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Global and Planetary Change 38 (2003) 93–99
GLOBAL AND PLANETARY CHANGE
www.elsevier.com/locate/gloplacha

The role of aerodynamic roughness for runoff and snow evaporation in land-surface schemes—comparison of uncoupled and coupled simulations

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Received 24 January 2001; received in revised form 14 April 2002; accepted 21 June 2002

Abstract

This paper describes the impact of changes in aerodynamic roughness length for snow-covered surfaces in a land-surface scheme (LSS) on simulated runoff and evapotranspiration. The study was undertaken as the LSS in question produced widely divergent results in runoff, depending on whether it was used in uncoupled one-dimensional simulations forced by observations from the PILPS2e project, or in three-dimensional simulations coupled to an atmospheric model. The LSS was applied in two versions (LSS1 and LSS2) for both uncoupled and coupled simulations, where the only difference between the two versions was in the roughness length of latent heat used over snow-covered surfaces. The results show that feedback mechanisms in temperature and humidity in the coupled simulations were able to compensate for deficiencies in parameterizations and therefore, LSS1 and LSS2 yielded similar runoff results in this case. Since such feedback mechanisms are absent in uncoupled simulations, the two LSS versions produced very different runoff results in the uncoupled case. However, the magnitude of these feedback mechanisms is small compared to normal variability in temperature and humidity and cannot, by themselves, reveal any deficiencies in a parameterization. The conclusion we obtained is that the magnitude of the aerodynamic resistance is important to correctly simulate fluxes and runoff, but feedback mechanisms in a coupled model can partly compensate for errors.

Keywords: aerodynamic roughness; runoff; snow evaporation; land-surface scheme; PILPS

1. Introduction

Validation of model simulations against observations is always an important part of a model development process. However, reliable observations are not always available for this purpose, which includes, for example, regional precipitation and evapotranspiration patterns as part of a model's water budget. An observational data set that can be used for water-budget validations is river discharge data. River flow reflects the difference between precipitation and evapotranspiration over a drainage basin. The advantage is that measurements of river discharge are quite common and are often sampled for long time intervals (Oki et al., 1999). For small drainage basins, runoff produced by the model can be directly compared with river discharge data, while for large

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doi:10.1016/S0921-8181(03)00009-2

drainage basins a routing scheme is necessary if one wants to correctly catch the variability in time of the river flow.

Today, most land-surface schemes (LSSs) used in general circulation models (GCMs), or in regional models, are designed to simulate hydrological processes, where runoff is the final output. This runoff can be used to verify the ability of a coupled atmosphere–LSS model to simulate river flow. However, good agreement between simulated and observed river discharge using a coupled model system does not necessarily mean that the parameterizations in the model are without deficiencies. Feedback mechanisms in a model system may compensate for deficiencies in parameterizations.

One example where such an event can occur concerns the parameterizations of fluxes between the land surface and the atmosphere. In these parameterizations, the aerodynamic resistance, which is a function of wind speed, stability and surface roughness, plays an important role. In simulations forced by observations, Beljaars and Viterbo (1994) show that the roughness length for heat is very important for the simulated evaporation. This is also shown by Chen et al. (1997), who studied the impact of the ratio between momentum and heat roughness lengths on simulations of surface heat fluxes and surface skin-temperature in numerical weather prediction models. Based on the same data set as the present study, van den Hurk and Viterbo (2003-this issue) also showed the sensitivity of aerodynamic exchange on snow sublimation.

This paper shows how the presence or absence of feedback mechanisms affects simulated runoff and how such mechanisms can compensate for deficiencies in an LSS with respect to the parameterization of aerodynamic resistance.

2. Background

This study was performed within the framework of PILPS2(e), which is one of many subprojects in the Project for Intercomparison of Land-surface Parameterization Schemes (PILPS) (Bowling et al., 2003-this issue). The purpose of PILPS2(e) is to quantify the accuracy with which current land schemes represent high-latitude land processes. PILPS2(e) took place over the Torne and Kalix river systems in northern Scandinavia. We participated in this project with an LSS which is part of a coupled land–atmosphere–ocean model system developed within the Swedish Regional Climate Modelling Program (SWECLIM) (Rummukainen et al., 2001). This LSS has its origin in a surface scheme used for operational weather forecasts and has been extensively improved for use in climate simulations in SWECLIM (see Section 3). For this study, we have used the LSS stand alone in one-dimensional simulations (uncoupled) forced by observations, as well as in three-dimensional simulations (coupled) with an atmospheric model.

The version of the LSS used for our first uncoupled PILPS2(e) simulations produced far too little runoff, although corresponding runs with the same LSS used in coupled simulations showed reasonable runoff. These results initiated sensitivity tests of the evaporation from snow-covered surfaces in the coupled model on the roughness length for latent heat. These tests resulted in a modification of the roughness length for latent heat, which only slightly improved the results. However, when this modification was introduced in the uncoupled version of the LSS, it resulted in a dramatic improvement of the runoff results.

3. The land-surface scheme

The land-surface scheme used for this study consists of two soil layers, which include prognostic variables for soil temperature and soil moisture. The heat capacity of the soil is modified in the temperature range $+1$ to -3 °C to simulate soil freezing (Viterbo et al., 1999). A soil temperature relaxed to ECMWF reanalysis data is used as the bottom boundary condition. The land use is described as fractions of water, forest, and open land where a bare-soil part is included in both forest and open land. Vegetation is described as a weighted combination of forest and open land vegetation characteristics. When snow is present, the top soil layer is replaced by a snow layer with a corresponding snow heat capacity. Interception of rain is parameterized similarly to Noilhan and Planton (1989). The land surface energy balance is represented by one surface temperature including all surface types. The sensible (H) and latent (λE) heat fluxes between the surface and the first atmospheric level at height z_{am} (70 m) are parameterized as

$$H = \rho c_p \frac{T_s - T_{am}}{r_a} \tag{1}$$

and

$$\lambda E = \rho c_p \lambda \frac{q_s(T_s) - q_{am}}{r_a + r_s}. \tag{2}$$

Here, ρ is air density, c_p is air heat capacity, λ is heat of vaporization, T_s is surface temperature, T_{am} is temperature at z_{am}, q_s is surface-saturated specific humidity, q_{am} is specific humidity at z_{am}, and r_a is aerodynamic resistance. The formulation of the surface resistance r_s depends on the surface; for snow and intercepted water, it is zero; for bare soil, it is formulated according to van den Hurk et al. (2000); and for vegetation according to Noilhan and Planton (1989). The aerodynamic resistance is given from the drag coefficient, C_h:

$$C_h = \frac{1}{u r_a} = \frac{k^2}{\ln(z_{am}/z_{0m})\ln(z_{am}/z_{0h})} f_h(Ri, z_{am}/z_{0h}), \tag{3}$$

where u is wind speed at z_{am}, k is the von Karman's constant, z_{0m} and z_{0h} are the roughness lengths for momentum and heat, Ri is Bulk–Richardson number, and f_h represents analytic stability expressions based on modified Louis et al. (1981) formulations.

For this study, the LSS has been used in two different versions where the only difference between them is in the roughness length for latent heat with respect to snow-covered surfaces on open land. In the first version (LSS1), the roughness length for latent heat equals that for momentum, z_{0m}, which is defined as the weighted average of the logarithmic roughness lengths for forest (1.0 m) and open land (0.2 m). In the second version (LSS2), the roughness length for latent heat for snow-covered surfaces on open land is reduced to 0.001 m, while the roughness length for latent heat for snow in forest remains at 1.0 m. These roughness lengths are applied in a parameterization where the resulting aerodynamic resistances are combined to a common aerodynamic resistance for snow, r_{a_sn}:

$$\frac{1}{r_{a\_sn}} = \frac{Fr_{opl}}{r_{a\_snopl}} + \frac{Fr_{for}}{\max(16 r_{a\_sn\ for}, 400)}. \tag{4}$$

Here, Fr_{opl} and Fr_{for} are fractional areas of open land and forest, respectively, and r_{a_snopl} and r_{a_snfor} are the aerodynamic resistances for snow in open land and forest according to Eq. (3), respectively. This parameterization is a modified version of the one suggested by Shuttleworth (1991). The factor 16 represents an increased aerodynamic resistance due to conditions between the surface and the air within the forest canopy. The aerodynamic resistance is also restricted to values above 400 to avoid too small values for unstable atmospheric surface-layer conditions. In the present LSS, there is only one surface temperature representing all different surfaces. Therefore, the sensible heat flux is also treated more simply than the latent heat flux with one single aerodynamic resistance for the whole grid square which equals that for momentum. A more detailed description of the LSS is found in Bringfelt et al. (2001).

4. Coupled three-dimensional simulations

The land-surface scheme is part of the Rossby Centre Regional Climate Model (RCA2) (Rummukainen et al., 2001). RCA2 has been developed primarily for multi-year integrations performed over the European–East Atlantic sector but has also been tested for its ability to simulate observed climate processes in other parts of the world. The atmospheric dynamics is based on a Semi-Lagrangian scheme (McDonald, 1993) and the physics package consists of a radiation scheme (Savijärvi, 1990), the CBR turbulence scheme (Cuxart et al., 2000), the Kain–Fritsch convection scheme (Kain and Fritsch, 1990), and a parameterization of resolved condensation processes following the one described in Rasch and Kristjansson, (1998). Further details regarding the model setup and the model physics are presented in Jones (2001).

The RCA2 model domain is shown in Fig. 1a. The domain is resolved by 70×60 grid points in the

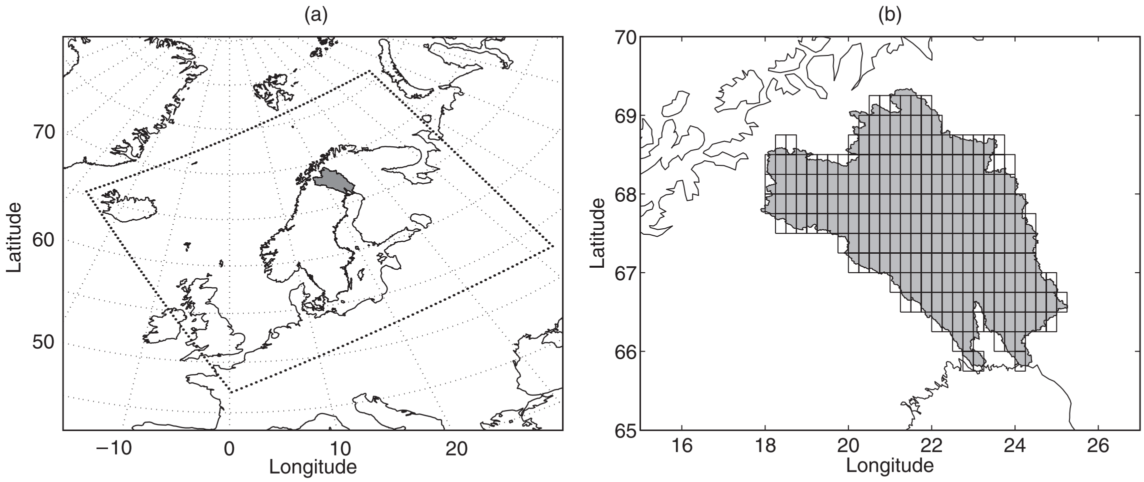

Fig. 1. (a) The model domain for the three-dimensional simulations is shown by the dotted frame and the Torne/Kalix drainage basin is the shaded area in northern Scandinavia. (b) The shaded area is the Torne/Kalix drainage basin and the grid represents that used for the uncoupled one-dimensional simulations.

horizontal, corresponding to $0.40° \times 0.40°$ resolution, and 24 levels in the vertical. The ECMWF reanalysis fields were used as forcing on lower (sea-surface and deep-soil temperatures) and lateral boundaries. The simulation was done for the period 1 September 1988–31 December 1993. Since the model system, and especially the soil column, needs some spin-up time, only the results from the period 1990–1993 were analyzed.

5. Uncoupled one-dimensional simulations

The one-dimensional simulations were performed in the framework of PILPS2e as described in Bowling et al. (2003-this issue). The Torne/Kalix river basin was divided into 218 grid points at a horizontal resolution of $0.25° \times 0.25°$, as shown in Fig. 1b. For each grid point, the following forcing data was provided with 1-h time resolution: rainfall, snowfall, 2 m air temperature, surface pressure, 2 m specific humidity, short- and long-wave downward radiation, and 10 m wind speed. Forcing data were available for the period 1979–1998, where the first 10 years were used for spin up purposes. From the RCA2 LSS driven by these data, fluxes and storage terms were calculated for the period 1989–1998. This paper concentrates on the results from the period 1990–1993.

6. Results

Fig. 2 shows simulated runoff and evapotranspiration, and simulated and observed precipitation for the Torne/Kalix river basin for the period 1990–1993. For each simulation time step, the results from all individual grid squares within the basin are averaged together and shown as cumulative sums over the 4-year period. For the coupled runs in Fig. 2a, the difference in runoff and evapotranspiration between the two LSS versions is not very large, although the tendency for smaller evapotranspiration in the LSS2 case is as expected. For the uncoupled runs in Fig. 2b, however, the difference is quite dramatic. For LSS1, the runoff is less than the evapotranspiration. The accumulated values of simulated and observed precipitation are all within the range of 250 mm at the end of the period. Therefore, the differences in runoff and evapotranspiration between coupled and uncoupled simulations cannot be explained from differences in precipitation.

The reason for the small difference in the coupled runs is that the atmospheric model is able to compensate for errors in the LSS; Fig. 3 shows the differences in temperature and saturation deficit between coupled simulations using LSS1 and LSS2, respectively. The results are monthly averages for the 4-year period. Since the change in the LSS only applies to snow-

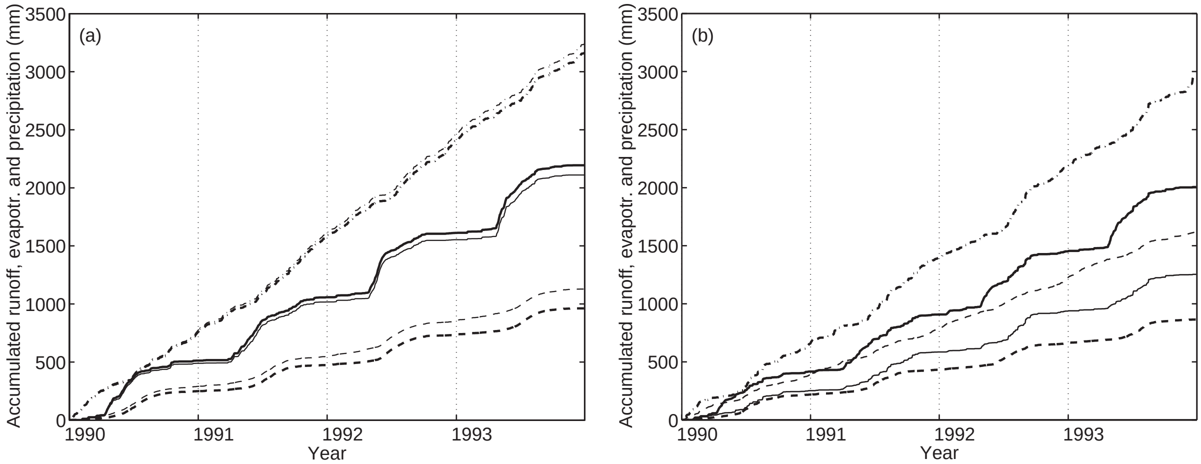

Fig. 2. Accumulated simulated runoff (solid lines) and evapotranspiration (dashed lines)in the Torne/Kalix drainage basin using the LSS in (a) a coupled three-dimensional land–atmosphere model and in (b) uncoupled simulations forced by observations at two meters height. The dash–dotted lines represent in (a) simulated and in (b) observed precipitation. The thin and thick lines show results based on LSS1 and LSS2, respectively.

covered surfaces, the largest differences should be expected during spring when most of the basin is still covered with snow at the same time as available net radiation is increasing with time. This also happens to be the case, as demonstrated by the maximum differences from March to May.

Simulations using LSS2 are slightly drier and warmer than those using LSS1 during the spring, shown as a positive difference in saturation deficit, at least for the three lowest levels. The atmospheric model increases the humidity at lower levels in the LSS1 case (as compared to the LSS2 case) in response to slightly higher evapotranspiration. The higher evapotranspiration is compensated by a slightly smaller upward sensible heat flux to keep the energy balance at the surface. This reduced sensible heat flux results in a cooling of the lower atmospheric levels. Thus, the higher upward sensible heat flux in the LSS2 case

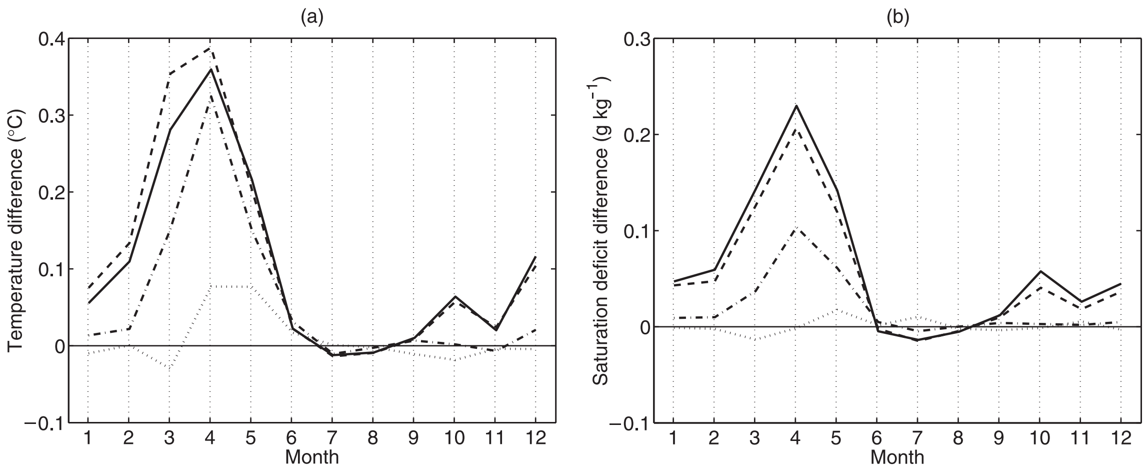

Fig. 3. Temperature (a) and saturation deficit (b) differences between three-dimensional simulations (LSS2–LSS1) for the Torne/Kalix drainage basin. The different lines represent the following atmospheric model levels above ground: 70 m (solid), 240 m (dashed), 760 m (dash–dotted), and 1100 m (dotted).

shows up as a positive difference in temperature during the spring period in Fig. 3a–b. These kind of feedback mechanisms between surface and atmosphere are described well by Jacobs and De Bruin (1992).

7. Conclusions

The use of gridded observations from the PILPS2e project as forcing data for the RCA land-surface scheme revealed deficiencies in the parameterization of latent heat flux. A too large roughness length for latent heat resulted in an overestimation of evapotranspiration, especially for evaporation from snow, and consequently in an underestimation of runoff. These errors were obvious, using observations as forcing in uncoupled simulations; whereas the same LSS used in coupled simulations with an atmospheric model gave reasonable results. The reason for this difference is that simulated temperature and humidity in the atmospheric model adjust to compensate for deficiencies in the LSS parameterization. Due to the nature of stand-alone forcing, this cannot occur in uncoupled simulations.

The adjustments needed to compensate for the errors are small compared to normal variability in temperature and humidity. Thus, these results show that it would be nearly impossible to detect this kind of error in parameterization based solely on biases in temperature and humidity from coupled simulations.

As Beljaars and Viterbo (1994) state, the magnitude of the aerodynamic resistance is important for winter evaporation. Even if the variability in observations is large, they suggest that reducing the ratio z_{0m}/z_{0h} to at least 10 is better than using $z_{0h} = z_{0m}$, which is still the case in some models. In this study, a fixed value of the roughness length for heat was used, but according to Chen et al. (1997), it would be more physical to relate the ratio z_{0m}/z_{0h} to the properties of the flow.

Using river discharge observations for validation of simulated runoff is an efficient way to check the performance of a model regarding its ability to correctly divide precipitation into evapotranspiration and runoff. The dramatic impact the change in roughness length had on uncoupled simulations in this analysis confirms this. However, as the results indicated, feedback mechanisms in a coupled model tend to compensate for deficiencies in a parameterization in relation to the sensitivity of the division of available energy between sensible and latent heat flux to aerodynamic resistance (Jacobs and De Bruin, 1992; Beljaars and Viterbo, 1994). Agreement in such validations against coupled models does not necessarily mean that the parameterizations in the model are correct, but rather that the model as such is able to simulate runoff. This provides a strong argument for continued process-by-process model validations using, wherever possible, rigorous measures such as stand-alone forcing experiments.

Acknowledgements

The authors acknowledge the PILPS2e project for providing us the data that made this study possible. We are also grateful for the valuable comments made by Stefan Gollvik and for the technical support by Anders Ullerstig. The Rossby Centre is part of the Swedish SWECLIM programme, and financed by the Swedish Foundation for Strategic Environmental Research (MISTRA) and the Swedish Meteorological and Hydrological Institute (SMHI).

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