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**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  $8.4 \text{ mW m}^{-2} \text{ K}^{-1}$ , 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  $16 \text{ mW m}^{-2} \text{ K}^{-1}$  are found in regions of high evapotranspiration, although relative humidity 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  $\text{mm d}^{-1}$ , while entropy production is reduced comparatively less from 4.2 to 3.1  $\text{mW m}^{-2} \text{ K}^{-1}$ . This is related to the reduction in relative humidity 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:** evapotranspiration, land surface, entropy production, principle of Maximum Entropy Production, vegetation effects.

## **INTRODUCTION**

The Earth system is dominated by irreversible processes (Peixoto et al. 1991, Goody 2000, Kleidon and Lorenz 2005). For instance, solar radiation, once absorbed at the surface, cannot be re-emitted at the same wavelengths, but is emitted as longwave 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 as it is the driver that maintains this state far from equilibrium (Pauluis and 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  $LH = 79 \text{ W m}^{-2}$ , the mean surface temperature of  $T_s = 288\text{K}$ , and an average temperature of condensation of  $T_c = 266\text{K}$  (Peixoto et al. 1991). The average number obtained in this way is  $EP = LH * (1/T_c - 1/T_s) \approx 23 \text{ mW m}^{-2} \text{ K}^{-1}$ . 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 and Lorenz 2005, Martyushev and 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 and Margulis 1972) in terms of ecosystem functioning (Ulanowicz and 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 and 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  $\mu_s$  and the air of the boundary layer  $\mu_{bl}$  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  $T_s$ :

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

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

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

where  $R_v = 461.5 \text{ J kg}^{-1} \text{ K}^{-1}$ . When eqns. (1) and (2) are combined, we obtain the following expression:

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

Where  $RH_{bl}$  is the relative humidity of the 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, eqn. (3) is incorporated into a climate system model of intermediate complexity (Plasim, 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  $5.5^\circ$  longitude \*  $5.5^\circ$  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. 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. The last 5 years are used to compute the climatological mean state.

## **RESULTS AND DISCUSSION**

Fig. 1 shows maps of annual mean evapotranspiration rates, relative humidity, 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  $8.4 \text{ mW m}^{-2} \text{ K}^{-1}$ , 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 ( $9.8 \text{ mW m}^{-2} \text{ K}^{-1}$ ) is more than twice of the respective value over land ( $4.2 \text{ mW m}^{-2} \text{ K}^{-1}$ ). 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 (see Fig. 1).

Entropy production varies greatly among regions, from values of near zero in the polar regions to rates as high as  $16 \text{ mW m}^{-2} \text{ K}^{-1}$  and more in the mid-latitude regions over the oceans. The peak values of  $EP_{\text{et}}$  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 Fig. 2, which shows a scatter plot of annual means of these variables for each land grid point. While entropy production is positively correlated with gross primary productivity and evapotranspiration, the relationship to annual mean values of relative humidity is more complicated. This reflects the combined effects of (a) lower relative humidity should result in higher entropy production rates (cf. eqn. (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  $2.4 \text{ mm d}^{-1}$  to about half its value of  $1.4 \text{ mm d}^{-1}$ ,  $EP_{\text{et}}$  decreases only by about 25% to  $3.1 \text{ mW m}^{-2} \text{ K}^{-1}$ . This reduced sensitivity of  $EP_{\text{et}}$  results from the lower relative humidity 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  $EP_{\text{et}}$ . This is more clearly shown in the zonal mean plots in Fig. 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, Kleidon 2006). Kleidon (2004) lists a decrease from  $13 \text{ mW m}^{-2} \text{ K}^{-1}$  down to  $5 \text{ mW m}^{-2} \text{ K}^{-1}$  for a comparable decrease in ET by  $1 \text{ mm d}^{-1}$ . 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 (eqn. 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. Tesar 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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**TABLES:**

	Land	Ocean	Global	“Desert”
Evaporation (mm d <sup>-1</sup> )	2.4	3.3	3.1	1.4
Relative humidity (%)	57	65	63	46
Entropy production (mW m <sup>-2</sup> K <sup>-1</sup> )	4.2	9.8	8.4	3.1

**TABLE I:** Annual mean values of evaporation, relative humidity 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.

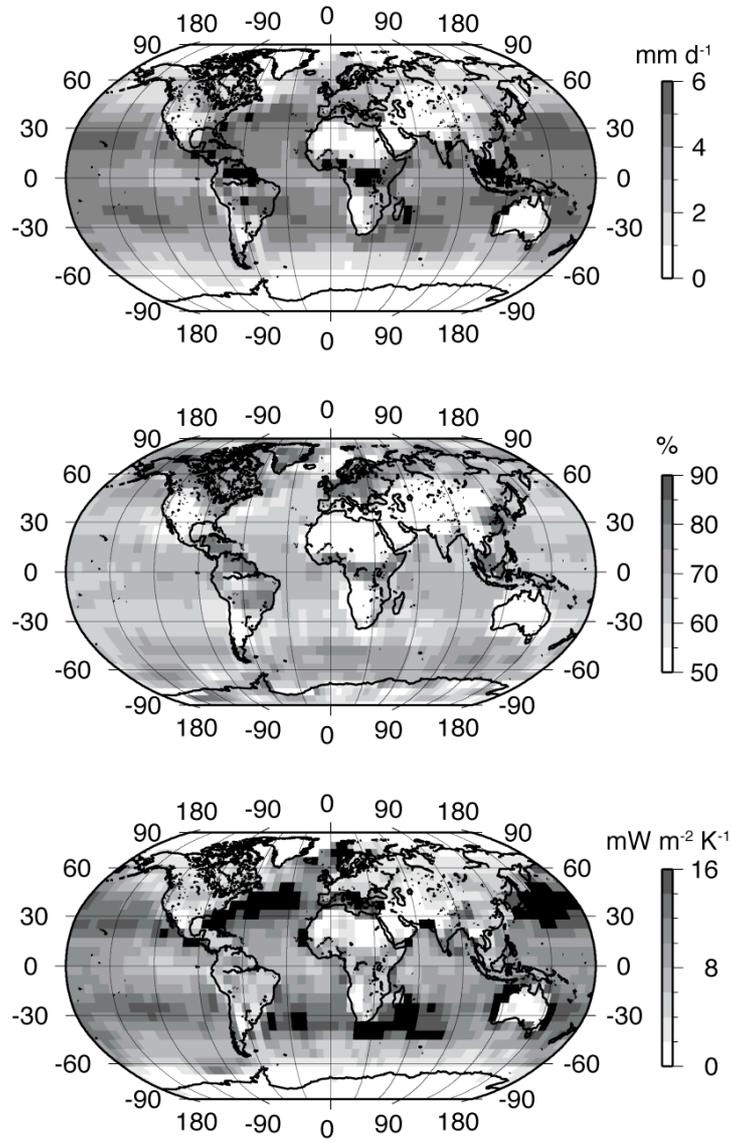
## **FIGURE CAPTIONS**

**FIGURE 1:** Annual mean values of evapotranspiration (top), relative humidity (middle) and entropy production associated with evapotranspiration (bottom) for the present-day “control” simulation.

**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 (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 (middle) and entropy production associated with evapotranspiration (bottom) for the “control” (solid lines) and “desert” (dotted lines) simulations.

**FIGURE 1:**



**FIGURE 2:**

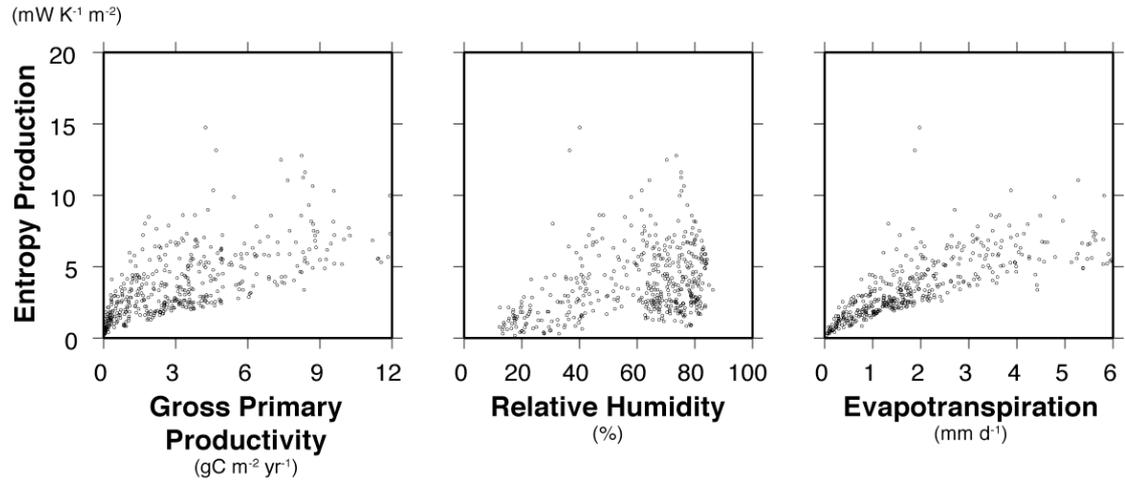


FIGURE 3:

