SNOW PACK SIMULATION IN THE SWISS ALPS – COMBINING GIS AND REMOTE SENSING TO MODEL SNOW COVER IN SWITZERLAND

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ABSTRACT

A simple model for snow pack simulation on the alpine scale is presented. Such a model can be of interest for natural risk management (avalanches, debris flows, flooding), hydropower generation or tourism.

The model allows distributed modelling at the spatial resolution of 1 km by combining snow maps derived from NOAA/AVHRR data with point information from all available snow gauges for validation. The computation of the snowmelt is based on the daily mean air temperature and the daily sum of precipitation. The interpolation of these two driving variables for the 1 km resolution grid is crucial for the quality of the simulation results.

First focus is testing the interpolation method in order to consider the relationship between temperature or – more uncertain – precipitation and altitude. For that, an ordinary kriging with detrending is compared to the up to now implemented method of 1st order inverse distance weighting. First results of simulating the snow cover with standard parameters for one simulation period of nine months are presented and discussed. Validation at the snow gauges on snow depth and snow water equivalent (SWE) show promising results. The comparison of the snow maps as derived by NOAA/AVHRR allow a first conclusion of the snow model generally overestimating the snow covered area.

INTRODUCTION

A snow model, developed for estimating the amount of melt in spring 1999 after heavy snowfalls in the alpine regions, is presented. A model simulating the snow pack on the alpine scale is of great interest. It can contribute to natural risk management (avalanches, debris flows, flooding), hydropower generation or tourism. Studies on climate change also have to take into account the influence of snow on the climate system. The high albedo and the large amount of energy consumed by the snow pack during the melting season affects environmental conditions up to continental scale (1).

The model - applied in the relatively small and homogeneous but still mountainous area of the Canton of Berne - is now tested for the whole of Switzerland. Due to its simplicity and the fact that only daily values of precipitation and air temperature as measured by the standard meteorological stations of MeteoSwiss are needed as input factors, the snow model could be applied to any region within the Swiss Alps by any interested users.

METHODS

As a distributed model the snow model accounts for the spatial distribution of the snow cover. It works at a spatial resolution of one kilometre, which on the one hand is predefined by NOAA/AVHRR's pixel size. On the other hand it can be seen as a middle course between the very high resolved models in small scale studies needing many more detailed input parameters such as information about wind speed and direction and the models on coarse resolutions at continental scales that do not take into account the complex topography of the Alps.

The model is built out of five modules being able to interchange the variables. These modules are the following: "Precipitation" (separation of snow and rain), "accumulation" (increase of SWE and snow depth), "metamorphosis" (decrease of snow depth depending on time and snow density), "cold content" (snow energy balance / calculation of potential and effective snow melt) and "mass balance" (amount of liquid water to be melted). The modules, their tasks and the variables are illustrated in Figure 1, together with the parameters that can be used for the calibration of the model. Detailed information on the snow model's architecture can be found at Kleindienst (1). The module "cold content" carries out an important role since it is responsible for the calculation of the energy balance. Based on the degree-day factor in combination with a radiation correction it is called an extended temperature-index model and represents a simplification of the energy balance. This kind of model has been used by several authors [(2), (3), (4)] and in comparison to physically based snow cover models showed good performance.

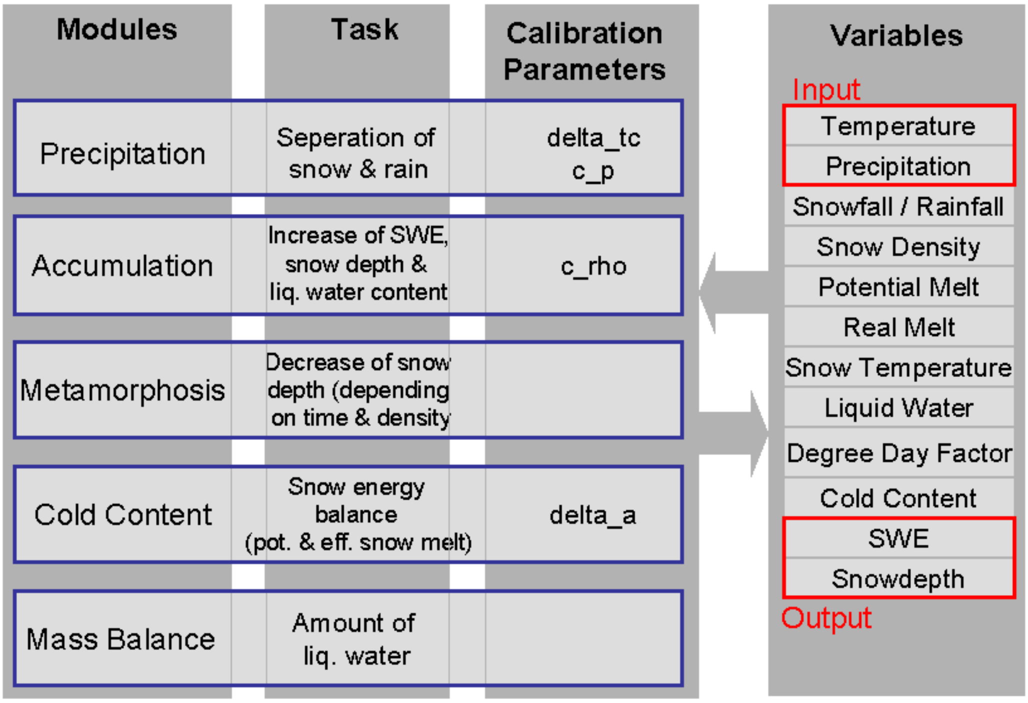

Figure 1: The five modules of the snow model that are responsible for the calculation of the snow pack and their tasks. The four calibration parameters can be used for calibrating the model. The variables are interchanged between the modules.

The snow model itself only needs two driving variables as input for the computation of the snow cover on daily intervals. These are the daily sum of precipitation and the daily mean of temperature as measured at the meteorological stations of MeteoSwiss. In Figure 2 it can be seen that snow depth and SWE from the measurement net of the Swiss Federal Institute for Snow and Avalanche Research Davos (SLF) as well as NOAA/AVHRR data are needed not as direct inputs into the model but for calibration and validation done interactively by the user. Processing of NOAA/AVHRR data is based on the algorithm for snow and ice detection as first presented by Gesell (5). The processing chain used within this work is described more closely by Wunderle (6).

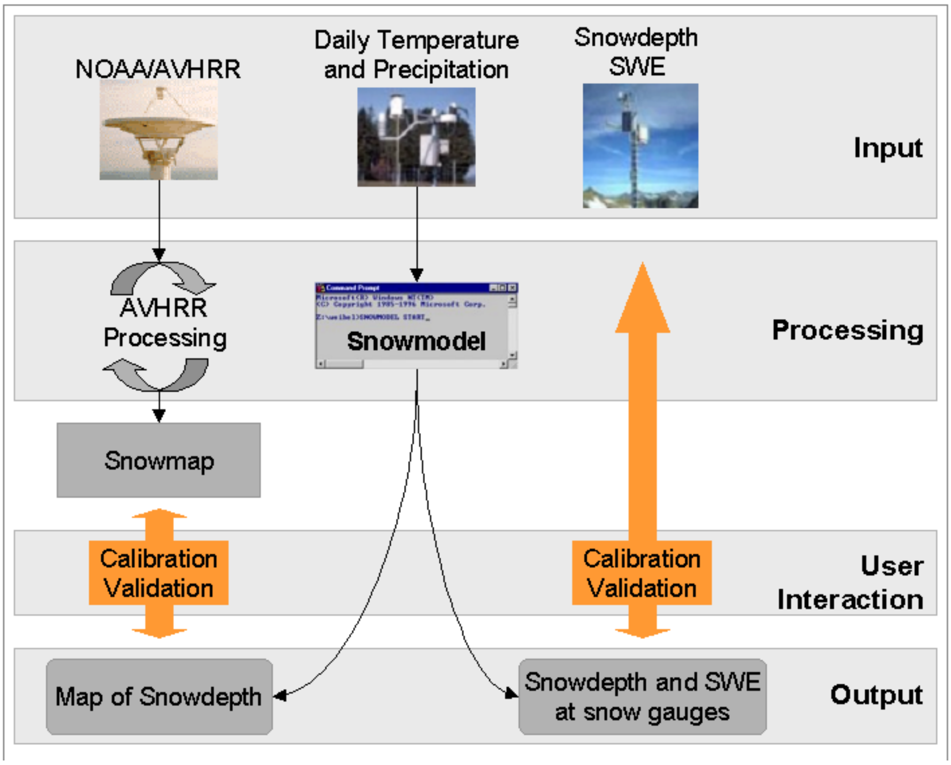

Figure 2: Overview of the functionality of the snow model

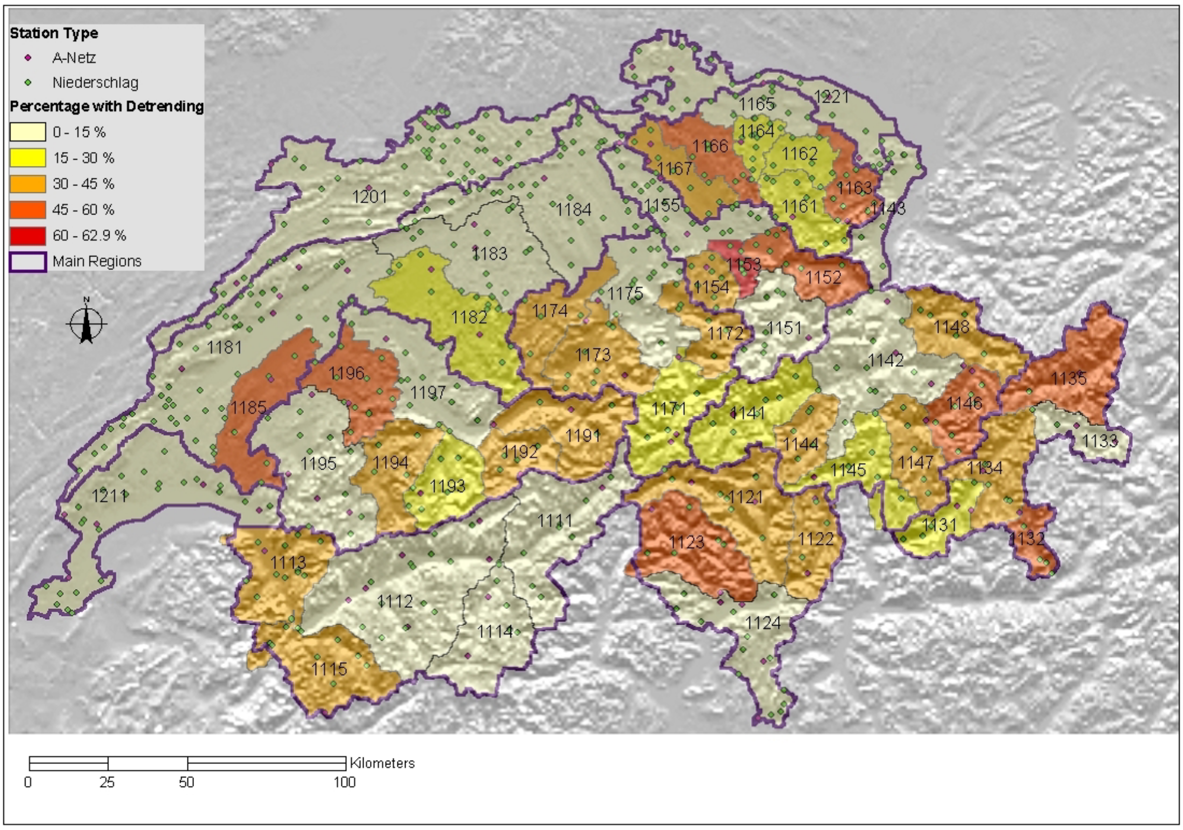

Figure 3: Percentage of days with $R^2 \geq 0.5$ (only rainy days included). Each of the 53 watersheds is labelled with a number

First, attention is paid to the interpolation of precipitation. It is tested whether the quite simple but rather fast interpolation method of inverse distance weighting (IDW) can be substituted by an ordinary kriging and/or be improved by regarding the precipitation-elevation relationship. Before interpolating a detrending of the observed values to a reference elevation with the application of a precipitation gradient might give some better results of interpolation. Various authors use kriging and kriging with detrending to interpolate precipitation in mountainous terrain (e.g. by (7), (8), (9)) but it is used mostly on climatic time scales and not on daily intervals. IDW weights the observed values inverse to the distance between each data point and the point that is estimated. Ordinary kriging on the other hand analyses the data applying geostatistics and weights the samples dependent on the spatial correlation between each sample and the estimated point. Therefore the semi-variogram that draws the semi-variance $\gamma(h)$ between the samples in function of the distance h is used. The semi-variance $\hat{\gamma}(h)$ of a sample with data points at x_i is estimated as follows:

$$\hat{\gamma}(h) = 1/2\,n\sum_{i=1}^{n}\{z(x_i) - z(x_i + h)\}^2$$

Further readings on geostatistical methods can be found at different authors [(10), (11) or (12)].

For the analysis in this work, gstat (13), an open source software tool for geostatistical applications is integrated into the development environment. Ordinary kriging is done with a search radius of 20 km and a distance lag of 5 km. For detrending, the total area of Switzerland is divided into 53 watersheds (Figure 2) within which the daily precipitation gradient is calculated based on observed precipitation values. The methods are tested during a typical simulation period starting on Oct 1 and ending on June 30 of winter 1998/1999.

Only about 30% of the whole dataset show a correlation coefficient R^2 of 0.5, after Sevruk and Mieglitz (14) this being the minimal correlation for explaining the distribution of precipitation values by the influence of altitude. Figure 3 shows the percentage of days within the watersheds where $R^2 \geq 0.5$. The total of 100% of the dataset only includes days with observed precipitation sum greater than 0.1 mm.

Table 1: Mean and mean standard deviation of the daily absolute mean errors (No: no detrending, yes: with detrending) of the two interpolation methods

	IDW (no)	Kriging (no)	IDW (yes)	Kriging (yes)
Mean absolute error	2.2 mm	1.8 mm	2.7 mm	2.4 mm
Mean standard deviation	2.7 mm	2.5 mm	3.2 mm	2.8 mm

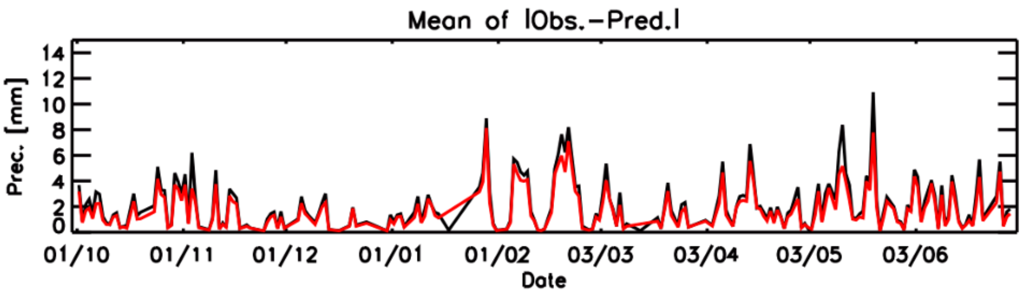

Figure 4: Comparison of the average of daily absolute errors of IDW (black) and ordinary kriging (red) without detrending

Table 1 lists some of the statistics done on the daily absolute average error ($\overline{|obs. - sim|}$) and allows comparing the two interpolation methods with or without detrending. Although cross-validation shows that kriging estimates the precipitation value at a point with a smaller mean of absolute errors, it is only a slight improvement. This can also be observed in Figure 4. Here, one can see the problematic periods for interpolation: During February when the heavy snow falls occurred and during May, probably as transition period from winter to summer, the errors are the highest. Out of this study it can clearly be seen that detrending of daily precipitation values does not bring any improvement for interpolation. It seems that daily precipitation sums depend more on single precipitation events than by a – at least climatically present - precipitation gradient derived by a precipitation-altitudinal relationship.

For the interpolation of the daily air temperature inverse distance weighting is applied with a constant gradient of –0.65°C/100m combined with a calculated gradient depending on R^2 to normalise the data to elevation:

$$\frac{dT}{dz} = \frac{R^2}{0.4} \cdot \left(\frac{dT}{dz}\right)_{calc} + \frac{0.4 - R^2}{0.4} \cdot \left(\frac{dT}{dz}\right)_{const}$$

where $\frac{dT}{dz}$ is the temperature gradient with the indices *calc* for the calculated and *const* for the constant value (1). Before interpolating observed air temperature, the data are analysed and stations more or less permanently lying within inversion layers are removed. A typical day for such situations was January 20, 1999 (Figure 5, right). On the contrary, Figure 5 (left) gives an example of a day with a clear correlation between temperature and altitude. Cross-validatation of the temperature interpolation results in an average of absolute errors of 1.4 K for the total area of Switzerland.

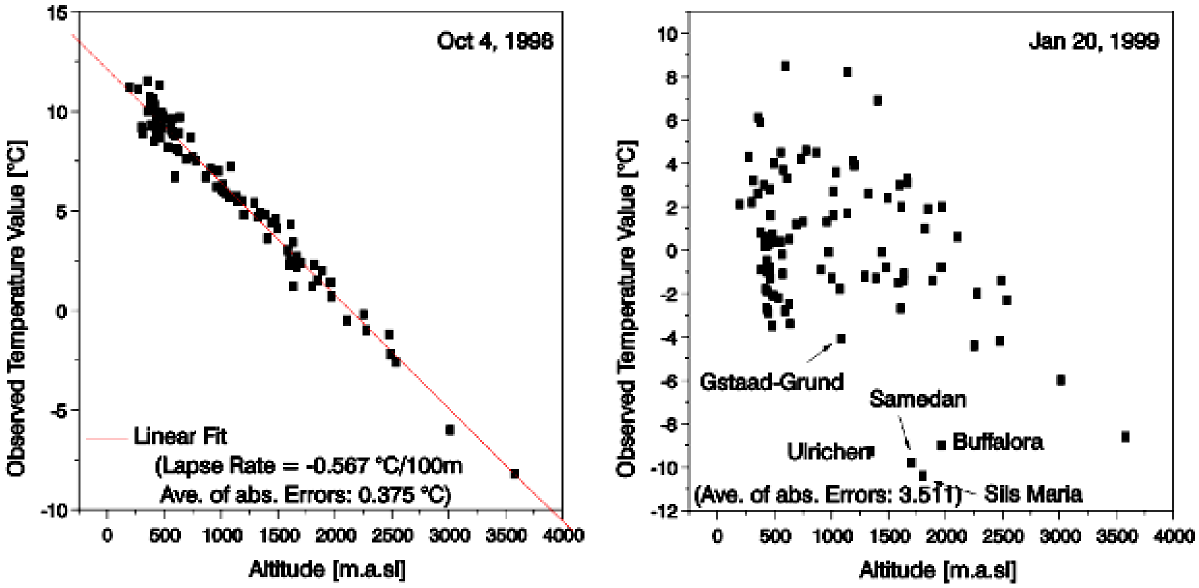

Figure 5: Temperature-elevation relationship for Oct 4, 1998 (left) with minimal interpolation error and for Jan 20, 1999 with a high interpolation error due to inversion layers

RESULTS

The results presented are based on a first simulation run with standard parameters. Validation for snow depth and snow water equivalient (SWE) is done by using point measurements. Observed snow depth is available daily during winter months whereas SWE is measured at fewer snow gauges and only on a fortnightly basis. Figure 6 shows observed snow depth and SWE against simulated values at selected snow stations in four different alpine regions at about the same height. Table 2 is a listing of some characteristics of the four stations with absolute average difference = $\overline{|obs. - sim|}$ and relative average difference = $\frac{1}{n}\sum\left(\frac{|obs - sim|}{\overline{obs}}\right)$.

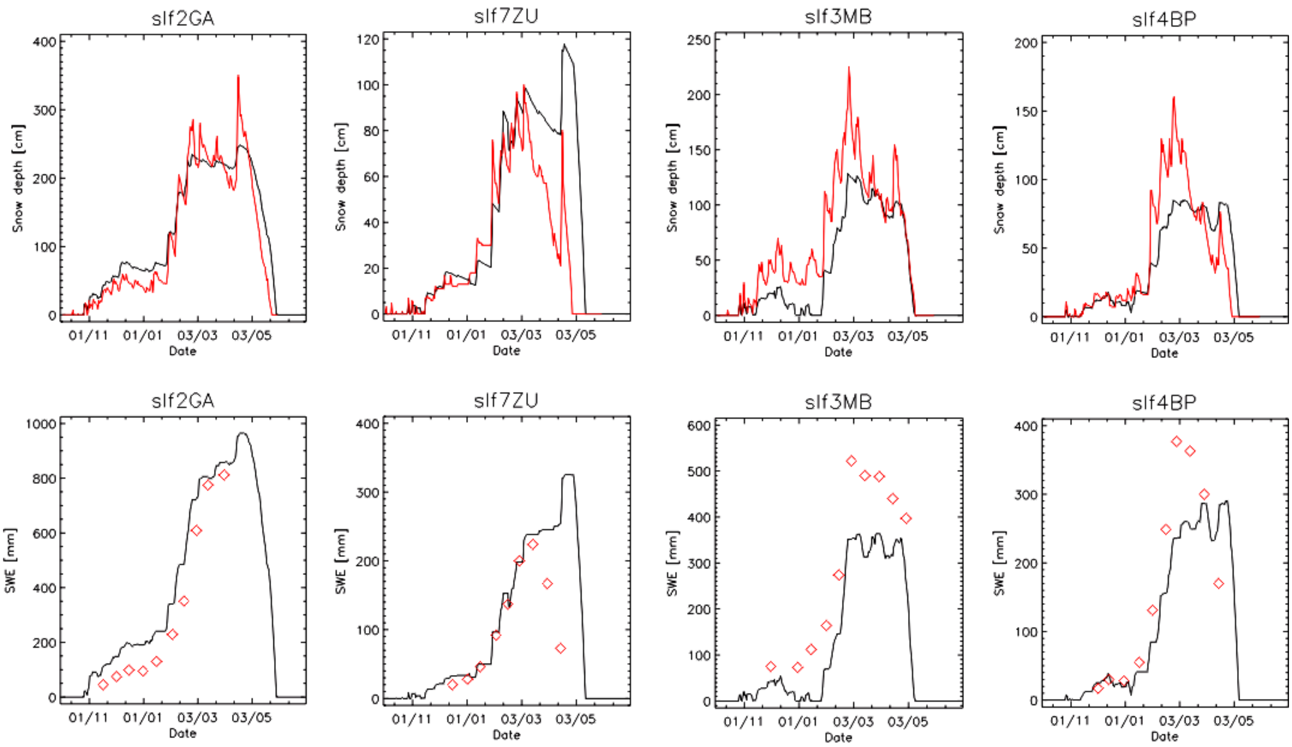

Figure 6: Observed (red) and simulated (black) snow depth (above) and snow water equivalent (below) at four selected stations in different regions. The stations and regions (as calibration centres) can be seen in Figure 7.

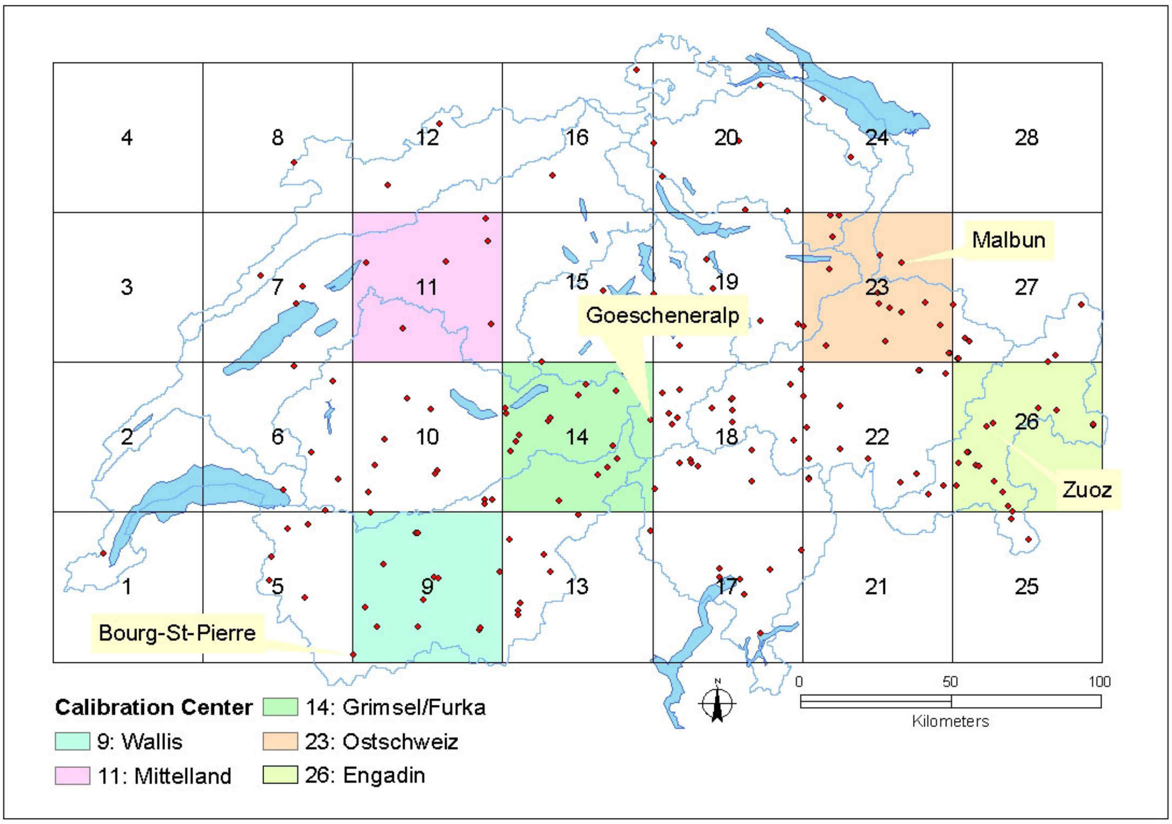

Figure 7: Calibration centres as referenced in Table 2 (numbered boxes filled with colours) and all the snow gauges relevant for this study (red points). The four stations discussed in this work are labelled.

The calibration centres are defined as illustrated in Figure 7. A visual validation of the plots shows quite good results. While Göscheneralp simulates an almost perfect fit, the two stations Malbun and Bourg-St-Pierre seem to suffer the same two things: too little snow accumulation and a too late and not very clear start of the ablation period. The latter is also occurring at Zuoz with the effect that at the beginning of March to the end of April there is much more snow simulated than observed. Detailed information about the absolute and relative average difference of snow depth and SWE respectively is given in Table 2.

Table 2: Characteristics of four selected snow gauges

Station code	**Name**	**m.a.sl.**	**Calibration centre**	**Abs. average difference of snow depth**	**Rel. average difference of snow depth**	**Abs. average difference of SWE**	**Rel. average difference of SWE**
Slf2GA	Göscheneralp	1750	14	21.3 cm	0.20	88.2 mm	0.27
Slf7ZU	Zuoz	1710	26	16.1 cm	0.59	36.3 mm	0.32
Slf3MB	Malbun	1610	23	22.2 cm	0.36	117.5 mm	0.38
Slf4BP	Bourg-St-Pierre	1610	9	14.2 cm	0.40	53.8 mm	0.31

Validation of the distribution of the snow cover is done by comparing the simulated snow maps with snow maps derived from NOAA/AVHRR. In Figure 8 validation on three dates is illustrated. Especially on February 12, shortly after the heavy snow falls, there is a remarkable overestimation of the snow cover by the snow model in less elevated areas. But also during the beginning of the ablation period (March 24) the snow model computes more snow than is observed by NOAA/AVHRR. This corresponds to the validation at the snow gauges (see above).

CONCLUSIONS

The tests on the interpolation of precipitation demonstrate that a normalisation of the observed precipitation values on a reference altitude (detrending) does not improve the results. Cross-validation of the interpolation results show a better estimate using ordinary kriging. However, there is not an outstanding difference between the interpolation method of inverse distance weighting (IDW) and ordinary kriging. Especially if one takes into consideration the lower computation speed of ordinary kriging and the straightforward way of implementing IDW, IDW still suits sufficiently the demands of the snow model.

Results of a standard simulation are quite good but they show the problems of the model without calibration. Calibration is an important process and still has to be done. By correcting the degree-day factor at the beginning of March it should be possible to improve the ablation, which starts too late. Many stations show a too low accumulation during the main accumulation phase. This can be accounted for by applying the calibration parameter that corrects the under-catchment of precipitation during snowfalls.

The snow model originally designed for the Bernese Alps is working on the whole area of Switzerland quite well even without any calibration. The possibility of calibrating the model within different calibration centres according to regions means that the model is quite flexible. A further advantage is its simplicity since just the two input parameters of daily air temperature and precipitation values are needed.

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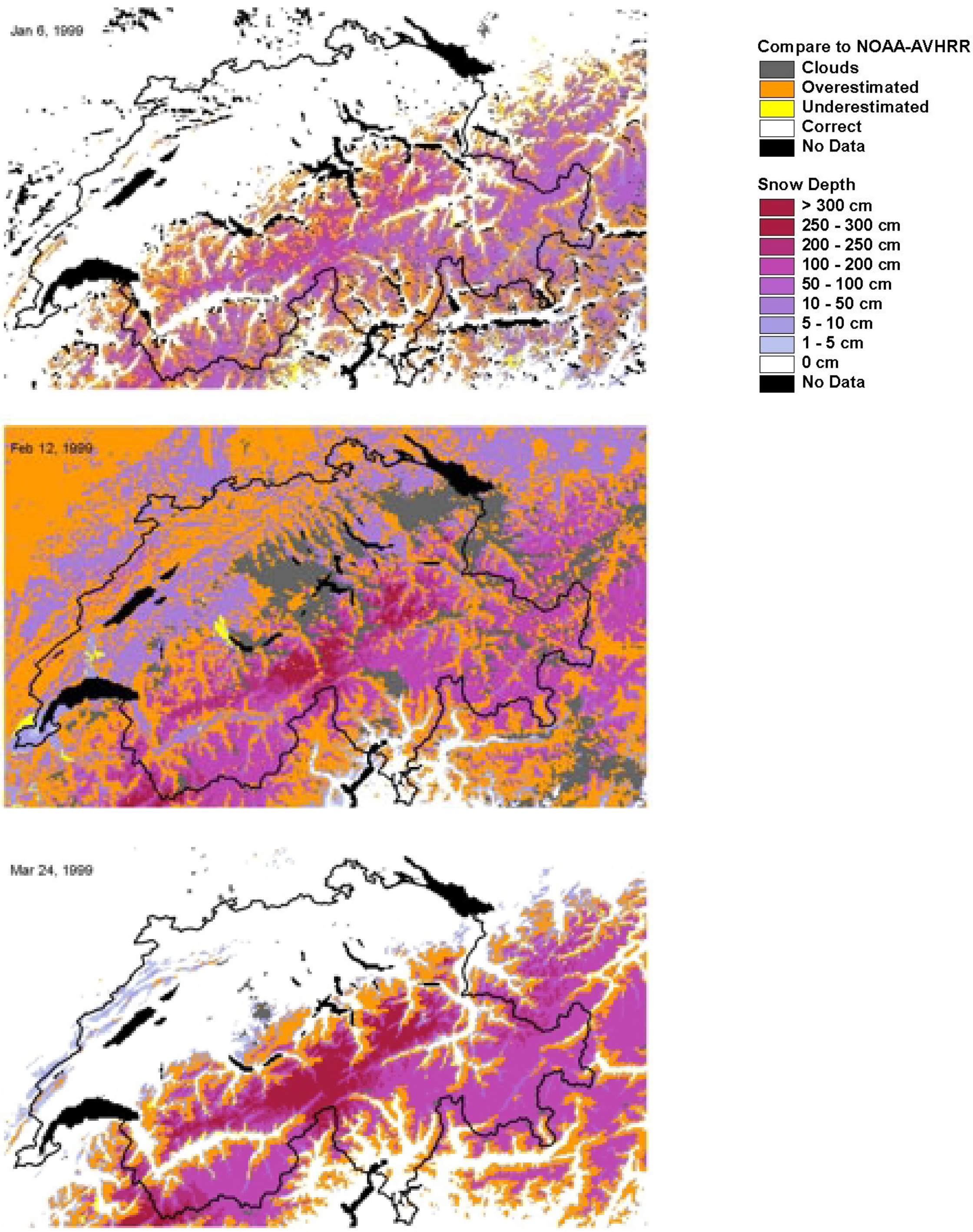

Figure 8: Comparison of simulated snow maps with snow maps derived from NOAA/AVHRR data on Jan 6 (top), Feb 12 (middle) and March 24, 1999 (bottom). Orange: overestimation by the snow model, yellow: underestimation by the snow model. Light blue to dark red: snow depth from 0 cm to > 3 m. Gray: Clouds, Black: no data and lakes.

REFERENCES

1. Kleindienst, H. 2000. Snow hydrological models as tools for snow cover assessment and water resources management. Inauguraldissertation, Universität Bern, Bern.
2. Kustas, W. P. and Rango, A. 1994. A simple energy budget algorithm for the snowmelt runoff model. WRR 30: 1515-1527.

3. Cazorzi, F. and Dalla Fontana, G. 1996. Snowmelt modelling by combining air temperature and a distributed radiation index. J. of Hydrology 181: 169-187.
4. Vehviläinen, B. 1991. A physically based snowcover model. Recent advances in the modelling of hydrologic systems, edited by Bowles, D. S. and O'Connell, P. E. (Dordrecht, Kluwer Academic Publishers**),** 113-135.
5. Gesell, G. 1989. An algorithm for snow and ice detection using AVHRR data. An extension to the APOLLO software package. Int. J. Remote Sensing Vol. 10 Nos. 4 and 5: 897-905.
6. Wunderle, S. and Hauser, A. 2001. Projektantrag SALSA. (Geographisches Institut Universität Bern, Bern)
7. Sevruk, B. 1997. Regional dependency of precipitation-altitude relationship in the Swiss Alps. Climatic Change 36(3/4): 355-369.
8. Phillips, D. L., Dolph, J. and Marks, D. 1992. A comparison of geostatistical procedures for spatial analysis of precipitation in mountainous terrain. Agricultural and Forest Meteorology 58: 119-141.
9. Hevesi, J. A., Istok, J. D. and Flint, A. L. 1992. Precipitation estimation in mountainous terrain using multivariate geostatistics. Part I: Structural analysis. Journal of Applied Meteorology 31: 661-688.
10. Isaaks, E. H. and Srivastava, R. M. 1989. Applied Geostatistics. (New York, Oxford University Press)
11. Burrough, P. A. and McDonnel, R. A. 1998. Principles of geographical information systems. (Oxford, Oxford University Press)
12. Jones, C. 1997. Geographical information systems and computer cartography. (Essex, Addison Wesley Longman)
13. Pebesma, E. J. (1999): gstat 2.2.1. (http://www.gstat.org/)
14. Sevruk, B. and Mieglitz, K. 2001. The effect of topography, season and weather situation on daily precipitation gradients in 60 swiss valleys. (preprint from B. Sevruk via e-mail on the 16.5.2001)