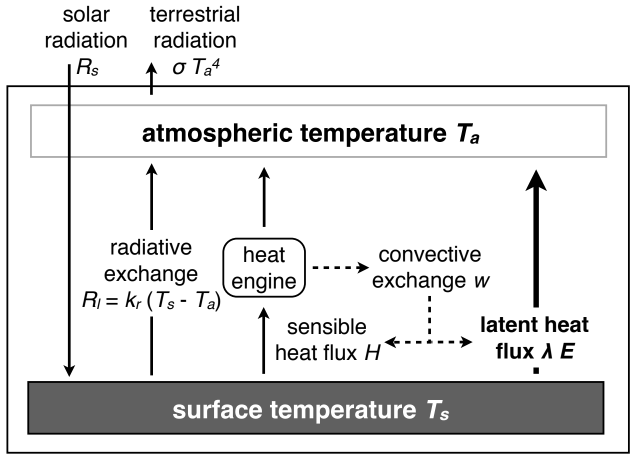

Fig. 1. Schematic illustration of the simple energy balance model that is used to describe the strength of the hydrologic cycle through the rate of surface evaporation, E, with the main variables and fluxes used here. After Kleidon and Renner (2013).

2 Model description

We use the approach of Kleidon and Renner (2013), which describes a thermodynamically consistent global steady state of the surface–atmosphere system in which the hydrologic cycle is represented by evaporation (which balances precipitation, $E = P$, in steady state). The layout of the model as well as the main fluxes is shown in Fig. 1. The model uses the surface and global energy balance to describe the surface temperature, T_s, as well as the (atmospheric) radiative temperature, T_a. The surface is assumed to be an open water surface, all absorption of solar radiation is assumed to take place at the surface, and it is assumed that the atmosphere is opaque for terrestrial radiation so that all radiation emitted to space originates from the atmosphere. Atmospheric dynamics, and particularly the turbulent heat fluxes, are not explicitly considered, but rather inferred from the thermodynamic limit of generating convective motion. The important point to note is that for convective exchange to take place in a steady state, motion needs to be continuously generated against inevitable frictional losses. This kinetic energy is generated out of heating differences akin to a heat engine (as shown in Fig. 1). The conversion of heat to kinetic energy by this heat engine is thermodynamically constrained, and such a thermodynamic limit sets the limit to the turbulent exchange at the surface. A brief derivation of this limit from the laws of thermodynamics is provided in the Appendix. We will refer to this limit and the associated state of the surface energy balance as the state of maximum power, with power being the physical measure of the rate at which work is being performed. We then measure the strength of the hydrologic cycle by the value of E at this maximum power state.

In the model, the surface energy balance is expressed as

$$0 = R_s - R_l - H - \lambda E, \tag{1}$$

where R_s is the absorbed solar radiation at the surface (which is prescribed), R_l the net cooling of the surface by terrestrial radiation, H the sensible heat flux, and λE the latent heat flux. We use simple, but common formulations for these fluxes which are simple enough to obtain analytical results. For the net radiative cooling, we assume a simple linearized form, $R_l = k_r(T_s - T_a)$. Here, k_r is a linearized radiative "conductance" that relates to the strength of the greenhouse effect. The sensible and latent heat fluxes are expressed as turbulent exchange fluxes in the form of $H = c_p \rho w (T_s - T_a)$ and $\lambda E = \lambda \rho w (q_{sat}(T_s) - q_{sat}(T_a))$. The heat capacity of air is $c_p \rho = 1.2 \times 10^3\,\mathrm{J\,m^{-3}\,K^{-1}}$, with a density of about $\rho = 1.2\,\mathrm{kg\,m^{-3}}$; w is a velocity which describes the rate of vertical mass exchange and is determined below from the thermodynamic maximum power limit; $\lambda = 2.5 \times 10^6\,\mathrm{J\,K^{-1}}$ is the latent heat of vaporization; $q_{sat} = 0.622\, e_{sat}/p$ is the saturation specific humidity; e_{sat} is the saturation vapor pressure, and $p = 1013.25$ hPa is surface air pressure. For the saturation vapor pressure, we use the numerical approximation of $e_{sat}(T) = e_0 \cdot e^{a-b/T}$ (Bohren and Albrecht, 1998), with $e_0 = 611$ Pa, $a = 19.83$ and $b = 5417$ K and temperature T in K. The global energy balance yields an expression for the temperature T_a:

$$0 = R_s - \sigma T_a^4, \tag{2}$$

where σ is the Stefan–Boltzmann constant.

The strength of the convective heat fluxes are derived from the assumption that surface exchange is driven mostly by locally generated buoyancy at the surface, and that the power to generate motion by dry convection, $H \cdot (T_s - T_a)/T_s$ is maximized. The Carnot limit has a maximum, because a greater value of H is associated with a smaller value of $T_s - T_a$ due to the constraint imposed by the surface energy balance. This tradeoff between H and $T_s - T_a$ results in a distinct state of maximum power associated with convective exchange at intermediate values for these two terms (see also Appendix). The maximization is achieved by optimizing the vertical exchange velocity w. At maximum power, the optimum value for the vertical exchange velocity, w_{opt}, is given by

$$w_{opt} = \frac{\gamma}{s+\gamma} \frac{R_s}{2 c_p \rho (T_s - T_a)}, \tag{3}$$

where $\gamma = 65\,\mathrm{Pa\,K^{-1}}$ is the psychrometric constant and $s = de_{sat}/dT_s$ is the slope of the saturation vapor pressure curve. This maximum power state results in an energy partitioning at the surface of

$$R_{l,opt} = \frac{R_s}{2} \quad H_{opt} = \frac{\gamma}{s+\gamma}\frac{R_s}{2} \quad \lambda E_{opt} = \frac{s}{s+\gamma}\frac{R_s}{2}. \tag{4}$$

The expression of E_{opt} is nearly identical to the equilibrium evaporation rate (Slayter and McIlroy, 1961; Priestley and Taylor, 1972), a concept that is well established in estimating evaporation rates at the surface, with the additional constraint that the net radiation of the surface at a state of maximum convective power is half of the absorbed solar radiation, R_s.

This partitioning between radiative and turbulent heat fluxes at the surface is associated with a characteristic temperature difference, $T_s - T_a$, which can be used to infer the associated temperatures. The radiative temperature of the atmosphere, T_a, follows directly from the global energy balance, eqn. 2, and is unaffected by the partitioning:

$$T_a = \left(\frac{R_s}{\sigma}\right)^{1/4}. \tag{5}$$

Surface temperature, T_s, at the maximum power state is derived from the expression of net radiative exchange, $R_{l,opt} = k_r(T_s - T_a) = R_s/2$, and is given by

$$T_s = T_a + \frac{R_s}{2 k_r}. \tag{6}$$

In Kleidon and Renner (2013), we showed that this model reproduces the global evaporation rate as well as poleward moisture transport very well. It is important to note, however, that the expression for evaporation given by Eq. (4) represents the maximum evaporative flux that is achieved by locally generated motion near the surface only. In practice, the equilibrium evaporation rate is often corrected by the Priestley-Taylor coefficient (Priestley and Taylor, 1972) of ca. 1.26, which can be understood as the effect of horizontal motion that is generated by horizontal differences in absorption of solar radiation (Kleidon and Renner, 2013). However, as this coefficient simply acts as a multiplier, it does not affect the relative sensitivity of evaporation to changes in the surface energy balance. Also note that evaporation driven by local convection by surface heating can already explain more than 70 % of the strength of the present-day hydrologic cycle (Kleidon and Renner, 2013). We will therefore consider only this locally driven rate of evaporation in the following derivation of the sensitivities.

3 Results

To derive the hydrologic sensitivity to surface temperature, we are interested in the expression $1/E\, dE/dT_s$. We first note that T_s is not the independent variable of our model, because $T_s = T_s(k_r, R_s)$ with the relationship given by Eq. (6), and that solar radiative forcing, R_s, and the greenhouse parameter, k_r, are our independent variables. We can, however, use Eq. (6) to make T_s and k_r our independent variables, and R_s our dependent variable. This sounds a bit backward, but is mathematically sound and allows us to compute $1/E\, dE/dT_s$ analytically.

We now use the expression of E_{opt} in Eq. (4) as the evaporation rate to derive the hydrologic sensitivity. This expression depends on s and R_s, which are both related to our independent variable T_s. The derivative is thus given by

$$\frac{1}{E}\frac{\mathrm{d}E}{\mathrm{d}T_\mathrm{s}} = \frac{1}{E}\frac{\partial E}{\partial s}\frac{\mathrm{d}s}{\mathrm{d}T_\mathrm{s}} + \frac{1}{E}\frac{\partial E}{\partial R_\mathrm{s}}\frac{\partial R_\mathrm{s}}{\partial T_\mathrm{s}}. \tag{7}$$

Since $\partial R_\mathrm{s}/\partial T_\mathrm{s} = (\partial T_\mathrm{s}/\partial R_\mathrm{s})^{-1}$, we can also express this as

$$\frac{1}{E}\frac{\mathrm{d}E}{\mathrm{d}T_\mathrm{s}} = \frac{1}{E}\frac{\partial E}{\partial s}\frac{\mathrm{d}s}{\mathrm{d}T_\mathrm{s}} + \frac{1}{E}\frac{\partial E}{\partial R_\mathrm{s}}\left(\frac{\partial T_\mathrm{s}}{\partial R_\mathrm{s}}\right)^{-1} \tag{8}$$

for which the derivative $\partial T_\mathrm{s}/\partial R_\mathrm{s}$ can be directly calculated from Eqs. (6) and (5). We refer to Eq. (8) as the hydrologic sensitivity.

The hydrologic sensitivity consists of two terms. The first term on the right hand side expresses the dependence of evaporation on s, which depends strongly on surface temperature, while the second term describes the dependence of evaporation on the solar radiative forcing, which also affects surface temperature.

When a difference in surface temperature, ΔT_s, is caused by changes in the atmospheric greenhouse effect (i.e., a different value of k_r), then the solar radiative heating is a constant and $1/E\, \mathrm{d}E/\mathrm{d}T_\mathrm{s} = 1/E\, \partial E/\partial s\, \mathrm{d}s/\mathrm{d}T_\mathrm{s}$. This sensitivity represents only a shift in the partitioning between the sensible and latent heat flux, as the overall magnitude of turbulent fluxes does not change since R_s does not change.

If ΔT_s is caused by a difference in R_s, then $1/E\, \mathrm{d}E/\mathrm{d}T_\mathrm{s}$ consists of two terms, expressing the change of evaporation due to a change in s that is caused by the increase in temperature, but also the overall increase in turbulent fluxes due to the increase in R_s. Hence, we would expect different hydrologic sensitivities to surface temperature, depending on the type of radiative change. Changes in the greenhouse effect affect the first term of the right hand side of Eq. (8) only, while changes in solar radiation affect both terms of the right hand side of Eq. (8) and thus should result in a greater sensitivity.

The first term in Eq. (8) expresses the change of evaporation, E, to surface temperature, T_s, by altering the value of s:

$$\frac{1}{E}\frac{\partial E}{\partial s}\frac{\mathrm{d}s}{\mathrm{d}T_\mathrm{s}} = \frac{\gamma}{s+\gamma}\frac{1}{s}\frac{\mathrm{d}s}{\mathrm{d}T_\mathrm{s}}. \tag{9}$$

We note that this sensitivity does not involve the relative change in saturation vapor pressure $1/e_\mathrm{sat}\, \mathrm{d}e_\mathrm{sat}/\mathrm{d}T_\mathrm{s}$, but rather the relative change in the *slope* in saturation vapor pressure $1/s\, \mathrm{d}s/\mathrm{d}T_\mathrm{s}$. The proportionality to the slope $1/s\, \mathrm{d}s/\mathrm{d}T_\mathrm{s}$, rather than $1/e_\mathrm{sat}\, \mathrm{d}e_\mathrm{sat}/\mathrm{d}T_\mathrm{s}$, is due to the fact that the intensity of the water cycle does not depend on $e_\mathrm{sat}(T_\mathrm{s})$, but rather on the difference of $e_\mathrm{sat}(T_\mathrm{s}) - e_\mathrm{sat}(T_\mathrm{a})$, which is approximated in our model by the slope s. Hence, the sensitivity of the hydrologic cycle does not follow $1/e_\mathrm{sat}\, \mathrm{d}e_\mathrm{sat}/\mathrm{d}T$, but rather $1/s\, \mathrm{d}s/\mathrm{d}T$. The sensitivity is further reduced by a factor $\gamma/(s+\gamma)$, which originates from the energy balance (and maximum power) constraint and ensures that E is not unbound with much higher values for T_s, but converges to an upper limit of $R_\mathrm{s}/2$.

To quantify this first term of the sensitivity for present-day conditions, we use $R_\mathrm{s} = 240\,\mathrm{W\,m^{-2}}$ and derive a value for $k_\mathrm{r} = 3.64\,\mathrm{W\,m^{-2}\,K^{-1}}$ indirectly from the observed global mean temperatures, $T_\mathrm{s} = 288\,\mathrm{K}$ and $T_\mathrm{a} = 255\,\mathrm{K}$ and from Eq. (6) above. With this radiative forcing and values of $\gamma = 65\,\mathrm{Pa\,K^{-1}}$ and $s = 111\,\mathrm{Pa\,K^{-1}}$, we obtain a numerical value of this sensitivity of

$$\frac{1}{E}\frac{\partial E}{\partial s}\frac{\mathrm{d}s}{\mathrm{d}T_\mathrm{s}} \approx 2.2\,\%\,\mathrm{K^{-1}} \tag{10}$$

which matches the mean sensitivity of climate models of $2.2\,\%\,\mathrm{K^{-1}}$ (Allen and Ingram, 2002; Held and Soden, 2006; Li et al., 2013).

The second term of Eq. (8) is due to a difference in absorption of solar radiation, ΔR_s, and is given by

$$\frac{1}{E}\frac{\partial E}{\partial R_\mathrm{s}}\cdot\left(\frac{\partial T_\mathrm{s}}{\partial R_\mathrm{s}}\right)^{-1} = \frac{4 k_\mathrm{r}\sigma^{1/4}}{2\sigma^{1/4}R_\mathrm{s} + k_\mathrm{r}R_\mathrm{s}^{1/4}}. \tag{11}$$

This sensitivity depends only on radiative properties and results in a sensitivity of

$$\frac{1}{E}\frac{\partial E}{\partial R_\mathrm{s}}\cdot\left(\frac{\partial T_\mathrm{s}}{\partial R_\mathrm{s}}\right)^{-1} \approx 1\,\%\,\mathrm{K^{-1}}. \tag{12}$$

This sensitivity is about half the value of the first term when evaluated using present-day conditions, so that the total hydrologic sensitivity to surface temperature change caused by solar radiation is about $3.2\,\%\,\mathrm{K^{-1}}$ and thus exceeds the above sensitivity to changes in the atmospheric greenhouse effect.

These sensitivities are shown graphically in Fig. 2a. The relative proportion of this sensitivity to that caused by changes in the atmospheric greenhouse is consistent with the proportions reported by Bala et al. (2008) and Andrews et al. (2009). In both studies, the authors reported a sensitivity to surface temperature caused by changes in the atmospheric greenhouse of $1.5\,\%\,\mathrm{K^{-1}}$, while the sensitivity to changes in solar radiation was given as $2.4\,\%\,\mathrm{K^{-1}}$. While the magnitude of the sensitivity is smaller compared to the sensitivities calculated here and most other climate models (Allen and Ingram, 2002; Held and Soden, 2006; Li et al., 2013), the sensitivity to temperature differences caused by differences in solar radiation is about 60 % greater than those due to differences in the greenhouse effect, which is similar to the difference that is estimated here.

We will next look at the sensitivities of convective mass exchange that is associated with these differences in hydrologic cycling. The sensible and latent heat flux are accomplished by convective motion, which exchanges the heated and moistened air near the surface with the cooled and dried air of the atmosphere. To evaluate the sensitivity of convective motion to surface temperature, we evaluate the relative difference in w in response to a difference in T_s, for which we use the expression of w_opt as given in Eq. (3):

$$\frac{1}{w}\frac{\mathrm{d}w}{\mathrm{d}T_\mathrm{s}} = \frac{1}{w}\frac{\partial w}{\partial T_\mathrm{s}} + \frac{1}{w}\frac{\partial w}{\partial R_\mathrm{s}}\cdot\left(\frac{\partial T_\mathrm{s}}{\partial R_\mathrm{s}}\right)^{-1}. \tag{13}$$

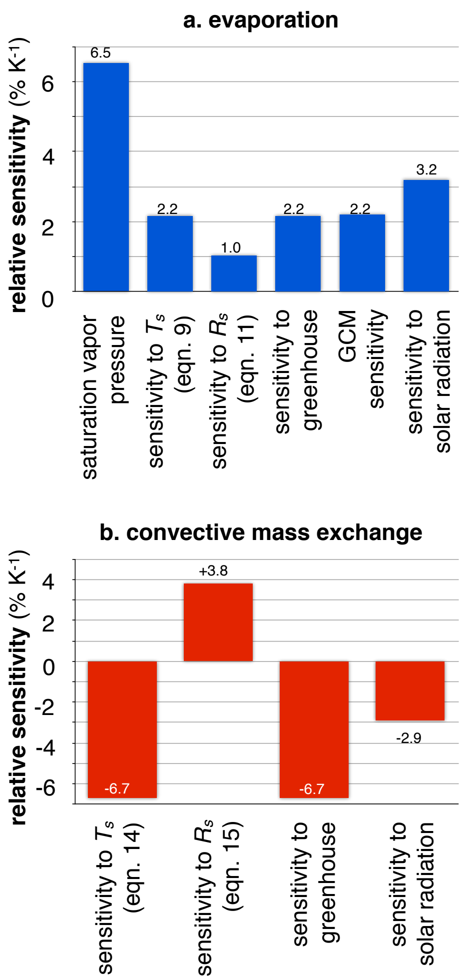

Fig. 2. Sensitivity of **(a)** the hydrologic cycle (evaporation E) and **(b)** convective mass exchange (exchange velocity w) to differences in surface temperature (T_s). Shown are the numerical values for the relative sensitivities as given in the text for present-day conditions. Also included in **(a)** is the sensitivity of saturation vapor pressure, $1/e_{sat}\, de_{sat}/dT_s$, as well as the mean sensitivity to greenhouse differences reported for climate models by Held and Soden (2006) ("GCM sensitivity").

As in the case of evaporation, the sensitivity consists of two terms, with the first term representing the direct response of w to s and T_s. This first term is given by

$$\frac{1}{w}\frac{\partial w}{\partial T_s} = -\frac{s}{s+\gamma}\frac{1}{s}\frac{ds}{dT_s} - \frac{1}{T_s - T_a}. \tag{14}$$

Using the values from above, this yields a sensitivity of $-6.7\,\%\,K^{-1}$. The sensitivity is negative, implying that convective mass exchange is reduced by a stronger greenhouse effect. This sensitivity is consistent with previous interpretations as described by Betts and Ridgway (1989) and Held and Soden (2006), and the estimates of about 4–8 % reported by Boer (1993).

The second term in Eq. (13) describes the indirect effect of differences in solar radiation on w through differences in T_s:

$$\frac{1}{w}\frac{\partial w}{\partial R_s}\cdot\left(\frac{\partial T_s}{\partial R_s}\right)^{-1} = \left(k_r + \frac{k_r^2}{2\sigma^{1/4}R_s^{7/4}}\right)\cdot\frac{4k_r\sigma^{1/4}}{2\sigma^{1/4}R_s + k_r R_s 1/4}. \tag{15}$$

This expression yields a sensitivity of $+3.8\,\%\,K^{-1}$, so that the total sensitivity of convective mass exchange to temperature differences caused by differences in absorption of solar radiation is $-2.9\,\%\,K^{-1}$. This sensitivity is noticeably less than the sensitivity to changes in the atmospheric greenhouse effect (see also Fig. 2b).

In summary, we have shown here that our analytical expressions for the sensitivity of evaporation rate, Eq. (9), can reproduce the reported mean sensitivity of climate models to greenhouse-induced temperature differences. Due to an additional term that relates to changes in absorbed solar radiation (Eq. 11), the hydrologic sensitivity is greater when the temperature increase is due to an increase in the absorption of solar radiation, which is also consistent with what is reported from climate model studies. Associated with these changes in the hydrologic cycle are changes in the intensity of vertical mass exchange, which depend on the type of change in the radiative forcing. Hence, our approach appears to represent a simple yet consistent way to capture the mean aspects of climate change that are reflected in surface temperature differences.

4 Discussion

Before we interpret our results in more detail, we first discuss some of the limitations of our approach and evaluate the extent to which these affect the results. We then interpret our results for the hydrologic sensitivity and relate this interpretation to previous explanations. We close with a brief discussion of one of the implications of our work for the climatic impacts of climate geoengineering by solar radiation management.

4.1 Limitations

Naturally, we have made a number of assumptions in our approach. These assumptions relate to the assumption of (a) the maximum power limit for convective exchange, (b) a steady state of the energy balances, (c) surface exchange being caused by local heating, and (d) a simple treatment of processes in our model.

The use of the maximum power limit provided a means to constrain the convective exchange in our model. If this limit would not have been invoked, the magnitude of the turbulent heat fluxes would be unconstrained, and some form of empirical treatment of these fluxes would be required, typically with an empirically derived value of the drag parameter. The application of a thermodynamic limit to convective exchange avoids this empirical parameter. This limit relates closely to the hypothesis that atmospheric motion maximizes material entropy production, noting that in steady state, power equals dissipation, and entropy production is described by dissipation divided by temperature. This hypothesis was first proposed by Paltridge (1975), and has been quite successful, for instance in predicting heat transport in planetary atmospheres (Lorenz et al., 2001), in deriving an empirical parameter related to turbulence in a general circulation model (Kleidon et al., 2003), and other applications in climate science (e.g., Ozawa et al., 2003). Hence, the assumption that atmospheric motion operates near such a thermodynamic limit, while not widely recognized, has considerable support. For the derivation of our sensitivities, this assumption only matters to the extent that it predicts that net radiation does not change in the case of greenhouse-induced warming. In other words, the derivation of the hydrologic sensitivity to greenhouse-induced warming (Eq. 9) could have been done with the assumption that net radiation does not change. Likewise, the hydrologic sensitivity to solar-induced changes of surface temperature (Eqs. 9 and 11) could have been derived from the assumption that the ratio between radiative and turbulent cooling remains fixed. Both of these assumptions can then be justified and explained by the maximum power limit.

We also assumed that the energy balances of the surface and the atmosphere are in a steady state. This assumption ignores the temporal variations on diurnal and seasonal timescales, which result in the dynamics of boundary layer growth and changes in heat storage. These aspects are most relevant on land, while over the ocean, these aspects are likely to play a minor role due to the large heat capacity of water. Since the sensitivity of the hydrologic cycle is dominated by the oceans, it would thus seem reasonable to neglect these variations.

Another assumption that we have made is that the turbulent exchange at the surface results only from local surface heating. This assumption neglects the fact that the large-scale circulation adds extra turbulence to the surface, thus generating more turbulence at the surface than what would be expected by local heating alone. This extra contribution would shift the partitioning in the surface energy balance towards turbulent heat fluxes. In the framework of the equilibrium evaporation rate, this shift can be interpreted by the Priestley-Taylor coefficient. We incorporated this effect in Kleidon and Renner (2013) by introducing a factor into the formulation of the sensible and latent heat flux, but we did not use this factor here. The reason for omitting this factor is that as long as this factor is independent of T_s, the *relative* sensitivities that we derived here are not affected as this factor would cancel out. Hence, this large-scale contribution to turbulent exchange is unlikely to result in substantially different sensitivities.

In addition, we implemented processes in our approach in a simplified way. We assumed that all absorption of solar radiation takes place at the surface, while observations (e.g., Stephens et al., 2012) state that it is only about $165\,\mathrm{W\,m^{-2}}$ rather than $240\,\mathrm{W\,m^{-2}}$ of solar radiation which is absorbed at the surface. We used this simplification to keep the model as simple as possible (otherwise, we would need to account for atmospheric absorption in the expression for T_a). For the hydrologic sensitivity, this simplification plays a minor role because the sensitivity is formulated in relative terms, which is independent of R_s (at least the first term in Eq. 8). In addition, we assumed that the atmosphere is mostly opaque for terrestrial radiation. This assumption does not hold for all regions. Particularly in dry and cold regions, the atmosphere is more transparent to terrestrial radiation. This would affect our model in which it is assumed that all terrestrial radiation to space originates from the atmosphere (cf. Eq. 2).

Overall, while we made several assumptions and simplifications in obtaining our results, it would seem that our results are rather robust. These assumptions may need to be revisited and refined when using this approach at different scales or conditions. For instance, when this approach is applied to land, then one would need to account for the additional constraint of water limitation. When it is applied to the diurnal cycle, one would clearly need to account for changes in heat storage. These factors can, of course, be included in an extension of the approach, but they should nevertheless not affect our results at the global scale in the climatic mean.

4.2 Interpretation

The interpretation of our results is relatively straightforward and can be attributed entirely to changes in the surface energy balance. This focus on changes in the surface energy balance is plausible, because after all, convective mass exchange, the associated transport of sensible and latent heat, and hence hydrologic cycling is caused by surface heating. It is important to note that the actual heating of the surface is solely due to the absorption of solar radiation, R_s, while terrestrial radiation, R_l, cools the surface. In the following, we explain these changes and illustrate these for an example of a surface warming of $\Delta T_s = 2\,\mathrm{K}$, which is shown in Fig. 3.

When the surface warming is entirely caused by an increase of the atmospheric greenhouse effect, R_s is effectively unchanged, but the cooling of the surface by terrestrial radiation is less efficient. In our model, this reduced cooling efficiency is reflected in a lower value of k_r. This lower value of k_r, however, does not affect the partitioning of absorbed solar radiation into radiative and turbulent cooling, R_l and $H + \lambda E$, at the maximum power state. This is noticeable in Eq. (4), since the partitioning does not depend on the value of k_r. Hence, R_s and R_l do not change (cf. Fig. 3, blue bars). However, because k_r is reduced, it requires a greater temperature difference, $T_s - T_a$, to accomplish the same radiative cooling flux, R_l. Since T_a is fixed by the global energy balance and is independent of k_r, this can only be accomplished by an increase in T_s. This surface warming is then associated with a different partitioning between sensible and latent heat, because the slope of the saturation vapor pressure curve, s,

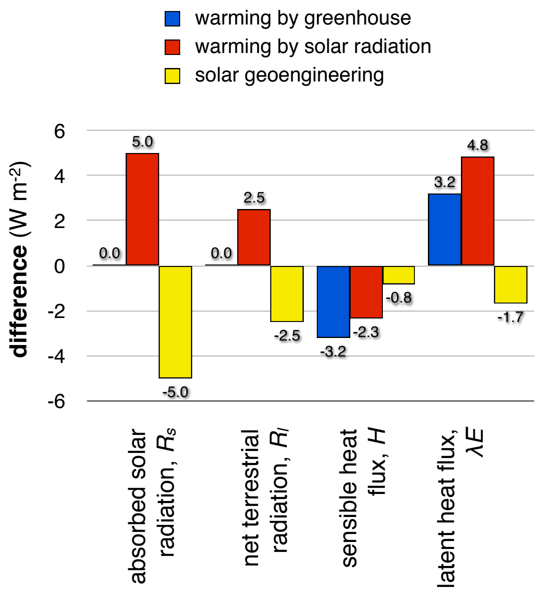

Fig. 3. Estimates of the changes in the surface energy balance components due to a warming of $\Delta T_s = 2$ K caused by an increase in the atmospheric greenhouse effect (blue, "warming by greenhouse"), by an increase in absorbed solar radiation (red, "warming by solar absorption"), and when a greenhouse warming of 2 K is compensated for by a reduction of solar radiation by some geoengineering management (yellow, "solar geoengineering"). The numbers are obtained using values of $R_s = 240\,\mathrm{W\,m^{-2}}$ and $k_r = 3.64\,\mathrm{W\,m^{-2}\,K^{-1}}$ for the present-day climate, a reduction of k_r to $k_r = 3.44\,\mathrm{W\,m^{-2}\,K^{-1}}$ to get a surface warming of 2 K by changes in the greenhouse effect, an increase of $\Delta R_s = 5\,\mathrm{W\,m^{-2}}$ to get a surface warming of 2 K by changes in absorption of solar radiation, and a combined change of $k_r = 3.44\,\mathrm{W\,m^{-2}\,K^{-1}}$ and $\Delta R_s = -5\,\mathrm{W\,m^{-2}}$ to implement the solar radiation management effects.

has a greater value at a warmer temperature, resulting in a greater proportion, $s/(s+\gamma)$, of the turbulent cooling being represented by the latent heat flux, thus resulting in a stronger hydrologic cycle (Fig. 2a). In the example shown in Fig. 3 the consequence of the warming is reflected merely in the shift from the sensible heat flux to the latent heat flux, but the magnitude of both does not change.

Since the difference $T_s - T_a$ is enhanced, the turbulent heat fluxes are accomplished by less convective mass exchange, which results in the negative sensitivity $1/w\,\partial w/\partial T_s$ (Fig. 2b). Since both sensitivities deal with the intensity of convective transport and its partitioning into sensible and latent heat, the sensitivities are expressed only in terms of related properties (s, γ, $T_s - T_a$, cf. Eqs. 9 and 14), but do not depend explicitly on radiative properties of the system (R_s, k_r). This interpretation is consistent with the general understanding of the greenhouse effect, but it emphasizes that the atmospheric greenhouse effect acts to reduce the efficiency by which the surface cools through the emission of terrestrial radiation.

The situation is different when the surface warms due to enhanced absorption of solar radiation (Fig. 3, red bars). In this case, the surface is heated more strongly (R_s is increased), so the rate of cooling, $R_l + H + \lambda E$, is increased as well. Apart from the difference in surface temperature and the associated differences in the partitioning between the sensible and latent heat flux, the overall magnitude of the turbulent fluxes is altered as well. Hence, the sensitivity is greater than in the case of greenhouse warming, which is noticeable in our example by the increase in sensible and latent heat (compare red vs. blue bars in Fig. 3). The additional contribution by the overall increase in turbulent fluxes depends on R_s and on the temperature difference, which depends on R_s and k_r. Consequently, the second term in the sensitivities depends explicitly on the radiative properties of the system (R_s, k_r, cf. Eqs. 11 and 15). This enhancement of the turbulent fluxes favors greater convective mass exchange, so that the sensitivity of convective mass exchange is reduced compared to differences caused by a stronger greenhouse effect.

Our interpretation is quite different from the common explanation for the hydrologic sensitivity (e.g., Mitchell et al., 1987; Allen and Ingram, 2002; Takahashi, 2009; Allan et al., 2013). The common explanation starts by considering the atmospheric energy balance. Surface warming results in a perturbation of this energy balance. It accounts for the extra release of latent heat, $\lambda\Delta P$, the change in radiative cooling of the atmosphere to space, ΔR_{toa}, the change in radiative fluxes from the surface, ΔR_l, and a change in the sensible heat flux, ΔH:

$$\lambda\Delta P = \Delta R_{toa} - \Delta R_l - \Delta H. \tag{16}$$

The common explanation for the lower sensitivity of precipitation to surface warming compared to the sensitivity of the saturation water pressure argues that the additional release of latent heat, $\lambda\Delta P$, is constrained by the ability to radiate away the additional heat by the term $\Delta R_{toa} - \Delta R_l$. The term ΔH in these considerations is commonly neglected because H is quite a bit smaller than the latent heat flux.

This energy balance is, of course, indirectly also obeyed in our model even though we do not explicitly consider it. First, we consider a steady state, so that $\Delta R_s = \Delta R_{toa}$, $\lambda\Delta P = \lambda\Delta E$, and, $1/P \cdot \mathrm{d}P/\mathrm{d}T_s = 1/E \cdot \mathrm{d}E/\mathrm{d}T_s$.

We first consider the case of greenhouse-induced surface warming. In this case, changes in the greenhouse effect do not change the radiative temperature of the atmosphere (which is entirely determined by R_s, Eq. 5), hence, $\Delta R_{toa} = 0$. The term R_l does not change either, because the surface heating by solar radiation did not change ($\Delta R_s = 0$) and the maximum power constraint results in an equal partitioning among R_l and $H + \lambda E$, no matter how strong the greenhouse effect is. Hence, the overall changes in the atmospheric energy balance reduce to

$$\lambda\Delta P = -\Delta H. \tag{17}$$

This implies that the weak, 2.2 % K^{-1} increase in the strength of the hydrologic cycle can simply be explained by the reduction of the sensible heat flux at the surface. This interpretation is identical to what we found for the changes in the surface energy balance: a greenhouse-induced surface warming merely affects the partitioning between sensible and latent heat, but does not affect the magnitude of the turbulent heat fluxes (see also example in Fig. 3, blue bars). This explanation is different to the common explanation, which neglects changes in the sensible heat flux. In our explanation, the hydrologic sensitivity due to greenhouse-induced surface warming is entirely due to the reduction of H.

The changes in the atmospheric energy balance are different if the surface temperature change was caused by changes in solar radiation. If absorbed solar radiation increases by ΔR_s, then the global energy balance requires that $\Delta R_{toa} = \Delta R_s$, so that the radiative temperature T_a must increase. The partitioning of energy at the surface changes as well. At a state of maximum power, the additional heating of ΔR_s results in an equal increase in radiative and turbulent fluxes of $\Delta R_l = \Delta R_s/2$, and of $\Delta(H + \lambda E) = \Delta R_s/2$. In addition, the increase in surface temperature alters the partitioning between H and λE. Hence, in this case, all four terms are going to change in Eq. (16), which is quite a different change than the greenhouse-induced warming.

Overall, our explanation is quite different to the common explanation of the hydrologic sensitivity. Yet, our explanation is simple, physically based, consistent with the atmospheric energy balance, and predicts the right value of the sensitivities.

4.3 Implications

An important implication of our interpretation of the hydrologic sensitivity is that the forcing of the surface cannot be simply lumped into a single, radiative forcing concept. The notion of a "radiative forcing" combines the changes in solar and terrestrial radiation into one variable. However, as these sensitivities show, solar radiation plays a very different role than terrestrial radiation. The strength of hydrologic cycling as well as convective mass exchange react quite differently if the surface is warmed due to stronger heating by solar radiation or due to a weaker cooling by a stronger greenhouse effect. An immediate consequence of this notion is that climate geoengineering cannot simply be used to undo global warming (see also Bala et al., 2008 and Tilmes et al., 2013). This can be illustrated using the sensitivities given above.

We consider the case of surface warming of 2 K caused by an enhanced greenhouse effect, as before, except that we look at the relative sensitivities rather than the absolute changes in the surface energy balance. Since this increase of surface temperature is caused by the greenhouse effect, the hydrologic cycle would be strengthened by the sensitivity $1/E\,\partial E/\partial s\,\mathrm{d}s/\mathrm{d}T_s$ (Eq. 9 and blue bars in Fig. 3). This sensitivity has a value of 2.2 % K^{-1}, so that E would increase by 4.4 % with a warming of 2 K. This increase is shown by the arrow in Fig. 4a from the original climatic state "A" to the state in which the surface is heated by 2 K (point "B"). The convective mass exchange would be reduced following the sensitivity $1/w\,\partial w/\partial T_s$ (Eq. 14), which has a value of −6.7 % K^{-1}. With a 2 K warming, the convective mass exchange would be reduced by 13.4 % (Fig. 4b).

If this surface warming is reduced by a reduction of absorbed solar radiation (cf. yellow bars in Fig. 3), as proposed

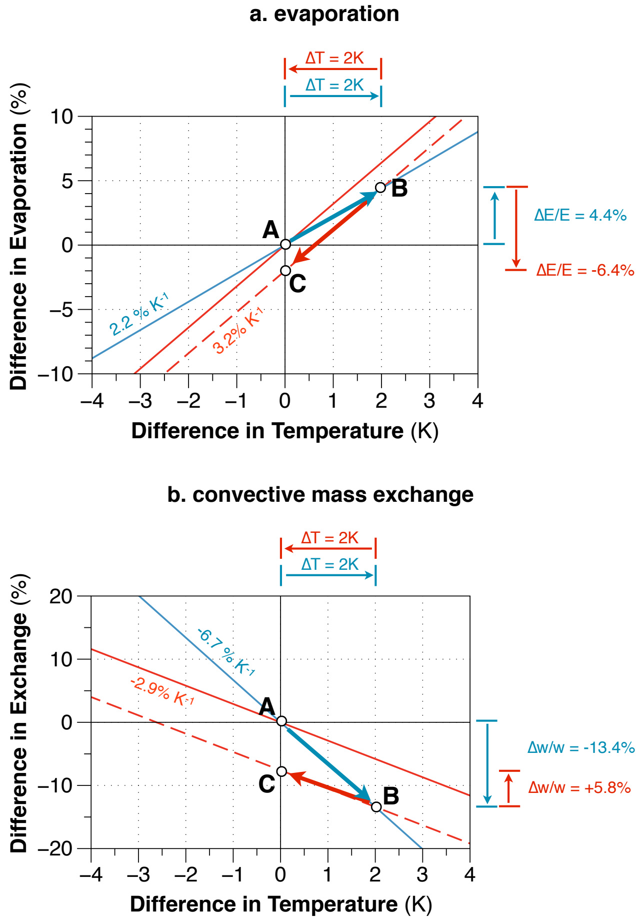

Fig. 4. Illustration of the contrary effects of greenhouse vs. solar-induced changes on **(a)** evaporation, E, and **(b)** convective mass exchange, w. The sensitivity of evaporation to a surface warming of $\Delta T_s = 2$ K by an elevated greenhouse effect results in an increase of 2.2 % K^{-1}, resulting in a change from point A to B (blue line). A geoengineering response aimed at compensating this increase in surface temperature by reducing solar radiation would change evaporation at a rate of 3.2 % K^{-1}, resulting in a change from point B to C (red line). Likewise, the surface warming would decrease the vertical exchange velocity, w, by −6.7 % K^{-1} (point A to B), while the geoengineering response would increase it at a rate of 2.9 % K^{-1} (point B to C). Hence, while the geoengineering response may undo differences in surface temperature, it cannot compensate changes in the hydrologic cycle and vertical mass exchange at the same time.

by some geoengineering schemes, then the value of R_s would change and we need to consider both terms in the sensitivities of evaporation (Eq. 8) and convective mass exchange (Eq. 13). The hydrologic cycle would be reduced not at a rate of 2.2 % K^{-1} as in the case above, but rather at the combined value of 3.2 % K^{-1}, which includes the additional sensitivity given by Eq. (11). Hence, with a cooling of 2 K that would be necessary to undo the surface warming, the strength of the hydrologic cycle would be reduced in total by 6.4 %. This is shown in Fig. 4a by the arrow from point B to C. Overall, the warming would be undone, but the strength of the hydrologic cycle at this state of geoengineering (point C) would be weaker by 2 % compared to the original state (point A). Likewise, the sensitivities of convective mass exchange do not compensate either. The cooling of 2 K by the reduction of absorbed solar radiation follows the weaker sensitivity of -2.9 % K^{-1}, so that the convective mass exchange would increase by only 5.8 % (see arrow from point B to C in Fig. 4b). Hence, overall, the greenhouse warming by 2 K and the geoengineering cooling by 2 K would weaken convective mass exchange by 6.8 % (compare point A and C in Fig. 4b, which is also seen in the yellow bars in Fig. 3).

Hence, such intervention by geoengineering may undo surface warming, but it cannot undo differences in hydrologic cycling and convective mass exchange at the same time. What this tells us is that it is important to consider the different roles of solar and terrestrial radiation separately in future studies on the strength of the hydrologic cycle and global climatic change (see also Jones et al., 2013).

5 Summary and conclusions

In this study we showed that the sensitivity of the hydrologic cycle to surface temperature can be quantified using a simplified surface energy balance and the assumption that convective exchange near the surface takes place at the limit of maximum power. This model yields analytical expressions for the hydrologic sensitivity and shows that it does not scale with the saturation vapor pressure, but rather with its slope. The hydrologic sensitivity scales with the slope of the saturation vapor pressure curve because hydrologic cycling relates to the differences in saturation vapor pressure between the temperatures at which evaporation and condensation takes place. This difference is approximated by the slope. The actual sensitivity is then further reduced by a factor $\gamma/(s+\gamma)$, which originates from the surface energy balance constraint. Our analytical expressions also show that surface warming caused by increases in absorbed solar radiation result in a greater sensitivity of the hydrologic cycle than warming caused by an increased greenhouse effect. This greater sensitivity for warming due to solar radiation is simply explained by the requirement for a greater total cooling rate by radiative and turbulent fluxes. Even though our approach is highly simplistic and omits many aspects, the analytical expressions yield sensitivities that are consistent with those found in rather complex climate models. We conclude that the hydrologic sensitivity to surface warming can be explained in simple, physical considerations of the surface energy balance.

An important implication of our results is that geoengineering approaches to reduce global warming are unlikely to succeed in restoring the original climatic conditions. Because of the difference in hydrologic sensitivities to solar vs. greenhouse induced surface warming, the changes in hydrologic cycling and convective mass exchange do not compensate even if surface temperature changes are compensated by solar radiation management. This example emphasizes the different roles that solar and terrestrial radiation play in the surface energy balance and challenges the frequently used radiative forcing concept, which lumps these two components together. It would seem insightful to extend our study in the future to other aspects of global climatic change, in which the different roles in solar and terrestrial radiation are explicitly considered in a thermodynamically consistent way.

Appendix A

Thermodynamic limits

The first and second law of thermodynamics set a fundamental direction as well as limits to energy conversions within any physical system. We apply it here to derive the limit to how much kinetic energy can be derived from the differential radiative heating between the surface and the atmosphere. The following derivation summarizes Kleidon and Renner (2013).

To derive the limit, we consider a heat engine as marked in Fig. 1 that is driven by the sensible heat flux, H. In the steady-state setup used here, the first law of thermodynamics requires that the turbulent heat fluxes in and out of the engine, H and H_{out}, are balanced by the generation of kinetic energy, G:

$$0 = H - H_{out} - G. \tag{A1}$$

The second law of thermodynamics requires that the entropy of the system does not decrease during the process of generating kinetic energy. This requirement is expressed by the entropy fluxes associated with the heat fluxes H and H_{out} that enter and leave the heat engine at the temperatures of the surface and the atmosphere:

$$\frac{H}{T_s} - \frac{H_{out}}{T_a} \geq 0. \tag{A2}$$

In the best case, the entropy balance equals zero, which then allows us to express the flux H_{out} as a function of H, T_s, and T_a:

$$H_{out} = H \cdot \frac{T_a}{T_s}. \tag{A3}$$

When combined with Eq. (A2), this yields the well-known Carnot limit of the power generated by a heat engine:

$$G = H \cdot \frac{T_s - T_a}{T_s}. \tag{A4}$$

In steady state, this generated power is dissipated by friction, so that $G = D$.

While the Carnot limit provides a constraint on how much power can be generated, it does not directly provide a limit on the value of H. Such a limit is obtained when we consider that the temperature difference, $T_s - T_a$, is not independent of H, but rather constrained by the energy balance (Eq. 1). This temperature difference can be expressed using the expression for the net radiative exchange flux, R_l, from above as

$$T_s - T_a = \frac{R_s - H - \lambda E}{k_r}. \tag{A5}$$

Hence, the surface energy balance demands that the temperature difference, $T_s - T_a$, decreases with an increasing value of H. When combined, this yields an expression for the Carnot limit of

$$G = H \cdot \frac{R_s - H - \lambda E}{k_r}. \tag{A6}$$

Due to the contrasting effects of H on G in the two terms of the right hand side, the Carnot limit has a maximum value at intermediate values of H.

We obtain this maximum in the Carnot limit when we first use the formulation of the sensible and latent heat flux from above in terms of the vertical velocity, w to express the temperature difference:

$$T_s - T_a = \frac{R_s}{k_r + c_p \rho w (1 + s/\gamma)}, \tag{A7}$$

and then combine this expression with the formulation of the sensible heat flux, which yields:

$$G = \frac{c_p \rho w}{T_s \left(k_r + c_p \rho w (1 + s/\gamma)\right)^2} \cdot R_s^2. \tag{A8}$$

This equation has a maximum value with respect to w, which can be derived analytically when we neglect that T_s in the denominator depends on H as well. The maximization is achieved by $\partial G/\partial w = 0$. This yields the expression of the optimum exchange velocity (Eq. 3) and results in an optimal energy partitioning as given by Eq. (4).

This maximum power limit describes the upper thermodynamic limit by which convective motion can be generated to sustain the sensible heat flux out of local radiative heating by absorption of solar radiation at the surface. It does not necessarily imply that this limit is achieved. This would rather formulate a hypothesis, namely, that the surface-energy partitioning would operate near this maximum power limit. This hypothesis is very closely related to the proposed principle of Maximum Entropy Production (MEP, Ozawa et al., 2003; Kleidon et al., 2010), noting that in steady state, $P = D$, and entropy production is described by D/T. A more complete discussion on this relationship is given in Kleidon and Renner (2013). In this paper, we assume that natural processes operate at this thermodynamic limit to derive the analytical expressions.

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