Entropy Production by Evapotranspiration and its Geographic Variation

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Abstract: The hydrologic cycle is a system far from thermodynamic equilibrium that is characterized by its rate of entropy production in the climatological mean steady state. Over land, the hydrologic cycle is strongly affected by the presence of terrestrial vegetation. In order to investigate the role of the biota in the hydrologic cycle, it is critical to investigate the consequences of biotic effects from this thermodynamic perspective. Here I quantify entropy production by evapotranspiration with a climate system model of intermediate complexity and estimate its sensitivity to vegetation cover. For present-day conditions, the global mean entropy production of evaporation is 8.4 mW/m^2/K, which is about 1/3 of the estimated entropy production of the whole hydrologic cycle. On average, ocean surfaces generally produce more than twice as much entropy as land surfaces. On land, high rates of entropy production of up to 16 mW/m^2/K are found in regions of high evapotranspiration, although relative humidity of the atmospheric boundary layer is also an important factor. With an additional model simulation of a "Desert" simulation, where the effects of vegetation on land surface functioning is removed, I estimate the sensitivity of these entropy production rates to the presence of vegetation. Land averaged evapotranspiration decreases from 2.4 to 1.4 mm/d, while entropy production is reduced comparatively less from 4.2 to 3.1 mW/m^2/K. This is related to the reduction in relative humidity of the atmospheric boundary layer as a compensatory effect, and points out the importance of a more complete treatment of entropy production calculations to investigate the role of biotic effects on Earth system functioning.

Keywords: hydrologic cycle; evapotranspiration; land surface; entropy production; principle of Maximum Entropy Production; vegetation effects

The Earth system is dominated by irreversible processes (Peixoto *et al.* 1991; Goody 2000; Kleidon & Lorenz 2005). For instance, solar radiation, once absorbed at the surface, cannot be re-emitted at the same wavelengths, but is emitted as long-wave radiation at the much colder temperatures of the Earth's surface compared to the hot emission temperature of the Sun. Hence, absorption of solar radiation at the prevailing temperatures of the Earth's surface is irreversible. Likewise, the process of evaporation from the Earth's surface into the atmospheric boundary layer cannot be reversed unless the evaporated moisture is cooled to saturation, usually through lifting by atmospheric motion. And atmospheric motion, in turn, is driven by the degradation of gradients in heating, and the associated generation of kinetic energy is dissipated by friction into heat, mostly within the planetary boundary layer. But it is not possible to convert this dissipated heat back into kinetic energy. Thus, the hydrologic cycle is intimately linked with the irreversible nature of Earth system processes.

This irreversibility of Earth system processes is associated with the production of entropy in steady state. When a system is maintained away from thermodynamic equilibrium, processes are directed such that they aim to bring the system back to thermodynamic equilibrium, and they produce entropy by doing so. The magnitude of entropy production hence characterizes the extent to which systems are maintained away from thermodynamic equilibrium. A critical aspect in maintaining this state is that the flexible conditions at the system boundary allow the entropy produced by the processes within the system to be exported to the surroundings. This steady state is then characterized by a comparatively low entropy of the system, a continuous rate of entropy production within the system, and, equivalently, by a net export of entropy to the surroundings.

When we investigate the hydrologic cycle from this perspective, we note that it is such a system far from thermodynamic equilibrium. The state of thermodynamic equilibrium of atmospheric moisture over a wet surface corresponds to the case in which the atmosphere is saturated with water vapor. In this equilibrium state, the process of evaporation and condensation would be reversible, because in the saturated atmosphere these phase transitions would occur at the same temperature. The case of an unsaturated atmosphere (i.e. relative humidity < 100%) thus reflects a state away from thermodynamic equilibrium. The process of evaporation at the surface then becomes irreversible and it produces entropy by the mixing of saturated air from the surface where evaporation takes place with the unsaturated air of the boundary layer. In other words, the process of evaporation is directed to bring the moisture content of the atmospheric boundary layer back closer to thermodynamic equilibrium. The atmospheric circulation plays a critical role here because it is the driver that maintains this state far from equilibrium (Pauluis & Held 2002a, b).

Previous studies have estimated the amount of entropy production (EP) by the hydrologic cycle from global averages of the latent heat flux of LH = 79 W/m^2, the mean surface temperature of T_s = 288 K, and an average temperature of condensation of T_c = 266 K (Peixoto *et al.* 1991). The average number obtained in this way is EP = LH × $(1/T_c - 1/T_s) \approx$ 23 mW/m^2/K. This estimate is derived in a rather crude way, and more detailed studies focus primarily on atmospheric processes (e.g. Goody 2000; Pauluis *et al.* 2000). The quantification of entropy production is important because (a) it characterizes its thermodynamic nature, which is only partially captured by the energy balance, and (b) because it has been suggested that sufficiently complex systems adopt steady states at which the rate of entropy production is maximized (proposed principle of Maximum Entropy Production, e.g. Ozawa *et al.* 2003; Kleidon & Lorenz 2005; Martyushev & Seleznev 2006). Applied to the hydrologic cycle, this line of reasoning would imply that the hydrologic cycle is maintained furthest away from thermodynamic equilibrium at a comparatively low average relative humidity, and that it is likely to be associated with maximum precipitation and evapotranspiration rates. Since the terrestrial biota substantially affects the exchange fluxes of energy and water at the land surface, this raises the question of what the role of terrestrial vegetation is in maintaining this thermodynamic state.

It has previously been suggested that life plays a critical role in maintaining the Earth system in a thermodynamic state far from equilibrium, e.g. in terms of atmospheric oxygen concentrations (Lovelock 1965; Lovelock & Margulis 1974) in terms of ecosystem functioning (Ulanowicz & Hannon 1987) and in terms of atmosphere-biosphere interactions (Kleidon 2004). This, however, requires a more detailed understanding of the dissipative nature of land surface exchange processes.

The goal of this paper is to perform a first step in quantifying entropy production associated with land surface evapotranspiration and its sensitivity to vegetation cover. This estimate is derived from a climate model simulation using realistic, present-day forcings. Using a climate model has the great advantage of being able to obtain a consistent global-scale estimate as well as its geographic variation, but also to account for the covariances and non-linearities in the variables that are needed in calculating entropy production rates (see methods section below). This, however, comes at the cost that the present-day climate may not be exactly representing observed variations. These discrepancies should be kept in mind in the interpretation of the results. In addition, a separate climate model simulation representing a "Desert" world is performed to obtain a first order estimate of the role of terrestrial vegetation for entropy production associated with evapotranspiration.

METHODS

Entropy production of evapotranspiration

Entropy production generally takes the shape of a product of a thermodynamic force and flux (e.g. Kondepudi & Prigogine 1998). In order to calculate entropy production rates EP_{et} associated with evapotranspiration, we need to consider the evapotranspiration rate (ET) as the mass flux as well as the difference in chemical potentials of the near-saturated air at the surface μ_s and the air of the boundary layer μ_{bl} as the thermodynamic driving force. The rate EP_{et} is calculated by considering the differences in chemical potential as well as the mass flux (ET) and the surface temperature T_s:

$$EP_{et} = (\mu_{bl} - \mu_s) \times ET/T_s \quad (1)$$

The chemical potential μ of moist air is related to the relative humidity RH and temperature T (Campbell & Norman 1998):

$$\mu = R_v \times T \times \ln RH \quad (2)$$

where:
R_v = 461.5 J/kg/K

When Eqs (1) and (2) are combined, we obtain the following expression:

$$EP_{et} = R_v \times ET \times \ln RH_{bl} \quad (3)$$

where:
RH_{bl} – relative humidity of the atmospheric boundary layer

Note that the relative humidity of the surface air ($RH_s \approx 100\%$) is omitted here because it is essentially saturated and therefore has a chemical potential of $\mu_s \approx 0$.

Model description

In order to compute EP_{et}, its geographic variation as well as its sensitivity to vegetation cover, Eq. (3) is incorporated into a climate system model of intermediate complexity (Fraedrich *et al.* 2005). This model explicitly simulates radiative transfer, atmospheric dynamics, moist processes and clouds as well as the surface energy- and water balance and includes a simple dynamic vegetation model. It runs on a spatial explicit grid of a resolution of about 5.5° longitude × 5.5° latitude with a time step of 45 minutes.

Land surface hydrology is simulated at every grid point of the model using a simple budget approach ("bucket" model). Evapotranspiration, as the major flux of interest here, is simulated using the "bulk" formula, i.e. it depends on simulated wind speed, atmospheric stability, vapor pressure deficit, and soil water availability. The relative humidity of the atmospheric boundary layer is calculated using the specific humidity and temperature of the lowest atmospheric layer. A dynamic vegetation model simulates biomass dynamics from simulated rates of gross primary productivity and respiration. See Fraedrich *et al.* (2005) and Kleidon (2006) for more details of the model.

Simulation setup

I conduct two model simulations: (i) a "Control" simulation representing the present-day climate is conducted to establish a reference case; and (ii) a "Desert" simulation is conducted in which the effect of vegetation on land surface functioning is removed (Kleidon *et al.* 2000; Kleidon 2006). The removal of vegetation in the model results primarily in a higher surface albedo, lower surface roughness, and a lower ability to extract moisture from the soil. Both simulations are run with prescribed climatological sea-surface temperatures, that is, oceanic feedbacks are explicitly excluded from the simulations. Both simulations are run for 40 years. The "Control" simulation is run with the dynamic vegetation model in an accelerated mode to reach the steady state after approx. 30 years.

Table 1. Annual mean values of evaporation, relative humidity of the atmospheric boundary layer and entropy production averaged over land (all non-glaciated land grid points), oceans, and globally; the last column refers to annual means averaged over land in the "Desert" climate model simulation

	Land	Ocean	Global	"Desert"
Evaporation (mm/d)	2.4	3.3	3.1	1.4
Relative humidity (%)	57	65	63	46
Entropy production ($mW/m^2/K$)	4.2	9.8	8.4	3.1

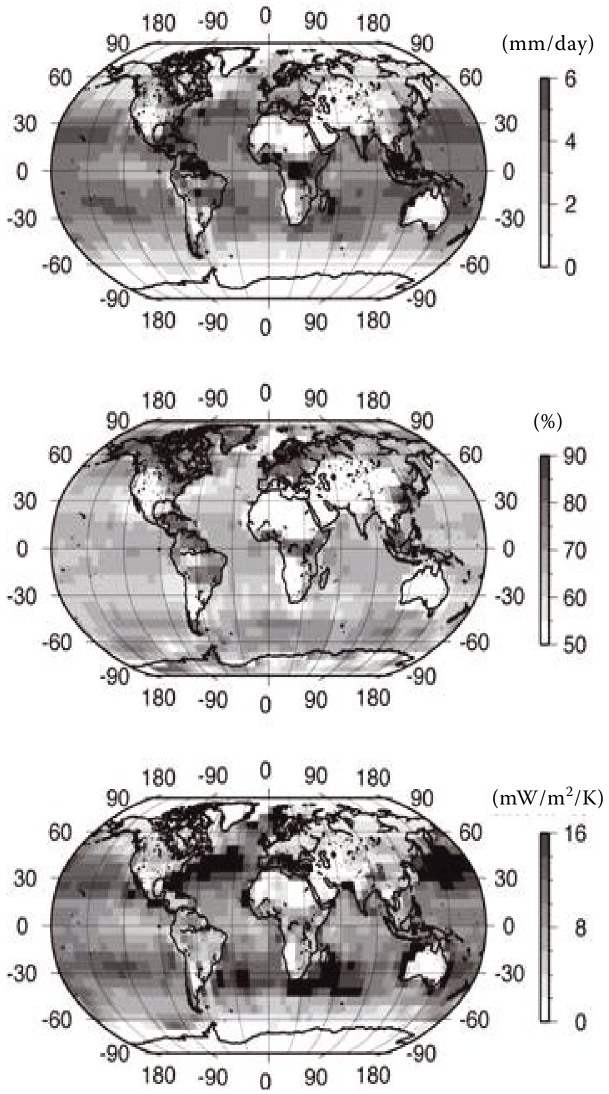

Figure 1. Annual mean values of evapotranspiration (top), relative humidity of the atmospheric boundary layer (middle) and entropy production associated with evapotranspiration (bottom) for the present-day "Control" simulation

The last 5 years are used to compute the climatological mean state.

RESULTS AND DISCUSSION

Figure 1 shows maps of annual mean evapotranspiration rates, relative humidity of the atmospheric boundary layer, and the associated rates of entropy production. The respective average taken over all ocean and land regions as well as global averages are shown in Table 1. Global mean entropy production associated with surface evapo(transpi)ration is 8.4 mW/m^2/K, which is about 1/3 of the estimated entropy production of the whole hydrologic cycle. The remaining 2/3 consequently would result from processes within the atmosphere, e.g. the dissipation of kinetic energy of falling raindrops (Pauluis *et al.* 2000). A disproportionate contribution to the entropy production rate comes from the ocean surface since the average per unit area over oceans (9.8 mW/m^2/K) is more than twice of the respective value over land (4.2 mW/m^2/K). This is clearly to a large extent due to the presence of deserts over land where ET is limited by water availability thus reducing the average value (Figure 1).

Entropy production varies greatly among regions, from values of near zero in the polar regions to rates as high as 16 mW/m^2/K and more in the mid-latitude regions over the oceans. The peak values of EP_{et} are clearly not only driven by patterns of ET alone, but reflect variations in relative humidity as well. On land, the geographic patterns broadly follow the patterns of evapotranspiration, but are not quite as pronounced. This is more easily seen in Figure 2, which shows a scatter plot of annual means of these variables for each land grid point. While entropy production is positively correlated

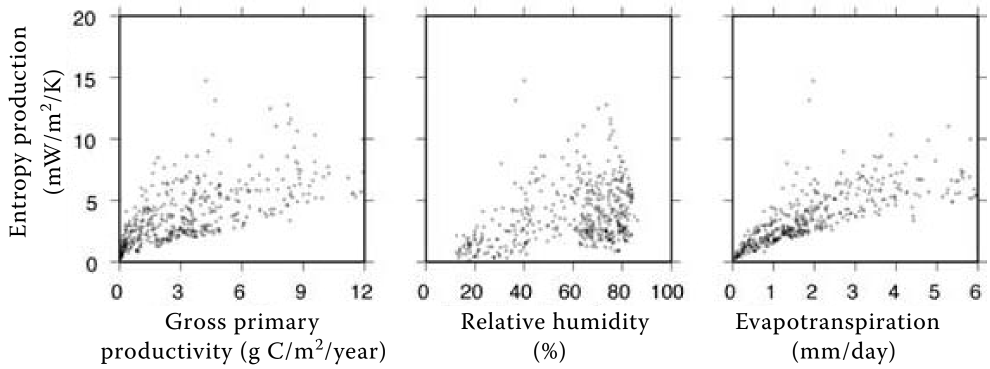

Figure 2. Scatter plot of simulated annual mean values of entropy production associated with land evapotranspiration in comparison to annual means of gross primary productivity (left), relative humidity of the atmospheric boundary layer (middle), and evapotranspiration (right); each circle represents the annual means of one land grid point respectively

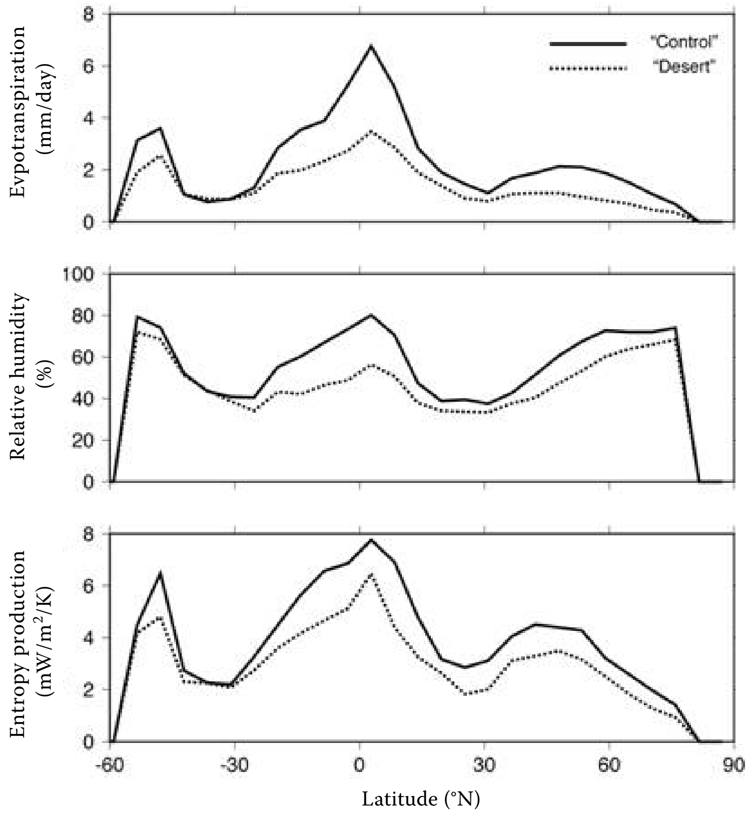

Figure 3. Annual land means of evapotranspiration (top), relative humidity of the atmospheric boundary layer (middle) and entropy production associated with evapotranspiration (bottom) for the "Control" (solid lines) and "Desert" (dotted lines) simulations

with gross primary productivity and evapotranspiration, the relationship to annual mean values of relative humidity of the atmospheric boundary layer is more complicated. This reflects the combined effects of (a) lower relative humidity should result in higher entropy production rates (cf. Eq. (3)), but (b) regions of high water availability (and thus evapotranspiration rates) also tend to have a higher relative humidity as well.

The land averages of the "Desert" simulation are also shown in Table 1. While land evapotranspiration decreases from 2.4 mm/d to about half its value of 1.4 mm/d, EP_{et} decreases only by about 25% to 3.1 mW/m^2/K. This reduced sensitivity of EP_{et} results from the lower relative humidity of the atmospheric boundary layer over land (which decreased from an average of 57% down to 46%). This, in turn, compensates for the decrease in ET in the overall sensitivity of EP_{et}. This is more clearly shown in the zonal mean plots in Figure 3. The reduction of ET to 50% of its value are clearly visible and spread relatively uniformly across the vegetated regions. The decrease in relative humidity follows similar patterns as the decrease in ET, emphasizing the tight coupling between evapotranspiration, relative humidity, and, more generally, continental moisture recycling in the model. Consequently, the reduction in EP_{et} is not as pronounced.

This decrease in entropy production is less than that reported by previous modelling studies (Kleidon 2004, 2006). Kleidon (2004) lists a decrease from 13 mW/m^2/K down to 5 mW/m^2/K for a comparable decrease in ET by 1 mm/d. However, this has been estimated following the approach described in the introduction, not the more direct approach of using the mass flux and chemical potential difference (Eq. (3)), and hence does not constitute the rate of entropy production with respect to evapotranspiration, but rather represent a rough estimate of total entropy production associated with water cycling over land.

In summary, I presented an analysis of entropy production of evapotranspiration, how it varies geographically, and how sensitive it is to vegetation cover. Clearly, more work and analysis needs to be done in this direction. A more detailed thermodynamic analysis of the coupled energy- and water budget over land in models/theory (Kleidon *et al.* 2007) and field observations (see e.g. Tesař *et al.* 2007) would seem to provide important insights about how the water exchange fluxes work, how these are affected by the biota, and how these are integrated within the dissipative Earth system.

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