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Using conceptual hydrological modelling to develop better sub-grid variability in the Rossby Centre Regional Atmospheric Model

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Abstract To address deficiencies in the treatment of sub-grid processes in the Rossby Centre Regional Atmospheric Model (RCA), the variability parameter approach for soil moisture from conceptual hydrological modelling was introduced to the RCA model. This improves representation of sub-grid soil moisture variability. Relationships between fraction of snow cover and snowpack volume from hydrological modelling experience were formulated and introduced to the snow accumulation and depletion calculations in the RCA model. These changes reduce the sub-grid variability of these processes to a statistical representation of the variability in each grid square. This provides a varied distribution of soil moisture and snowpack over each atmospheric model grid and produces a gradually increasing runoff response with increasing soil moisture.

Key words land surface; sub-grid variability; hydrological modelling; HBV model; atmospheric modelling; soil moisture; snow

INTRODUCTION

Under the Swedish Regional Climate Modelling Programme (SWECLIM), the Rossby Centre Regional Atmospheric Model (RCA) was developed to perform climate simulations over the Nordic region. It was recognized from the outset that improvements to the land surface scheme were needed to properly represent the hydrological cycle in this model. A large-scale hydrological model of runoff in the Baltic basin was used to compare key runoff process variables such as snow, soil moisture and evapotranspiration from RCA simulations to hydrological simulations (Graham & Bergström, 2000; Graham & Jacob, 2000). This incorporated both the runoff record and hydrological modelling experience into atmospheric model development.

Identification of deficiencies in sub-grid processes led to increased cooperation between hydrologists and meteorologists within SWECLIM to improve the surface processes of the climate model. As a result, components of hydrological modelling and modelling experience have been incorporated into the RCA model. These modifications are described in this paper.

HYDROLOGICAL MODELLING WITH HBV

The HBV model is a conceptual, semi-distributed hydrological model, which computes runoff from observations of precipitation and air temperature, typically on a daily time step (Bergström, 1995; Lindström *et al.*, 1997). Water is stored as snow, capillary water in the snowpack, soil moisture, groundwater and in lakes. The soil moisture routine originates from the simple bucket approach, but with the important additional condition that the water holding capacity of the soil in the basin has a statistical distribution representing many different buckets (Bergström & Graham, 1998). Runoff generation is based on a contributing area concept; i.e. only those buckets that have reached their field capacity will contribute to runoff in the event of rain or snowmelt. This approach implicitly accounts for the sub-basin (or sub-grid) variability in both soil water holding properties and inputs in the form of rain or snowmelt. The characteristics of HBV soil moisture and runoff were introduced into the RCA model as described below.

A simple snow parameterization scheme

For HBV modelling, sub-basins are subdivided into elevation bands and land-use classes to account for differences in climate and snow conditions. Snow conditions can vary considerably, even within an elevation zone, particularly above the timberline. In the HBV-96 model version (Lindström *et al.*, 1997), snow classes were introduced as a simple way to describe this variability. In the RCA model, no distribution of elevation occurs within grid squares. Introduction of elevation zones and snow classes into the RCA model would result in a large number of computational surfaces and thus increase computation time; simpler methods for snow distribution were needed.

Simulations with the HBV model for six basins in different parts of Sweden were used to evaluate relationships between snow cover and snow amounts as shown in Fig. 1. Entire basins were typically snow covered during snow accumulation in autumn and winter, whereas the snow-covered fraction gradually decreased during snowmelt. A clearer relationship was found between snow cover and the remaining fraction of the snowpack, than between snow cover and the current snowpack itself. This was formulated as a linearly decreasing snow covered fraction during snowmelt as shown in Fig. 2. Bare ground starts to appear as soon as the remaining snowpack reaches a threshold fraction, *sfdist*, of the maximum season snowpack, *snowmax*, i.e. the maximum reached thus far during that particular winter season. If *sfdist* is set greater than 100%, a portion of the area will remain snow free during the whole winter.

This formulation is similar to the snow cover depletion curves used by Vehviläinen (1990) and to results from Häggström (1994). Häggström compared HBV model snowpack, as a fraction of maximum snowpack, to snow cover data based on NOAA-AVHRR analyses. By assuming a rectangular distribution of snow at maximum accumulation, *sfdist* can be related to the coefficient of variation, *CV*, of the snowpack by:

$$sfdist = 0.5 \cdot CV\sqrt{12} \qquad (1)$$

Table 1 shows estimates of *CV* for snow distributions and *sfdist* values derived from different sources. Typical values of *sfdist* should be greater than 0.5 for mountainous areas and 0.2–0.6 for forested areas.

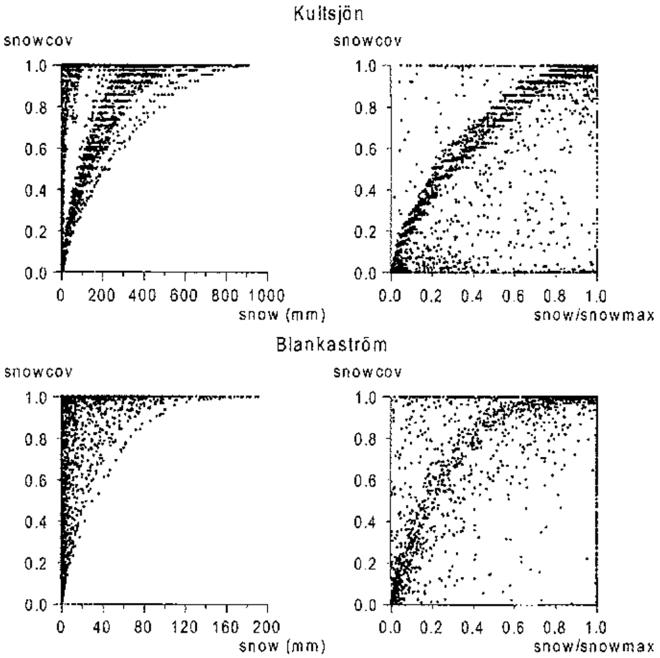

Fig. 1 Snow-covered fraction of the Kultsjön (northern Sweden) and Blankaström (southern Sweden) basins *vs* current snowpack (left), and current snowpack as a fraction of maximum season snowpack (right), from results by the HBV model.

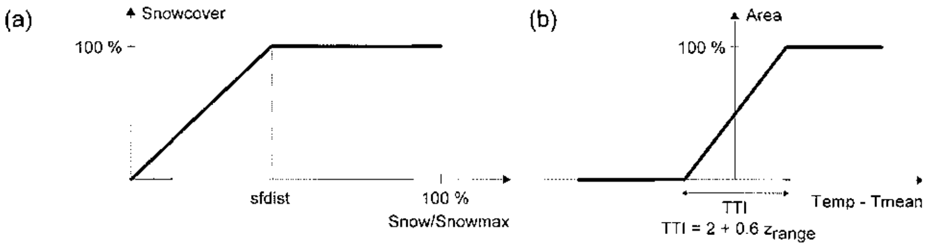

Fig. 2 Modelled snow cover as a function of (a) snow amount and (b) modelled temperature distribution over an area.

Table 1 Estimates of *CV* for snow distributions and derived values for *sfdist*.

Area	*CV*	*Sfdist*
Bare mountain*	0.3–0.7	0.5–>1
Mountainous basins [†]	-	0.5–0.8
Mountainous basins [‡]	0.5–0.9	0.8–>1
Mountainous basins [§]	-	~0.8
Other open areas*	0.09–0.19	0.2–0.3
Forest*	0.12–0.22	0.2–0.4
Forest [‡]	0.34	0.6
Forested basins [§]	-	~0.6

* Gottschalk & Jutman (1979); [†] Häggström (1994); [‡] Marchand & Killingtveit (1999); [§] HBV model simulations.

With a temperature lapse rate of 0.6°C per 100 m (as commonly used in the HBV model), a temperature interval, *TTI*, was assumed over a grid square as a function of the elevation range, *zrange*, as shown in Fig. 2. For simplicity, a rectangular elevation distribution was assumed based on the mean and standard deviation of the elevation in each grid square. The temperature can then be distributed evenly between $Tmean - TTI/2$ and $Tmean + TTI/2$. The lowest *TTI* value of 2°C was based on experience to account for temperature variations during the day and between different parts of an area even without elevation differences.

Precipitation was assumed to fall as rain over the part of the area, *Amelt*, where the temperature, *Temp*, is above zero, and as snow over the rest of the area ($1 - Amelt$). Snowmelt, *melt*, was assumed to take place with a rate of:

$$melt = cfmax \cdot Tpos \cdot snowcov \cdot Amelt \qquad (2)$$

where *Tpos* is the average temperature for the area with snowmelt (*Amelt*). The snow melt rate, *cfmax*, is the well-known degree-day factor commonly used in hydrological modelling. Typical values of *cfmax* from experience are 2 mm per °C day for forested areas and 3.5 for open areas.

HBV test of the simple snow parameterization scheme

Results of the snow scheme tested in the HBV model for the 1095 km^2 Kultsjön basin in northern Sweden are shown in Fig. 3. The parameter *Sfdist* was set to a standard value of 0.6. Automatic calibration against a mean of the Nash-Sutcliffe efficiency criterion, R^2, for discharge and snow cover resulted in R^2 values of 0.86 and 0.97, respectively.

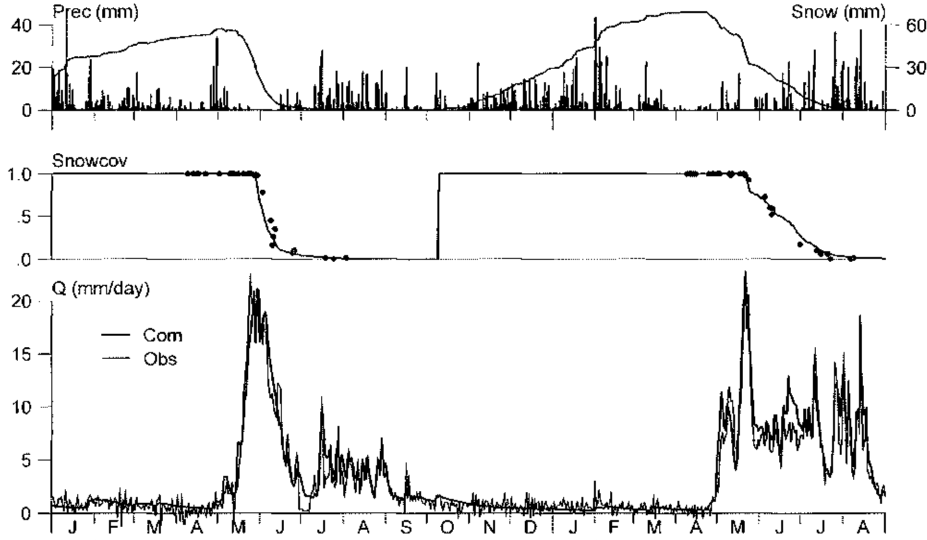

Fig. 3 HBV model application for Kultsjön; thick lines show computed values and thin lines show observations. The observed snow cover data, shown as dots, are from analysis of NOAA-AVHRR images (after Häggström, 1994).

As atmospheric models use a much shorter time step than the typical daily step of the HBV model, the influence of using shorter time steps in the degree-day method was studied by running the snowmelt model with both 12- and 24-h time steps in five basins. Good agreement in computed snow amount was achieved between these two time steps. Slight improvement was shown by reducing the snowmelt rate in the 12-h application by 10–20%. As a first approximation, the degree-day model could therefore be used at finer time resolution without major modifications.

ATMOSPHERIC MODELLING WITH RCA

The Rossby Centre Regional Atmospheric Climate Model (RCA) is based on the international High Resolution Limited Area Modelling (HIRLAM) forecast model (Källén, 1996). Many of the parameterization schemes of HIRLAM are retained in the RCA model, but changes in the surface treatment were necessary to enable climate integrations. The RCA model also differs from HIRLAM in that it includes an interactive treatment of regional water bodies—the Baltic Sea and Nordic lakes—by including a regional ocean model and a lake model. The RCA model has 19 vertical levels and has thus far been applied with horizontal grid resolutions of 0.8° (~88 km), 0.4° (~44 km) and 0.2° (~22 km). Two GCMs have been used to provide lateral boundary driving conditions along the perimeter of this domain. These are the UKMO HadCM2 GCM of the Hadley Centre (UK) and the ECHAM4/OPYC3 GCM of the Max-Planck-Institute for Meteorology (Germany). More detail is given by Rummukainen *et al.* (2001).

The task of a land surface scheme is to calculate time-tendency forecasts of surface variables (e.g. soil temperature and soil moisture) while fulfilling both the energy and water balances at the land/atmosphere interface. The main improvements to the RCA surface scheme (from the original HIRLAM model) are in the hydrological cycle and treatment of evapotranspiration processes. Parameterization of soil temperature is much the same as in the HIRLAM model, but with a modification for frozen soil moisture (not discussed here).

Soil water processes and runoff in RCA

The land surface as parameterized in the RCA model is shown in Fig. 4. Soil moisture is represented by two soil layers, a shallow surface layer, w_s, and a deeper layer, w_d. The HBV model approach to soil moisture and runoff was incorporated into these layers. Snow depth, *SN*, accumulates as a single layer on top of the uppermost soil layer. The parameters shown in the figure are listed in Table 2.

The RCA model uses modifications of the HBV model equations to give the fraction of rainfall plus snowmelt used for drainage flow, Q_1, to the second soil layer and for final or total runoff from the deep layer, Q_2, as follows:

$$Q_1 = (RA_F + SN_M) \cdot \left(\frac{w_s}{w_{FC1}} \right)^{\beta_1} \tag{3}$$

$$Q_2 = Q_1 \cdot \left(\frac{w_d}{w_{FC2}} \right)^{\beta_2} \tag{4}$$

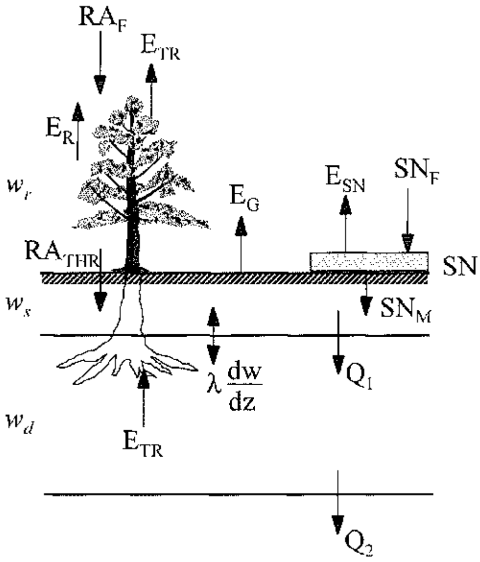

Fig. 4 The soil water and vegetation processes in RCA.

Table 2 The soil water and vegetation parameters in RCA.

Parameter	Description
SN	snow depth (kg m^{-2})
SN_F	snowfall rate (kg m^{-2} s^{-1})
RA_F	rainfall rate (kg m^{-2} s^{-1})
RA_{THR}	throughfall rate (kg m^{-2} s^{-1})
SN_M	snowmelt rate (kg m^{-2} s^{-1})
E_{TR}	transpiration from vegetation canopy (kg m^{-2} s^{-1})
E_R	evaporation of liquid water intercepted on the canopy (kg m^{-2} s^{-1})
E_{SN}	evaporation from snow cover (kg m^{-2} s^{-1})
E_G	evaporation from bare soil (kg m^{-2} s^{-1})
w_s	soil moisture in the surface soil layer (kg m^{-2})
w_d	soil moisture in the deep soil layer (kg m^{-2})
w_r	canopy water (kg m^{-2})
w_{FC1}	model field capacity for surface soil layer (20 mm)
w_{FC2}	model field capacity for deep soil layer (220 mm)
Q_1	drainage from the surface soil layer to the deep soil layer (kg s^{-1})
Q_2	total runoff generation (occurs from the deep layer) (kg s^{-1})
λ	Hydraulic conductivity (m s^{-1})
veg	fraction of surface covered with vegetation (%)
β_1	index of heterogeneity for the surface layer
β_2	index of heterogeneity for the deep layer

The exponents β_1 and β_2 reflect the areal variability of soil moisture. Thus, in contrast to many climate models, runoff can occur even if the area averages w_s and w_d are below field capacity, w_{FC}. These equations provide a much stronger relationship between runoff and soil moisture than those in the original HIRLAM model. As in nature, runoff response to infiltration increases (decreases) nonlinearly as soil moisture increases (decreases). The diffusive flux–gradient relationship for soil moisture, $\lambda \mathrm{d}w/\mathrm{d}z$, was retained to allow for more effective drainage of soil water as the top layer dries out.

Evapotranspiration in RCA

Evaporation and transpiration are treated explicitly in the RCA model using sub-grid averaging of vegetation parameters (fraction of forest/open land). This is a considerable change from the original HIRLAM model, where evaporation occurred only from the surface (bare soil). Transpiration, E_{TR}, from dry vegetation uses a formulation for surface resistance, r_s, based on the ISBA (Interaction between Soil, Biosphere and Atmosphere) scheme (Mahfouf *et al.*, 1995; Noilhan & Planton, 1989). This depends on photosynthetically active radiation (short wave), air water vapour pressure (forest only), air temperature, soil water stress and occurrence of frozen soil water as follows:

$$E_{TR} = \rho L \frac{veg \cdot \Delta q}{r_s + r_a} \tag{5}$$

$$\Delta q = q_{\mathrm{sat}}(T_S) - q_a \tag{6}$$

where $q_{\mathrm{sat}}(T_S)$ is the saturation specific humidity at surface temperature, q_a is the specific humidity at the lowest model level, L is the latent heat of vaporization, ρ is the density of air, and r_a is the aerodynamic resistance between the lowest model level and the surface.

In summer, the evaporation of rainfall intercepted on forest vegetation can amount to some 25% of total rainfall. The remaining throughfall to the ground contributes to the soil water budget. For forest and open land, the RCA model uses a formulation of rainfall interception similar to that used in the ISBA scheme. Evaporation from intercepted rain, E_R, is calculated as a function of water stored on the canopy, w_r, as follows:

$$E_R = \rho L \frac{veg \cdot \Delta q}{r_a} \cdot \left(\frac{w_r}{w_{r\max}} \right)^{2/3} \tag{7}$$

where $w_{r\max}$ is set according to vegetation type and w_r is determined from initial canopy water plus rainfall.

Although not described in detail here, evaporation also occurs from bare ground and from the snow surface as shown in Fig. 4. Transpiration from a saturated canopy is also included, which is a further refinement over the ISBA scheme.

Snow processes in RCA

With some modification, the snow relationships from hydrological modelling as described above are included in the RCA model. They provide a statistical sub-grid treatment of the fractional area of snow cover and snowmelt rate using topographical variance. Maximum snow depth is a new forecast variable that is used to calculate distribution of the snow cover.

SUMMARY AND CONCLUSIONS

Co-operation between hydrological and meteorological modellers has led to further development of the RCA climate model. The HBV variability parameter approach introduced to the RCA model provides a simple yet more realistic runoff response to soil moisture. This is complemented by the variable distribution of snowmelt from hydrological modelling. Furthermore, evaporation and transpiration are now explicitly represented.

This work is in progress and only short period, limited tests have thus far been made for the climate model. Longer integrations using re-analysis data are currently underway. The authors do not expect these changes to overcome all the problems of sub-grid variability in the land/atmosphere interface. However, we anticipate that we have created a suitable platform from which we can continue to improve representation of these processes in the RCA model.

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