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Landscape variability and surface flux parameterization in climate models

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Abstract

The Earth's surface shows variability at the landscape scale (1–10 km); the influence of surface variability at this scale has been analysed to provide a parameterization for use in large-scale atmospheric models with a grid size unable to solve the landscape scale explicitly. Landscape variations are found to add drag to the atmosphere, owing to sudden changes in vegetation height. The drag increases momentum flux and indirectly influences the transfer of heat and gases between the landscape and the atmosphere. Consequently, the exchange between a variable landscape and the atmosphere deviates from a simple sum of the exchanges between landscape elements and the contiguous air layer. Strong influences are found for tree lines and forest edges. Most of the existing aggregation schemes for grid-averaged fluxes in large-scale models strongly underestimate the consequences of landscape variability owing to the neglect of drag at surface transitions. The supplementary drag can easily be incorporated in an aggregation scheme of surface fluxes in a large-scale model. New experiments on the landscape scale are recommended to improve the accuracy of the method.

1. Introduction

Large-scale atmospheric models for climate or weather forecasting appear sensitive to the parameterization of the exchanges of momentum (Sud et al., 1988), sensible heat and water vapour (Mintz, 1984) between the land surface and the atmosphere. When soil moisture or surface albedo is changed, significant changes in precipitation and temperature take place over the corresponding region. The need to find a satisfactory representation of the land surface in large-scale models has resulted in several

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international experiments, such as HAPEX and FIFE (see review by Shuttleworth (1991)). These experiments are being coordinated by the World Climate Research Programme (WCRP) and the International Geosphere–Biosphere Programme, section Biospheric Aspects of the Hydrological Cycle (IGBP-BAHC). Special emphasis is placed on the aggregation of land surface–atmosphere exchange at scales below the grid size of global climate models. BAHC (Biospheric Aspects of the Hydrological Cycle, 1992) recognizes three important scales for research: the patch (surface dimensions less than 1 km), the region (10–100 km) and the continent (more than 100 km).

Between patch and region we define the landscape scale as having dimensions up to 10 km and consisting of an array of patches. This paper discusses land surface–atmosphere exchange at this landscape scale. Initially, aggregation of surface fluxes at this scale was considered to be less important for climate models. We will start by discussing the significance of this forgotten scale, and subsequently present a concept for incorporating landscape variability in larger-scale models.

2. The problem of flux averaging

In the first generation of global climate models, the surface fluxes were calculated from the dominant surface type in the grid. A significant improvement was obtained by calculating the fluxes for each apparent surface type in the grid, weighted for the area of the relevant surface type (Avissar and Pielke, 1989; Koster and Suarez, 1992). In this so-called mosaic approach, each surface type is assumed to be independently coupled to the (grid averaged) atmospheric boundary layer; interaction by horizontal advection is neglected. Hence this approach is presumably valid for the region scale (10–100 km) where, on the other hand, secondary motions owing to organized surface variations could influence sub-grid scale advection—a process not yet considered in any surface parameterization. To assess the validity of the mosaic approach of averaging surface fluxes, one needs to know whether the advection terms are important or not.

Let us imagine, for example, a nearly neutrally stratified flow over a landscape consisting of a large number of alternating dry and wet strips. Owing to higher evaporation, the air above the wet strips will tend to be wetter and cooler than above the dry strips. When the wind blows perpendicularly to the strips, advection will result in dryer and warmer air above the wet strips and vice versa above the dry strips. As a result, advection enhances evaporation over the wet strip and reduces evaporation over the dry strip. Therefore, on average, the effects of advection counteract (McNaughton, 1976). However, in the case where the dry strips are completely dried out, the evaporation flux of the dry strips is nil and cannot be reduced further by advection, whereas evaporation from the wet strip is enhanced by the dry air advecting from the dried-out strips. In this example, the landscape-averaged evaporation is increased by advection. This situation corresponds, for example, to an oasis in a desert, and to irrigated fields in an arid region. This area-averaged effect of advection is physically explained by non-linear terms in the evaporation equations.

A second example is heat fluxes in the winterly Arctic and Antarctic Marginal Ice

Zones (Claussen, 1991a; Stössel and Claussen, 1994). In those regions, the surface layer is often stably stratified. Nevertheless, the regional heat flux can be directed upward, i.e. counter to the areally averaged vertical temperature gradient, owing to the presence of small patches of open water which cause a strong upward heat flux locally. The upward heat flux results in a destabilization of the surface layer, leading to an increase of momentum flux transfer and stress on the ice floes. Neglecting this feedback would result in an unrealistically small interaction between the atmosphere and cryosphere.

These examples show that the advective fluxes at successive patches are not necessarily cancelling out. The question remains whether this effect is important for parameterization of surface fluxes in climate models. It was first thought that the effect would be small on the landscape scale up to 10 km, as only the lowest part of the planetary boundary layer is influenced by advection (Shuttleworth, 1988). It was quickly recognized that this assumption had not been validated (Shuttleworth, 1991). The importance of advection on the landscape scale is estimated from a short literature review on average roughness.

3. Landscape roughness

Results of a large number of models to calculate average roughness are shown in Fig. 1, adopted from Vihma and Savijärvi (1991) and extended with calculations using the model of Klaassen (1992). To illustrate the principles most simply, neutral conditions are assumed, resulting in the logarithmic wind profile

$$U(z) = U_*/k \ln(z/z_0) \tag{1}$$

where U is the wind velocity, z the height, z_0 the roughness length, k von Kármán's constant and U_* the friction velocity, defined by

$$\tau = \rho U_*^2 \tag{2}$$

with τ the momentum flux to the surface and ρ the air density. One could suppose that the average roughness length can be found by logarithmic averaging. This approach yields the straight line in Fig. 1. However, taking advection into account, most models result in a slightly larger roughness length, and some even in a much larger average roughness length. The model by Mason (1988) has been recalculated for a landscape scale of 1 km, resulting in a larger roughness length than shown by Vihma and Savijärvi (1991) for a 37 km landscape scale. According to Fig. 1, the roughness length estimates vary from 0.002 m to 0.2 m at 10% forest coverage. These differences between the estimates of z_0 strongly influence the estimates of the surface flux as shown in the following example.

The drag coefficient $C_D(z) = [U_*/U(z)]^2$ can be interpreted as the dimensionless momentum flux and varies at 10 m above the surface from 2×10^{-3} to 10×10^{-3} when the roughness length varies from 0.002 m to 0.2 m. The example shows that the various models may result in a factor of five difference in the estimate of the momentum flux of a strongly varying landscape. It should be noted that this example is based

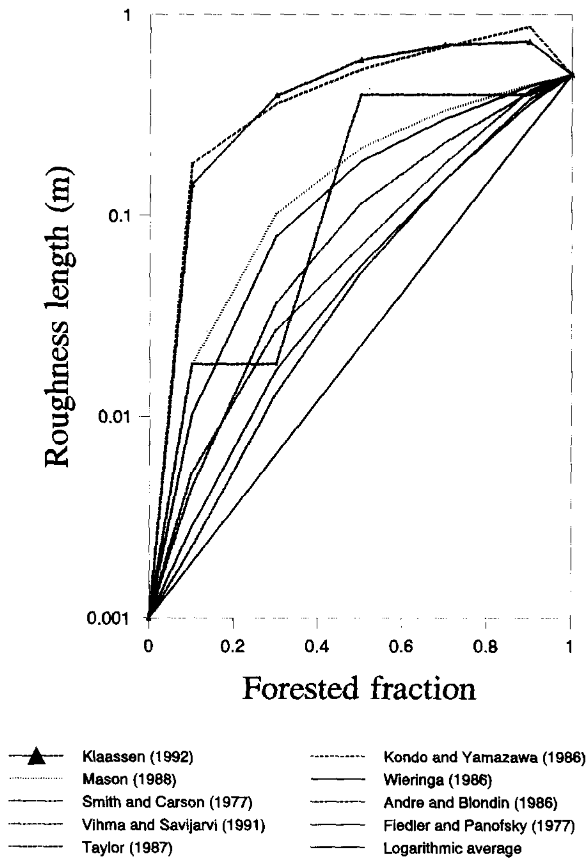

Fig. 1. Average roughness length of a landscape of 1 km length in the wind direction, consisting of one part open water and the remaining part forest as a function of fractional forest cover, using several methods. The figure is based on a figure by Vihma and Savijärvi (1991) and extended with the model result of Klaassen (1992). By simulating explicitly airflow into and out of the forest edges (Klaassen, 1992) results in a high average roughness, in agreement with the empirical model by Kondo and Yamazawa (1986).

on an extreme variation in surface roughness and that in most landscapes the uncertainty will be smaller. Provisionally, we conclude that large variations in roughness at the landscape scale result in an uncertainty in average roughness which seems unacceptable for use in large-scale climate models.

The large scatter of Fig. 1 is mainly caused by the large roughness found by Kondo and Yamazawa (1986) using empirical fitting and the recent model by Klaassen (1992). In particular, the result that a heterogeneous landscape may be aerodynamically rougher than its roughest element is striking. The departure from previous models is caused by a multi-layer representation of the vegetated surface. In this way, the wind can blow not only over the vegetation, but also through the edges of the vegetation. This stronger coupling of surface and atmosphere results in a stronger effect of advection and explains the strong influence of surface variations on the modelled landscape roughness.

An average roughness above the expected roughness of the roughest element is in agreement with observations of Chen and Schwerdtfeger (1989) and Dolman et al. (1992) on tiger bush in Australia and Niger, respectively. Both studies indicated an average roughness of 15–20% of the vegetation height, i.e. above the common value of 13% (Monteith, 1965) for closed vegetation. The roughness length in the agricultural region of HAPEX-MOBILHY was measured by aircraft as of the order of 1 m (Mahrt and Ek, 1993), which is an order of magnitude larger than used in numerical models. The order of magnitude increase was explained by additional drag from scattered trees, forest edges and other obstacles. The average roughness of an extended flat landscape, dominated by windbreaks, was estimated as 1.2 m (Wang and Klaassen, 1995). The average roughness appeared to be in good agreement with the model of Klaassen (1992) and deviated strongly from a simple logarithmic average of the roughnesses of the landscape elements.

From this discussion, we conclude that some models and observations indicate that regional fluxes deviate significantly from a simple averaging procedure, and that this effect is strong enough that it needs to be incorporated in larger-scale atmospheric models. However, the model of Klaassen (1992) is too complicated for use as a lower boundary condition in a large-scale atmospheric model. Therefore, two questions remain: Is it possible to incorporate landscape variability in a simple way in climate models? How can we reduce the uncertainty about the magnitude of the effect (Moore et al., 1993)?

4. The concept of aggregation at the landscape scale

The method of independently coupling surface types to the average atmosphere (as suggested for climate modelling) already implicitly accounts for some advection. It does so in the following way. The air between the land surface and the lowest level of the atmospheric model is assumed to be completely adjusted to the underlying surface type (no advection), and the air above it is assumed to be a uniform mixture of air advected from the various land surface types in the area under consideration. By decreasing the height of the lowest level in an atmospheric model, the height of the adjusted layer is decreased and consequently the influence of advection is increased, in agreement with the results of Mason (1988). A large number of the methods represented in Fig. 1 use a very similar concept to incorporate advection: complete adjustment up to a certain level and complete mixing above. The main difference is in the choice of the height of the transition between completely adjusted air in the atmospheric surface layer and the mixed air above. This level has been designated the 'blending height' (Wieringa, 1986). Instead of taking the blending height equal to the arbitrarily chosen height of the lowest level in the atmosphere model, the blending height has been suggested to be dependent on the height of the planetary boundary layer (Deardorff, 1972), on patch dimensions (Mason, 1988; Claussen, 1991b) or on vegetation height (Wieringa, 1993). By selecting an adequate blending height the problem of aggregation at the landscape scale was thought to be solved.

However, by using a fixed or variable blending height the average roughness is always lower than the roughness of the roughest element. This is in disagreement with observations and model results, as discussed in the previous section. To find a realistic drag from landscapes, a supplementary mechanism for exchange must be distinguished. The 'drag partition' theory (Schlichting, 1936) is a promising method, given by

$$\tau = \tau_D + \tau_S \tag{3}$$

The total momentum absorption of the surface (τ) is given by the sum of the shear stress of the surface elements (τ_S) and the drag of obstacles (τ_D). Now the obstacle drag is identified as the supplementary mechanism for surface–atmosphere exchange. The drag partition theory has already been used to describe the drag around obstacles in a wind-tunnel (Wooding et al., 1973), the large-scale exchange of Arctic pack ice (Arya, 1975), Antarctic sea-ice (Stössel and Claussen, 1994), wind erosion (Raupach et al., 1993) and landscape exchange (Claussen and Klaassen, 1992). For landscape exchange the obstacle drag arises from transitions in vegetation height between patches. In particular, forest edges are important owing to the large height difference with the adjacent fields. By including drag around vegetation edges a more realistic landscape exchange can be obtained, using a method simple enough to incorporate in climate models (Claussen and Klaassen, 1992).

5. Conclusion and recommendation

Experiments and models have shown that landscape exchange is not just a simple sum of homogeneous surface elements. Supplementary drag arises at surface transitions, in particular at tree lines and forest edges. This effect can be described by a drag partition theory. The theory needs to be developed further. In particular, the empirical drag coefficient C_D related to differences in vegetation height needs to be determined more accurately. For instance, the preliminary formulation of Claussen and Klaassen (1992) resulted in $C_D \approx 0.2$ for a landscape with a dense windbreak network, whereas Wang and Klaassen (1995) estimated from measurements in the Great China Plains $C_D \approx 1$. Moreover, the existing experiments of Chen and Schwerdfeger (1989), Dolman et al. (1992) and Wang and Klaassen (1994) have been executed in landscapes with distances between the surface transitions of the order of 0.1 km.

New experiments are set up to examine heterogeneities at an order of magnitude larger scale. It is planned to test the suggested parameterization scheme within the NOPEX and SLIMM experiments. Within the NOPEX project, in Scandinavia, the landscape is dominated by forest with scattered clearings, and the SLIMM project, in the Netherlands, will be executed in a landscape with moorland, agricultural fields and scattered forests. The theory might also be validated within the so-called supersites of experiments at the 100 km scale.

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