Physically based modelling of snow cover dynamics in Alpine regions

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Abstract

Modelling the amount, spatial variability and temporal changes of snow water storage in Alpine headwatersheds helps to quantify the available water resources and to estimate the timing of their entrainment. Therefore it is necessary to understand the processes leading to the observed spatial patterns of an Alpine snow cover. We present the physically based snow component of the integrated Global Change Decision Support System DANUBIA, which has been developed in the framework of the GLOWA-Danube project (www.glowa-danube.de). With the snow component, the energy balance, the water equivalent, the liquid water storage and the melt rate of the snow cover is simulated. Wind-induced snow transport is considered using parameterised results of an explicit modelling at the process scale.

Keywords: climate change, lateral snow transport processes, liquid water content, snow cover modelling

1 Introduction

In high mountain regions a significant portion of precipitation is temporally stored as snow and later released as snowmelt. Snowmelt is an important component of water supply for the downstream population of large mountain-foreland river systems. Modelling the amount, the spatial variability and temporal changes of the snow water storage in the headwatersheds helps to quantify the available water resources and to estimate the timing of their entrainment. It is therefore necessary to understand the processes leading to the specific spatial heterogeneity of an Alpine snow cover (Pomeroy 1998, Liston 2004). This article describes the physically based snow component of the Global Change Decision Support System (DSS) DANUBIA. For the Upper Danube catchment, simulated snow cover dynamics of historical periods as well as of future climate change scenarios are presented. In the outline we describe the implementation of a glacier model and the consideration of lateral snow transport processes.

2 The DANUBIA snow component

2.1 Energy balance of the snow surface

For the computation of the energy balance, melting condition (snow surface temperature ≥ 273.16 K) and no melt (snow surface temperature < 273.16 K) are distinguished first. Snow temperature is substituted by air temperature as reliable data is lacking. In the first case, a snow surface temperature of 273.16 K is assumed and melt can occur. If air temperature is < 273.16 K, an iterative procedure to adopt the snow surface temperature for closing the energy balance is applied. The energy balance for a snow pack can be expressed as:

$$Q + H + E + M + A + B = 0 \tag{1}$$

where Q is the shortwave and longwave radiation balance, H the sensible heat flux, E the latent heat flux, M the energy potentially available for melt, A the advective energy supplied by solid or liquid precipitation, and B the soil heat flux for the current time step. All energy flux densities are expressed in $W \cdot m^{-2}$.

The amount of shortwave radiation absorbed by the snow surface is determined by the albedo (Rohrer 1992, U.S. Army Corps of Engineers 1956):

$$a = a_{min} + a_{add} \cdot e^{-kn} \tag{2}$$

where a_{min} is the minimum albedo of (old) snow, a_{add} is an additive albedo (with $a_{min} + a_{add}$ representing the maximum snow albedo), k is a recession factor depending on air temperature (which determines snow surface temperature) and n the number of days since the last considerable snowfall (i.e., at least $0.5\ mm \cdot h^{-1}$); each time such snowfall occurs the snow albedo is reset to its maximum value.

Longwave emission of the snowcover $Q_{l\uparrow}$ is calculated with snow emissivity ε and the Stefan-Boltzmann-constant σ:

$$Q_{l\uparrow} = -\sigma \cdot \varepsilon \cdot T_s^{\,4} \tag{3}$$

where T_s is the snow surface temperature.

The turbulent fluxes are modelled over empirical descriptions valid for medium roughness and a wide range of wind speeds. The parameterisations for the turbulent fluxes used here have been proposed by Kuchment & Gelfan (1996) for a wide range of neutral or stable conditions. Thereby, the sensible heat flux H is expressed with wind speed W in $m \cdot s^{-1}$ as

$$E = 18.85 \cdot (0.18 + 0.098 \cdot W) \cdot (T - T_s) \tag{4}$$

and the latent heat flux E is calculated as

$$E = 32.82 \cdot (0.18 + 0.098 \cdot W) \cdot (e_l - e_s) \tag{5}$$

where W is the measured wind speed ($m \cdot s^{-1}$), e_1 is the water vapour partial pressure at measurement level and e_s the water vapour saturation pressure at the snow surface, with both water vapour pressures being calculated using the Magnus formula and expressed in hPa.

The small mass changes δe in mm generated by sublimation or resublimation are simulated with t being the duration between two model time steps (3600 s):

$$\delta e = \frac{E \cdot t}{l_s} \tag{6}$$

where l_s is the sublimation/resublimation heat of snow.

The advective energy A supplied by precipitation P depends on its phase. A threshold wet-bulb temperature T_w of 2°C is assumed for the distinction between snow and rain; the wet-bulb temperature is iteratively determined by solving the psychrometer formula. Then, the energy advected by P in mm is calculated for rainfall on snow with

$$A = P \cdot c_{sw} \cdot (T - 273.16) \tag{7}$$

where c_{sw} is the specific heat of water. For snowfall, the advective energy is computed with

$$A = P \cdot c_{ss} \cdot (T_w - T_s) \tag{8}$$

where c_{ss} is the specific heat of snow.

For the case of melting condition (air temperature ≥ 273.16 K), all fluxes are calculated with an assumed snow surface temperature of 273.16 K. If energy remains available for melt, its amount in mm is calculated with

$$melt = \frac{M \cdot t}{c_i} \tag{9}$$

where c_i is the melting heat of ice.

In the case of no melting condition, an iterative scheme to close the energy balance by adopting the snow surface temperature and recalculation of the respective fluxes is applied.

2.2 Liquid water content of the snow pack

Generally, the liquid water storage is assumed to be a fractional volume inside the snowpack. It can be filled by either melt water or rain. The content of the liquid water storage is considered in the mass balance of the snow pack. No water leaves the snow pack until a threshold determined as a fraction of the snow water equivalent is reached. Potential rain surplus is assumed to directly drain from the snowpack, whereas melt is then increased by a corresponding amount of formerly stored liquid water. If liquid water is stored in the snow pack and the air temperature is lower than 273.16 K, part of the actual liquid water storage can refreeze, and the snow surface temperature is adopted by iteratively closing the energy balance (Prasch et al. 2008).

2.3 Validation of model results

The modelled temporal evolution of the snow cover is validated at several weather stations with different physiogeographic characteristics in the Upper Danube catchment, whereas the modelled snow cover extent at a particular date is compared with corresponding snow cover maps derived from satellite data.

2.3.1 Point scale

The stations for the point scale validation in the Upper Danube catchment are Kühroint (1,407 m a.s.l.) and Grainet-Rehberg (655 m a.s.l.). Model results for different winter seasons are illustrated in figure 1, including measured snow water equivalent. Resulting Nash-Sutcliffe efficiency (Nash & Sutcliffe 1970) amounts to 0.97 for Kühroint (2005/06) and 0.98 for Grainet-Rehberg (1978/79).

2.3.2 Catchment Scale

The modelled spatially distributed snow cover for the Upper Danube catchment is validated using NOAA/AVHRR derived snow cover maps (Bach et al. 2004, Appel et al. 2006) of the winter season of 2005/2006 (figure 2). Modelled and remotely sensed snow covered area are similar, 32% of the catchment is classified, whereas

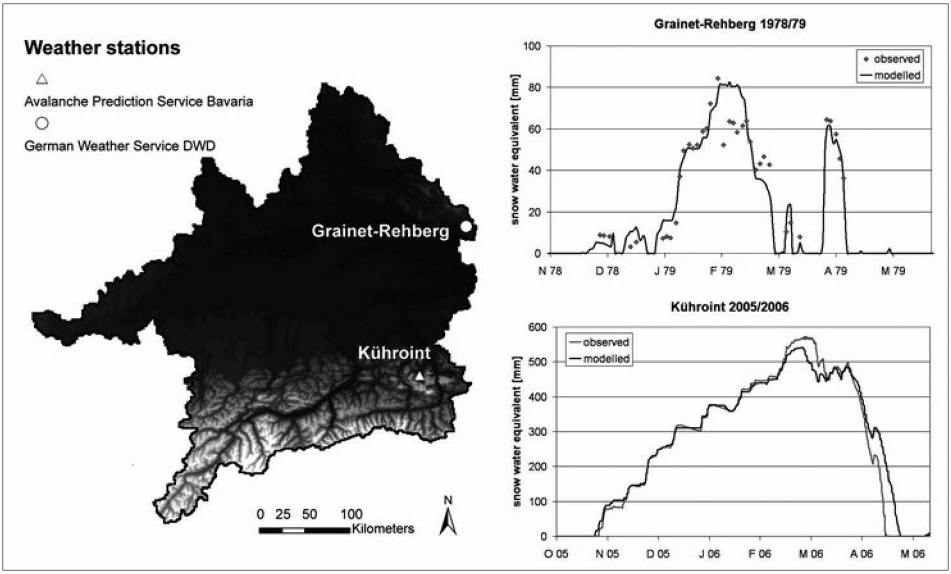

Figure 1: Location of the weather stations Grainet-Rehberg and Kühroint in the Upper Danube catchment and course of measured and modelled snow water equivalent.

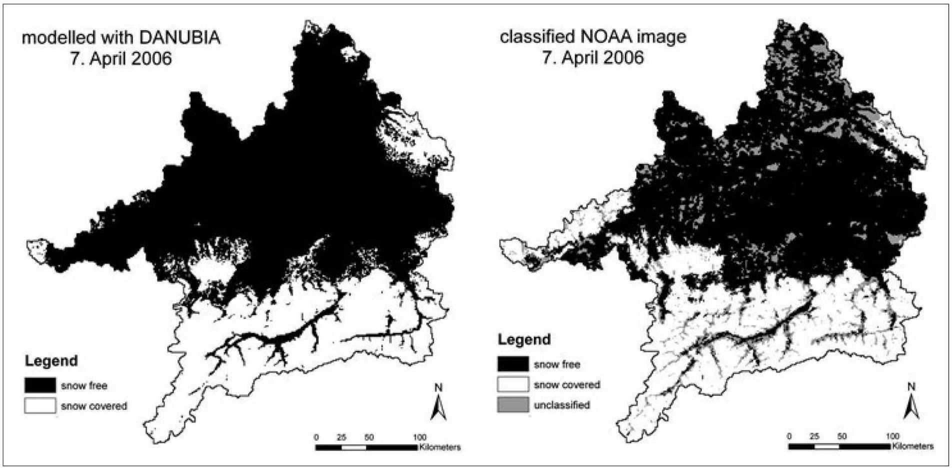

Figure 2: Modelled and satellite derived snow cover for the Upper Danube catchment on April 7, 2006.

33% is modelled as snow covered. Shadow, clouds and fog constrain 10% of the image so that the snow free area varies between 57% (NOAA/AVHRR) and 67% (DANUBIA). As the constrained areas are spread over the image, the spatial pattern of the snow cover is reproduced well.

3 Effect of future climate

The impact of future climate change on modelled snow cover is analysed in comparing the mean evolution of the snow water equivalent for the future period 2041–2070 with the past period 1971–2000 (figure 3). A stochastic weather generator provides future climate data for an IPCC B2 scenario (Mauser et al. 2007). For Kühroint the simulation shows a future decrease of water quantity stored as snow. Furthermore the snow cover melt out is shifted from June to the end of May. In Grainet-Rehberg a similar effect appears in the future period with a higher reduction of snow water equivalent.

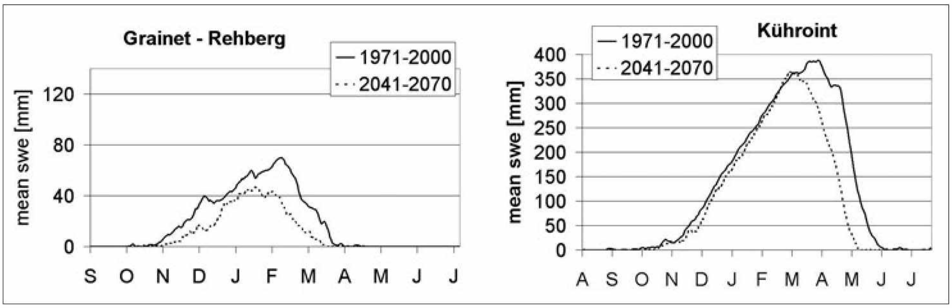

Figure 3: Evolution of mean snow water equivalent for 1971–2000 and the future period of 2041–2070.

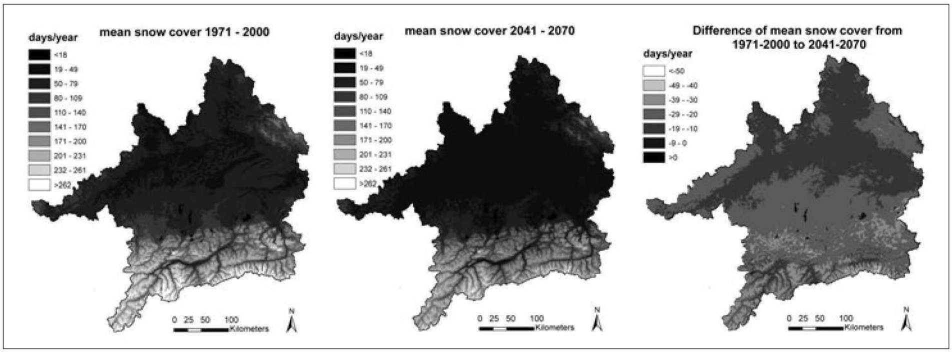

Figure 4: Mean annual snow cover in the Upper Danube catchment for the periods 1971–2000 and 2041–2070.

The comparison of the duration of the annual snow cover for the Upper Danube catchment shows the following: historical mean snow cover duration varies between 34 days per year in low valley regions, and more than 300 days per year in the Alps from 1971 to 2000. For our 2041–2070 scenario, mean annual duration of the snow cover is reduced in all parts of the catchment: DANUBIA model results indicate a decrease not only in the lowlands, but also in the high mountain regions. With a reduction of more than 40 days per year in the Alps as well as in the Alpine foreland the effect reaches a maximum, whereas in the river valleys the duration of the snow cover is reduced only by twenty days (figure 4).

4 Discussion and outlook

The modelled snow water equivalent at the point scale as well as at the catchment scale with the physically based snow component of DANUBIA, simulating the energy balance of the snow pack, shows high similarities to observed data. As an effect of potential future climate change, reduced annual snow cover duration was modelled at the weather stations Kühroint and Grainet-Rehberg as well as at the catchment scale. Due to the temperature increase the precipitation phase (rain or snow) in winter changes and causes less snowfall in the future although the amount of precipitation in winter almost stays the same in the generated scenario (Mauser et al. 2007). As less water is stored as snow, the melt out is reached earlier. Since this effect appears in regions with temperatures around the freezing point, the decrease in higher mountain regions with temperatures mostly below the freezing point during winter is lower than in low mountain ranges. This impact of climate change will modify the temporal as well as the quantitative dynamics of the flow regimes of the rivers.

For our future model developments, we want to include a numerical description of glacier response on climate change. For that purpose DANUBIA will be extended with a module simulating subscale glacier advance and retreat utilising a glacier inventory providing geometry and flow characteristics for each glacier. Hence, the individual dynamics of any single glacier and the effect of the reduction of the

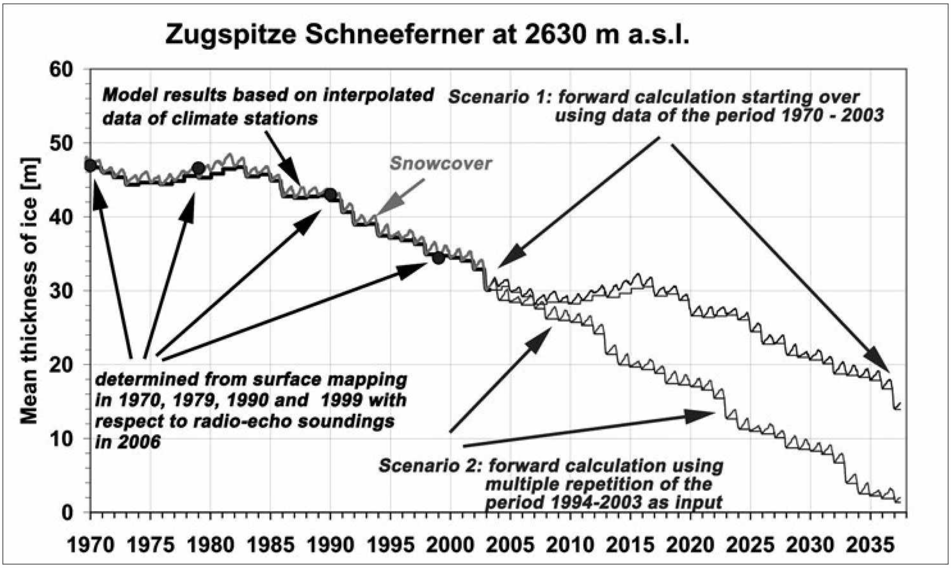

Figure 5: The evolution of mean thickness of ice and the snow cover for the Schneeferner at the Zugspitze in Germany in the past and for two future scenarios.

ice resources in the Alpine headwatersheds can be investigated. First results of the model are shown in figure 5 (Weber et al. 2007).

A prerequisite for a proper consideration of subscale snow accumulation and ablation areas is an adequate representation of lateral snow transport processes, namely wind-induced and gravitationally. If these two processes are neglected, the models tend to overestimate snow depth in the highest regions where temperature is always low. However, explicit simulation of snow transport processes in coarse scale models (> 1 km^2) is complicated due to:

- interpolated or statistically derived wind fields will not produce correct snow transport rates in Alpine terrain (Bernhardt et al. 2008), nor do they allow to describe the correct erosion and ablation zones (Liston 2004);
- both simulated snow drift as well as gravitational transport rates strongly depend on the scale of the DEM (Sturm 2004, Gruber 2007).

Hence, simulations were computed at a much finer scale (30 m and 50 m respectively) and the results were related to different topographic and meteorological parameters. The scheme is relatively transferable since the test site can be considered as typical Alpine and the topographic and meteorological parameters are mostly available.

For the modelling of wind-induced snow transport processes we used SnowTran-3D (Liston & Sturm 1998) in combination with MM5 predicted wind fields (Bernhardt et al. 2008). With respect to computational performance, a library of 220 wind fields was separately simulated, representing the most important wind situations for the test site. German weather service Lokalmodell re-analysis data was used to synchronise actual transport modelling with the best-fitting library file (Bernhardt et al. 2008).

For the description of gravitational snow transport we used a mass-conserving, multiple direction flow propagation algorithm (Gruber 2007) enabling a physically based modelling of snow transport from steep slopes to their base, according to what can be visually observed.

The long-term effect of these two processes of lateral snow transport and the general increase of snowfall with altitude are responsible for the temporal permanence of snow depth distribution in a region (Sturm 2004). A correlation of snow depth and transport rates with topographical features (altitude [m a.s.l.], curvature [θ], slope [°], aspect [°], mean wind speed [m/sec] and mean wind direction [°]) at the process model scale enables the transfer of these parameters to a coarser scale. Any specific combination of topography and wind speed/direction generates a specific snow depth distribution per pixel from which the snow-free fraction of any pixel can be derived (Liston 2004). As a result, the energy and moisture fluxes of both the snow-free and snow-covered fraction of the pixel area can be predicted.

The presented scheme is not locally calibrated and thus generally transferable. Within the framework of the EU-funded *Brahmatwinn* research project (www.brahmatwinn.uni-jena.de), DANUBIA will be applied for regional hydrological modelling in the Himalayan catchment of the Upper Brahmaputra. There, we focus on the effect of snow and glacier melt on the spatial variation and temporal dynamics of runoff generation in the Upper Brahmaputra watershed considering scenarios of future climate change up to the IPCC horizon (2001). Respective meteorological data will be provided by means of climate modelling (Marke & Mauser 2008).

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