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A simple explanation for the sensitivity of the hydrologic cycle to global climate change

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Abstract

The global hydrologic cycle is likely to increase its strength with global warming. Climate models generally predict an increase in strength of 2.2 % K^{-1}, which is much weaker than what would be expected from the increase in saturation vapor pressure of 6.5 % K^{-1}. Furthermore, it has been reported that the sensitivity of the hydrologic cycle to surface temperature differences caused by solar radiation is about 50 % greater than by an equivalent difference induced by the greenhouse effect. Here we show that these sensitivities can be derived analytically from an extremely simple surface energy balance model that is constrained by the assumption that vertical convective transport within the atmosphere operates at maximum power. Using current climatic mean conditions, this model predicts a sensitivity of the hydrologic cycle of 2.2 % K^{-1} to surface temperature induced by differences in the greenhouse effect, and a sensitivity of 3.2 % K^{-1} for differences caused by absorbed solar radiation. These sensitivities can be explained by considering the changes in the surface energy balance in which the heating by solar radiation is partitioned equally into radiative and turbulent cooling at a state of maximum power of convective exchange. This explanation emphasizes the different roles that solar and terrestrial radiation play in the surface energy balance and hydrologic cycling that cannot be lumped together into a radiative forcing concept. We illustrate one implication of this explanation for the case of geoengineering, which aims to undo surface temperature differences by solar radiation management, but will nevertheless result in substantial differences in hydrologic cycling due to the difference in sensitivities. We conclude that the overall sensitivity of the hydrologic cycle to surface temperature can be understood and predicted by very simple physical considerations.

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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 surfaces warm, 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).

Some studies of 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 of 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 and they found similar hydrologic sensitivities to surface temperature. Govindasamy et al. (2003) and Lunt et al. (2008) 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 different hydrologic sensitivity to surface temperature that would depend on whether the surface temperature difference was caused by differences in solar or terrestrial radiation.

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While these changes are often explained in terms of radiative changes in the atmosphere (Mitchell et al., 1987; Takahashi, 2009), we show here that these 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 a state 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 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 and illustrate one implication of these results in the context of 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.

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. Atmospheric dynamics are not explicitly considered, but rather constrained by the entropy exchanges taking place at the boundaries at temperatures T_s and T_a. The energy partitioning at the surface is then determined

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from the assumption that the generation of convective motion within the atmosphere is maximized, corresponding to a state of maximum power.

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), $R_l = k_r(T_s - T_a)$ is the net cooling of the surface by terrestrial radiation (with a linearized radiative conductance, k_r, that relates to the strength of the greenhouse effect), $H = c_p\rho w(T_s - T_a)$ is the sensible heat flux, and $\lambda E = \lambda\rho w(q_{sat}(T_s) - q_{sat}(T_a))$ is the latent heat flux, where $c_p\rho = 1.2\,\mathrm{J\,m^{-3}\,K^{-1}}$ is the heat capacity of air with a density of about $\rho = 1.2\,\mathrm{kg\,m^{-3}}$, w is a vertical effective exchange velocity that is determined by maximization as described below, $\lambda = 2.5 \times 10^6\,\mathrm{J\,K^{-1}}$ 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 $p = 1013.25\,\mathrm{hPa}$ 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 $e_0 = 611\,\mathrm{Pa}$, $a = 19.83$ and $b = 5417\,\mathrm{K}$ 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. 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

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$$w_{\mathrm{opt}} = \frac{\gamma}{s+\gamma} \frac{R_\mathrm{s}}{2c_p\rho(T_\mathrm{s} - T_\mathrm{a})} \quad (3)$$

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

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

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, $T_\mathrm{s} - T_\mathrm{a}$, which can be used to infer the associated temperatures. The radiative temperature of the atmosphere, T_a, follows directly from the global energy balance, Eq. (2), and is unaffected by the partitioning:

$$T_\mathrm{a} = \left(\frac{R_\mathrm{s}}{\sigma}\right)^{1/4}. \quad (5)$$

Surface temperature, T_s, at the maximum power state is derived from the expression of net radiative exchange, $R_{\mathrm{l,\,opt}} = k_\mathrm{r}(T_\mathrm{s} - T_\mathrm{a}) = R_\mathrm{s}/2$, and is given by

$$T_\mathrm{s} = T_\mathrm{a} + \frac{R_\mathrm{s}}{2k_\mathrm{r}}. \quad (6)$$

In Kleidon and Renner (2013), we showed that this model reproduces key characteristics of the hydrologic cycle very well. It is important to note, however, that the expression for evaporation given by Eq. (4) represents the maximum evaporative flux that

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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 about 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 and discussion

The hydrologic sensitivity is expressed by the derivative of evaporation, E, to surface temperature T_s:

$$\frac{1}{E}\frac{\mathrm{d}E}{\mathrm{d}T_s} = \frac{1}{E}\frac{\partial E}{\partial T_s} + \frac{1}{E}\frac{\partial E}{\partial R_s} \cdot \left(\frac{\partial T_s}{\partial R_s}\right)^{-1}. \tag{7}$$

This sensitivity consists of two terms. The first term on the right hand side expresses the direct dependence of evaporation on surface temperature, while the second term describes the dependence of evaporation on the solar radiative forcing, which also affects surface temperature. Note that the independent variable in our setup is R_s rather than T_s, which is why we express the derivative in the second term as $(\partial T_s/\partial R_s)^{-1}$ rather than $\partial R_s/\partial T_s$. 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 unaffected, so that $\partial E/\partial R_s = 0$ and $(1/E) \cdot \mathrm{d}E/\mathrm{d}T_s = (1/E) \cdot \partial E/\partial T_s$. This sensitivity represents only a shift in the partitioning between the sensible and latent heat flux, as the overall partitioning at maximum power is constrained by the

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unaffected solar radiative heating, R_s. If ΔT_s is caused by a difference in solar radiative heating, then $(1/E) \cdot \mathrm{d}E/\mathrm{d}T_s$ consists of two terms, expressing the direct change of evaporation due to ΔT_s, but also the indirect effect due to enhanced absorption of solar radiation, ΔR_s. The latter term accounts for the fact that when R_s changes, the turbulent heat fluxes change in magnitude as well, and not just their partitioning. Hence, we would expect different hydrologic sensitivities to surface temperature, depending on the type of radiative change.

The first term in Eq. (7) expresses the direct change of evaporation, E, to surface temperature, T_s, and using the expression for E_{opt} from above, we obtain:

$$\frac{1}{E}\frac{\partial E}{\partial T_s} = \frac{\gamma}{s+\gamma}\frac{1}{s}\frac{\mathrm{d}s}{\mathrm{d}T_s}. \tag{8}$$

We note that this sensitivity does not involve the relative change in saturation vapor pressure $(1/e_{sat})\mathrm{d}e_{sat}/\mathrm{d}T_s$, but rather the relative change in the *slope* in saturation vapor pressure $(1/s)\mathrm{d}s/\mathrm{d}T_s$, reduced by $\gamma/(s+\gamma)$, which originates from the constraint of the surface energy balance and includes characteristics of the sensible heat flux through the value of γ. To quantify this expression for present-day conditions, we use $R_s = 240\,\mathrm{W\,m^{-2}}$ (assuming that all radiation is absorbed at the surface for simplicity) and derive a value for $k_r = 3.6\,\mathrm{W\,m^{-2}\,K^{-1}}$ indirectly from the observed global mean temperatures, $T_s = 288\,\mathrm{K}$ and $T_a = 255\,\mathrm{K}$ and from Eq. (4) 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

$$\frac{1}{E}\frac{\partial E}{\partial T_s} \approx 2.2\,\%\,\mathrm{K^{-1}} \tag{9}$$

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

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The second term of Eq. (7) is due to a difference in absorption of solar radiation, ΔR_s, and is given by:

$$\frac{1}{E}\frac{\partial E}{\partial R_s} \cdot \left(\frac{\partial T_s}{\partial R_s}\right)^{-1} = \frac{4k_r\sigma^{1/4}}{2\sigma^{1/4}R_s + k_r R_s^{1/4}}. \qquad (10)$$

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

$$\frac{1}{E}\frac{\partial E}{\partial R_s} \cdot \left(\frac{\partial T_s}{\partial R_s}\right)^{-1} \approx 1\,\%\,K^{-1}. \qquad (11)$$

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 % 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 % K^{-1}, while the sensitivity to changes in solar radiation was given as 2.4 % 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), 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.

Before we interpret these sensitivites, we will first 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

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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 by Eq. (3):

$$\frac{1}{w}\frac{\mathrm{d}w}{\mathrm{d}T_s} = \frac{\partial w}{\partial T_s} + \frac{\partial w}{\partial R_s} \cdot \left(\frac{\partial T_s}{\partial R_s}\right)^{-1}. \tag{12}$$

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

$$\frac{1}{w}\frac{\partial w}{\partial T_s} = -\frac{s}{s+\gamma}\frac{1}{s}\frac{\partial s}{\partial T_s} - \frac{1}{T_s - T_a}. \tag{13}$$

Using the values from above, this yields a sensitivity of $-6.7\,\%\,\mathrm{K}^{-1}$. 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. (12) describes the indirect effect of differences in solar radiation on w through differences in T_s:

$$\frac{1}{w}\frac{\partial w}{\partial R_s} \cdot \left(\frac{\partial T_s}{\partial R_s}\right)^{-1} = \left(k_r + \frac{k_r^2}{2\sigma^{1/4} R_s^{7/4}}\right) \cdot \frac{4 k_r \sigma^{1/4}}{2\sigma^{1/4} R_s + k_r R_s^{1/4}}. \tag{14}$$

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

The interpretation of these sensitivities is simple and straightforward and they can be attributed entirely to changes in the surface energy balance. This is plausible, because after all, convective mass exchange, the associated transport of sensible and latent

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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. When R_s is unchanged but the atmospheric greenhouse effect is increased, the cooling by terrestrial radiation is less efficient. In the model used here, this is reflected in a lower value of k_r. The maximum power state of convective motion partitions the solar heating, R_s, equally into net radiative cooling, R_l, and turbulent cooling, $H + \lambda E$. This partitioning depends only on R_s, but not on k_r. Hence, since k_r is reduced, it requires a greater temperature difference, $T_s - T_a$, to accomplish the same radiative cooling flux. Since T_a is fixed by the global energy balance and is independent of k_r, this results in 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, has a greater value at a warmer temperature, resulting in a greater proportion $s/(s+\gamma)$ of the turbulent flux being represented by the latent heat flux, thus resulting in a stronger hydrologic cycle (Fig. 2a). 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 $(1/w)\mathrm{d}w/\mathrm{d}T_s$ (Fig. 2b). Since both sensitivities deal only 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. 8 and 13), 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. 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, the overall magnitude of the turbulent fluxes is altered, which depends on R_s and on the temperature difference, which depends on R_s and k_r. Hence, the second term in the sensitivities depends explicitly only on the radiative properties of the system

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(R_s, k_r, cf. Eqs. 10 and 14). Overall, the hydrologic cycle is strengthened more than by the changes in the greenhouse because in addition to the surface warming, the turbulent fluxes are increased as well. 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.

An important implication of this difference in sensitivities 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 cannot simply be geoengineered to undo global warming (see also Bala et al., 2008). This can be illustrated using the sensitivities given above. Imagine that due to global warming, the surface heats by 2 K. This warming would result in a strengthening of the hydrologic cycle of 4.4 %, and a reduction of convective mass exchange by 13.4 %. The surface warming could be reduced by some geoengineering scheme that reduces solar radiation, which would require a cooling of 2 K to counteract the warming. Then, the hydrologic cycle would be reduced by 6.4 % and the convective mass exchange would be enhanced by 5.8 %. Overall, the combination of the enhanced greenhouse effect and the solar radiation management would result in no surface temperature difference, but the hydrologic cycle would weaken by 4.4 − 6.4 % = −2 % and the convective mass exchange would be reduced by −13.4 % + 5.8 % = −6.8 %. Hence, such intervention by geoengineering may undo surface warming, but it cannot undo differences in hydrologic cycling and other critical processes within the Earth system 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).

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4 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 a state of maximum power. This model yields analytical expressions for the sensitivity and shows that it does not scale with the saturation vapor pressure, but rather with its slope, reduced by a factor that results from the surface energy balance constraint. 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. These sensitivities can be explained in simple, physical terms.

That the hydrologic sensitivity of complex climate models can be explained in such simple physical terms is surely reassuring. At the same time, our explanation also points out deficits in the concept of radiative forcing that is often used in studies of climatic change, because solar radiation and terrestrial radiation play rather different roles in the surface energy balance.

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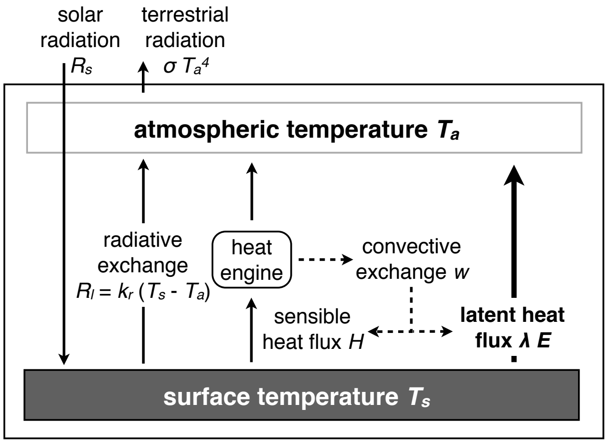

Fig. 1. Schematic illustration of the simple energy balance model that is used to describe the strength of the hydrologic cycle through the rate of surface evaporation, E, with the main variables and fluxes used here. After Kleidon and Renner (2013).

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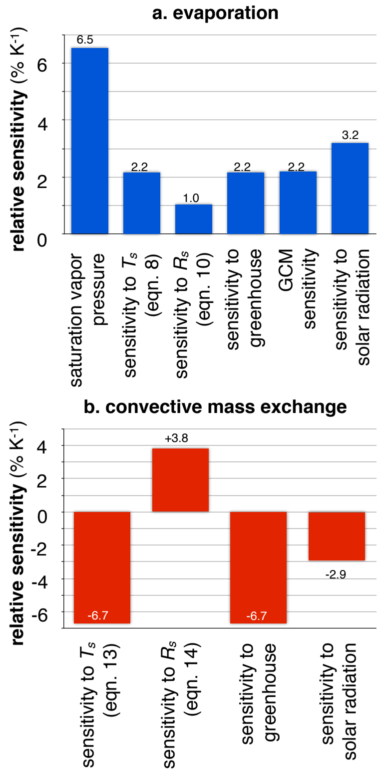

Fig. 2. Sensitivity of **(a)** the hydrologic cycle (evaporation E) and **(b)** convective mass exchange (exchange velocity w) to differences in surface temperature (T_s). Shown are the numerical values for the relative sensitivities as given in the text for present-day conditions. Also included in **(a)** 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").

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