Maximum entropy production and the strength of boundary layer exchange in an atmospheric general circulation model

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[1] Boundary layer turbulence plays a central role in determining the strength of the overall atmospheric circulation by affecting the intensity of exchange of heat, mass, and momentum at the Earth's surface. It is often parameterized using the bulk formula, in which the von-Karman parameter plays a critical role. Here we conducted a range of sensitivity simulations with an atmospheric general circulation model in which we modified the strength of boundary layer turbulence by varying the von-Karman parameter. These simulations show that the maximum of entropy production associated with boundary layer dissipation is consistent with the observed value of the von-Karman parameter of 0.4 and maximizes the planetary rate of entropy production with the global radiative temperature being close to its minimum value. Additional sensitivity simulations were conducted with an increased concentration of atmospheric carbon dioxide, which affects the relative radiative forcing of tropical vs. polar regions. We find that the global climate sensitivity is more-or-less independent of the assumed strength of boundary layer turbulence in our idealized setup. The difference in climate sensitivities of tropical and polar regions is at a minimum at a climatic state of MEP. **Citation:** Kleidon, A., K. Fraedrich, E. Kirk, and F. Lunkeit (2006), Maximum entropy production and the strength of boundary layer exchange in an atmospheric general circulation model, *Geophys. Res. Lett.*, *33*, L06706, doi:10.1029/2005GL025373.

1. Introduction

[2] The strength of the atmospheric circulation is determined by the balance of generation and dissipation of kinetic energy. The latter term is strongly dependent on the exchange of momentum at the surface. Several studies suggest that the atmospheric circulation maintains a state that is close to maximizing the rate of entropy production (MEP) [e.g., *Lorenz*, 1960; *Paltridge*, 1978; *Ozawa and Ohmura*, 1997; *Lorenz et al.*, 2001; *Paltridge*, 2001; *Ozawa et al.*, 2003; *Kleidon and Lorenz*, 2005]. This notion is also consistent with the upper-bound theory of heat transport, developed by W. V. R. Malkus and F. H. Busse [*Malkus*, 1954, 1956; *Busse*, 1970], which is equivalent to MEP with fixed temperatures as boundary conditions [*Ozawa et al.*, 2001]. *Dewar* [2003, 2005] provided a theoretical foundation to MEP. However, MEP has been primarily used in energy balance climate models, and the question arises whether and how MEP is applicable to more detailed treatment of fluid dynamics, as is the case in the atmospheric general circulation models (GCMs).

[3] *Kleidon et al.* [2003] demonstrated the relevance of MEP with sensitivity simulations with respect to model resolution and boundary layer turbulence with a dynamic core GCM. Here we extend our earlier study and investigate the role of turbulent exchange at the surface in more detail in a GCM of intermediate complexity (the Planet Simulator [*Lunkeit et al.*, 2004; *Fraedrich et al.*, 2005a, 2005b]. In this model, the surface-to-atmosphere exchange fluxes of momentum and heat, $F_{u,v}$ and F_t are expressed by the bulk formula approach [e.g., *Louis*, 1979; *Louis et al.*, 1981; *Roeckner et al.*, 1992]:

$$F_{u,v} = \rho C_m |\vec{v}|(u, v) \tag{1}$$

$$F_t = c_p \rho C_h |\vec{v}|(T_a - T_s) \tag{2}$$

where c_p is the specific heat capacity of air, ρ the density of air, $\vec{v}$ the wind vector with components (u, v), and T_a and T_s are the near surface atmospheric and surface temperature respectively. The latent heat flux F_q is parameterized similarly. The drag coefficients C_m and C_h are of the form

$$C_{(m,h)} = (k/\ln(z/z_0))^2 \cdot f_{(m,h)}(Ri, z/z_0) \tag{3}$$

where z is height from the surface, k is the von-Karman parameter, z_0 is the surface roughness, and f is an empirical function dependent on stability (as expressed by the Richardson number Ri) and surface roughness.

[4] In order to investigate the role of turbulent exchange for atmospheric entropy production, we conduct a range of sensitivity simulations with the GCM in which we vary the value of the von-Karman parameter k. We chose to vary k because it directly affects the strength of $C_{(m,h)}$. If MEP is applicable, it should predict an optimum value of k close to the commonly used, empirical value of about $k = 0.4$.

2. Methods

2.1. Climate Model

[5] The Planet Simulator consists of a low resolution atmospheric general circulation model with 5 vertical layers and T21 spectral resolution, corresponding to a grid resolution of about 5.625° * 5.625° longitude/latitude, with a dynamical core and a physical parameterization package of intermediate complexity for unresolved subgrid-scale processes.

2.2. Entropy Budget Calculations

[6] We add diagnostic entropy flux calculations to the model code in order to determine the global entropy budget.

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For each heating and cooling term the associated entropy flux Q/T is calculated, with T being the associated temperature at which the heat Q is being removed or released. These fluxes are calculated for the different components of the simulated energy balance: absorption and emission of shortwave and longwave radiation, vertical turbulent exchange fluxes at the surface and in the atmosphere, and horizontal heat fluxes in the atmosphere.

[7] For most components, the addition of these calculations is straightforward since the heating terms and the respective temperatures are readily available within the model code. The calculation of entropy fluxes associated with longwave radiative transfer is more complex since the radiative transfer code yields net radiative heating fluxes, but the entropy flux computations require information regarding the temperatures at which the gross radiative fluxes were emitted and absorbed. We perform this computation by duplicating the radiative transfer code for calculations of Q/T in addition to Q. Entropy fluxes due to horizontal heat transport out of an atmospheric column are calculated by adding up the overall net heating divided by temperature for each of the atmospheric model layers.

[8] In the climatic steady state, the net flux of entropy across the system boundary balances the production of entropy within the system [e.g., *Peixoto et al.*, 1991]. We use this steady-state assumption to compute the individual components of the global entropy budget by averaging each of the simulated components. Since the energy fluxes in the model do not balance exactly (which is often the case in general circulation models, with discrepancies in the order of 5 W m^{-2} or less), we apply a correction scheme where we compute a scaling factor to match the sum of all cooling terms to be equal to the sum of all heating terms. This scaling factor is then applied to the entropy fluxes before these are integrated to yield entropy production terms. Since this correction results in a shift only, it affects the magnitude of entropy production, but leaves the overall conclusions of the study unaffected.

[9] In addition to the entropy production terms of the energy balance components, the entropy budget also allows us to compute (a) the planetary rate of entropy production by using the entropy fluxes at the top of the atmosphere (TOA), as well as (b) the effective radiative temperature, by dividing the outgoing TOA flux of terrestrial radiation by its respective entropy flux.

2.3. Simulation Setup

[10] We use an idealized model setup in which the amount of absorbed solar radiation is fixed in order to exclude complicating feedbacks by clouds and sea-ice, so that we can compare the results obtained here to our previous study [*Kleidon et al.*, 2003]. This setup includes a global ocean surface (an "Aquaplanet" setup) with a 50m mixed-layer, excludes oceanic heat transport, excludes water cycle and sea-ice calculations in order to keep the planetary albedo fixed for all sensitivity simulations. The simulations also include realistic computations of the seasonal cycle in solar radiation representative of present-day conditions. In this setup, entropy is produced by radiative transfer (absorption of solar and terrestrial radiation at the surface and in the atmosphere), the turbulent transport of sensible heat in the vertical, and horizontal heat transport by the large-scale atmospheric circulation. Due to the exclusion of the water cycle, entropy production associated with the hydrologic cycle [*Pauluis and Held*, 2002a, 2002b] is not considered.

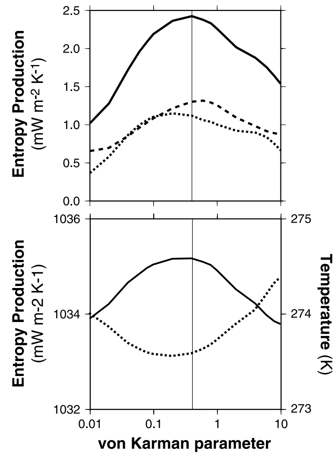

Figure 1. Sensitivity of globally averaged entropy production rates to different values of the von-Karman parameter. (top) Entropy production by poleward heat transport σ_{aht} (dotted line), sensible heat flux σ_{sh} (dashed line), and the sum of both $\sigma_{turb} = \sigma_{aht} + \sigma_{sh}$(i.e., entropy production by overall atmospheric turbulence, solid line). (bottom) Planetary entropy production σ_{tot} (solid line) and radiative temperature T_{rad} (dotted line). The thin vertical line marks the value of the von-Karman parameter of the control simulation of $k = 0.4$.

[11] We conduct a series of sensitivity simulations in which we vary the value of the von-Karman parameter k. The standard value is $k = 0.4$. Sensitivity simulations are conducted with values of $k = 0.01$, 0.02, 0.04, 0.06, 0.08, 0.1, 0.2, 0.4, 0.6, 0.8, 1.0, 2.0, 4.0, 6.0, 8.0 and 10.0. The von-Karman parameter is externally prescribed to one global value (i.e., regional variations are not allowed for) and is kept constant throughout the simulation. Each sensitivity simulation is run for 30 years, with the last 5 years used for evaluations. The time step in the simulations is decreased to appropriate values in the more extreme simulations in order to ensure the numerical stability.

[12] The sensitivity simulations are repeated for a setup in which the atmospheric concentration of carbon dioxide pCO_2 is increased by a factor of 10. These simulations are conducted to estimate the extent to which the sensitivity of the entropy budget changes under altered radiative forcings. A factor of 10 was chosen to significantly alter the radiative forcing in the model. Note that the overall climate sensitivity to pCO_2 of the model is substantially reduced due to the exclusion of the water cycle and sea-ice feedbacks.

3. Results and Discussion

[13] Figure 1 shows the global averages of entropy production associated with vertical, horizontal, and total turbulent heat transport in the atmosphere. The maximum in entropy production associated with atmospheric turbulence is reached with $k = 0.4$ (although the maxima for the two

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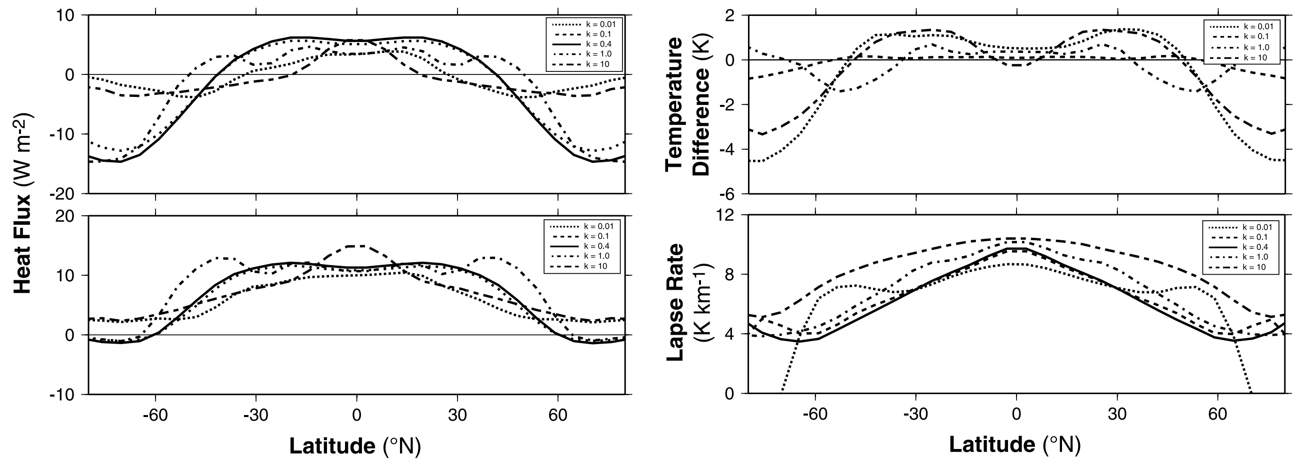

Figure 2. Zonal averages of net radiative and sensible heat fluxes (top and bottom left respectively) and respective zonal surface temperatures (top right, difference to "Control" simulation) and lapse rate (bottom right) for selected sensitivity simulations.

contributions occur at slightly different values - the existence of these maxima are consistent to previous studies [e.g., *Paltridge*, 1978; *Ozawa and Ohmura*, 1997]. The magnitude of entropy production is similar in value to our earlier study with a dynamic core GCM [*Kleidon et al.*, 2003].

[14] Figure 1 also shows that the planetary rate of entropy production is at a maximum at $k = 0.4$. Since the absorption of solar radiation is fixed, this maximum results from a lowering of the effective radiative temperature. The lowering of the effective radiative temperature in turn results from a more effective heat distribution by the atmospheric heat transport. This is, of course, to be expected as the increased production of entropy by atmospheric turbulence is ultimately rejected into space in form of longwave radiation. As more entropy is produced by atmospheric turbulence for certain values of k, the entropy flux to space is likely to increase accordingly (although shifts in the contributions by, e.g., absorption of radiation could partially compensate for enhanced entropy production by atmospheric turbulence).

[15] The climatic differences in heat fluxes, equator-pole temperature difference, and near-surface lapse rate are shown in Figure 2 for different values of k. We find that the climatic state that results in MEP is associated with the greatest magnitude of meridional heat transport, therefore resulting in the smallest equator-pole temperature difference in our simulations. The climatic state of MEP in the GCM results from a dynamic constraint, the 'barotrophic governor' [*James and Gray*, 1986]: With the overall kinetic energy increasing with lower values of k, the circulation becomes increasingly barotropic and stable to baroclinic disturbances. Furthermore, poleward heat transport does not increase with overall kinetic energy, but decreases beyond a certain point. In our simulations, the MEP climate is the baroclinically most unstable and is associated with the weakest jet stream (not shown). This is consistent with our earlier results [*Kleidon et al.*, 2003]. An implication is that the dynamic constraint, and therefore the MEP state, is sensitive to planetary rotation rate. This is an important difference to the simplified reasoning of flux-force tradeoffs in energy balance models that do not account for dynamical constraints and this addresses a critical limitation of previous applications of MEP to the atmospheric circulation.

[16] We also find that the sensible heat flux is generally near its maximum value (although not uniformly at all latitudes). This latter aspect is similar to Paltridge's maximum convection hypothesis [*Paltridge*, 1975], although here it results from the maximization of entropy production of total atmospheric turbulence, rather than the maximization of vertical turbulent heat fluxes.

[17] The optimum value of k is unaffected by elevated pCO_2 (not shown). The climate sensitivity to pCO_2 for different values of k is shown in Figure 3. While the global sensitivity is not affected by different values of k, the sensitivity of tropical (45°N – 45°S) versus extratropical regions (45° - poles) differs substantially. The climate sensitivity is most uniform across regions near values of k close to the MEP state. This results naturally from the results shown above: as heat transport is most effective at MEP, we should expect differences in radiative forcing to be smoothed out as much as possible, but not at different, non-MEP settings.

4. Summary and Conclusion

[18] We have shown with idealized climate model simulations that the strength of boundary layer turbulence

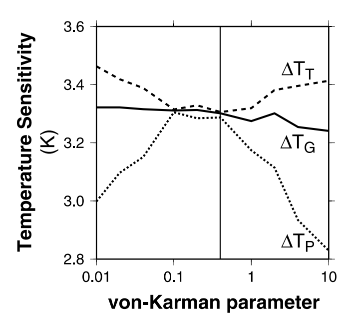

Figure 3. Sensitivity of global mean surface temperature (solid line), tropical surface temperature (dashed line) and polar temperature (dotted line) to an increase in pCO_2 by a factor of 10. The vertical line marks the value of the von-Karman parameter of the control simulation of $k = 0.4$.

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associated with the empirical value of the von-Karman parameter of $k = 0.4$ results in a climate in which the atmosphere works the hardest, thereby dissipating the most kinetic energy and maximizing its rate of entropy production for present-day conditions. The maximum strength of boundary layer turbulence is not affected by changes in the longwave radiative forcing induced by elevated concentrations of pCO_2 in the simulations. The extent to which this maximum strength is sensitive to other external forcing factors, such as solar radiation and planetary rotation rate, remains to be investigated. Further sensitivity simulations may show that the optimum value of the von-Karman parameter is not the same for all climatic settings, which may be relevant for the adequate simulation of atmospheric circulations on other planetary systems.

[19] The results presented here provide an important confirmation of the applicability of MEP to complex climate system models. Since MEP quantifies the upper bound of the strength of dissipative processes, MEP serves as a powerful guiding principle for the development and tuning of model parameterizations. Furthermore, the validation of MEP by state-of-the-art modeling tools strengthens the foundations for its applications to a wider range of topics related to the development of climate system theory and how the climate system evolves over time [e.g., *Ou*, 2001; *Kleidon*, 2004].

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