MODELLING SNOW ACCUMULATION AND SNOWMELT RUNOFF – PRESENT APPROACHES AND RESULTS

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ABSTRACT

The aim of the article is to give a review of methods applied for modelling the snow accumulation and snowmelt and to give a description of main processes governing the runoff from the snowpack. The progress in the understanding of processes running in the snowpack is documented both by worldwide results presented in many studies and by the results achieved by the authors in the selected small experimental catchments in the Czech Republic. The research is focused on 1) measuring the snowpack and analysing its spatial and temporal distribution, 2) assessing the role of different physical-geographical factors on snow accumulation and melting, 3) testing methods for interpolation of measured point data into area, and 4) modelling the snow accumulation and snowmelt in the local and regional scale. The main findings of the research show various ways of behaviour of snowpack accumulated in the forest and open areas in experimental catchments and show the most suitable interpolation methods taking into account one or more independent variables (slope, aspect, altitude, vegetation) for calculating the dependent variable (snow water equivalent, snow depth). The presented results also confirm the known problems with applying temperature-index snowmelt model, mainly for modelling the situations when the air temperature fluctuates near zero and for modelling the diurnal fluctuation of the snowmelt runoff.

Key words: snow accumulation, snowmelt, spatial interpolation, hydrological modelling, energy balance, degree-day method

1. Introduction

The snow melting caused by higher temperature, accompanied with precipitation, is a frequent cause of floods in the Czech Republic. Possibilities of runoff forecast in monitored profiles significantly improved thanks to mathematical modelling. For an accurate prediction it is necessary to enter catchment characteristics and also initial conditions representing altogether the catchment status at the beginning of the simulation. The most important initial condition for the flood forecast caused by snow melting is the snow layer status, mainly snow depth and snow water equivalent (SWE).

Numerous studies indicate various effects of both natural and anthropogenic factors on runoff formation (Jost et al. 2007; Váňová and Langhammer 2011) especially the role of the landscape and different land cover (forest, meadows, arable land). The majority of studies are focused on floods caused by liquid precipitation (Čurda et al. 2011; Jeníček 2009) or they assess the runoff on a long-term scale (Özdogan 2011; Kliment et al. 2011). Many physical-geographic factors are important in the course of accumulation and melting of snowpack in winter and in spring. The effects of altitude and air temperature are significant for large areas that exhibit high variability in terms of elevation (Essery 2003; Jost et al. 2007), while the effect of vegetation predominates locally, namely the presence of forests and open areas in the river basin (Jost et al. 2009; Tanasienko et al. 2009). Jeníček and Taufmannová (2010) has used the HEC-HMS model including the temperature-index snowmelt model for the runoff simulation from various vegetation covers (forest and open areas). The results indicate an applicability of temperature-index method for variant simulations of the rainfall-runoff process, especially for "rain-on-snow" events.

Numerous studies show that slower snow melting in the forest is caused by lower short-wave radiation (Verbunt et al. 2003). The forest also affects wind speed, interception and density of the snowpack (Holko et al. 2009). Essery (2003) tested the distributed MOSES model in high mountainous areas. This author indicates that a significant increase in accuracy was achieved by dividing the river basin to multiple zones based on elevation above sea level. Even better results were obtained by incorporating the effect of vegetation. Pobříslová and Kulasová (2000) applied the balance hydrological HBV-ETH model in experimental river basins in the Jizerské Mountains in order to simulate snow deposition and melting in the forest and in open areas. The results of direct observations confirmed markedly faster snow melting in deforested areas than in forests partly due to higher snow accumulation and partly due to higher short-wave radiation amount in the open areas. Weigert and Schmidt (2005) pointed to another problem by modelling the runoff in winter conditions. Their research was focused on infiltration processes under frozen soil by means of EROSION 3D model. The role of the frozen soil was investigated also by Bayard et al. (2005).

Concerning possible long-term changes in climate conditions there is a question of climate change impact

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on the runoff regime in the regional and local scales (Jain et al. 2011; Teutschbein et al. 2011). An important part of the research in mountain areas is therefore the issue of possible future changes in snowpack and glacier regimes. The main problem is evident; spatial and temporal distribution changes of the snowpack and glaciers will cause the changes in runoff regime (both seasonal regime changes and hydrological extremes changes) with consequences on the water and energy supplies (Janský et al. 2011).

Nevertheless, the modelling activity itself indicated a high level of mutual relationships among individual factors, which makes more difficult the proper identification of the model parameters and its calibration. Precisely, such identification and proper estimation of hydrological model parameters is one of important spheres of current research.

2. Models based on energy balance of the snowpack

Physical approach to modelling the runoff from snowpack is based on energy balance. This method considers and quantifies energy flux on the interfaces of atmosphere–snow–soil. Energy balance could be defined by (formula 1):

$$Q_m = Q_{nr} + Q_h + Q_e + Q_p + Q_g + Q_q \tag{1}$$

where Q_m is the energy balance accessible for snow melting [W m^{-2}], Q_{nr} is the radiation balance, Q_h and Q_e are turbulent sensible heat and latent heat fluxes on the interface of snow – atmosphere, Q_p is the heat supplied by precipitation, Q_g is the heat supplied by soil a Q_q is the heat change inside snowpack [W m^{-2}].

According to Singh and Singh (2001) it is possible to calculate the water volume from melting snow (formula 2):

$$M = 0.0031 \cdot Q_m \tag{2}$$

where M [mm d^{-1}] is the amount of melting water and Q_m [kJ m^{-2} d^{-1}] is the daily output of energy balance (only positive values). The value 0.0031 is determined for ripe snow and liquid water content equals 3%. The snowpack can be represented as one layer in the computation, but it is possible to derive the energy balance for more snow layers taking into account the stratification of the snowpack.

2.1 Shortwave and longwave radiation

Total radiation balance of the snowpack Q_n is possible to compute according to formula 3 (Singh and Singh 2001):

$$Q_{nr} = (1 - \alpha) \cdot S_i + (L_i - L_o) \tag{3}$$

where Q_{nr} is the total radiation, α is the albedo, L_i is the incoming radiation and L_o is the outcoming longwave radiation. S_i introduces the incoming shortwave radiation and it is sometimes replaced with global radiation G. The global radiation involves both direct and diffusive shortwave radiations. The amount of shortwave radiation captured by snow depends on several factors, mainly the slope, aspect, season, latitude, atmospheric diffusion and land cover, which can prevent the direct radiation into the snowpack.

The sources of the longwave radiation (6.8–100 μm) are the atmosphere and the Earth surface (Singh and Singh 2001). The radiation of surface is largely absorbed by the atmosphere, above all due to carbon dioxide and water vapour. An exception is the radiation within the interval 8–12 μm (infrared atmospheric window). The last two items of the formula 3 represent the radiation longwave balance of the snowpack. Based on Stefan-Boltzmann law using emissivity ε and snowpack temperature T_s [K] it is possible to express the relation between the incoming and outcoming longwave radiations (formula 4):

$$L_o = \varepsilon\sigma T_s^4 + (1 - \varepsilon) \cdot L_i \tag{4}$$

where σ refers to the Stefan-Boltzmann constant ($5.67 \cdot 10^{-8}$ W m^{-2} K^{-4}). The energy of incoming longwave radiation for sky without clouds is possible to derive by (formula 5):

$$L_i = (0.575 \cdot e_a^{1/7}) \cdot \sigma T_a^4 \tag{5}$$

where e_a [mbar] refers to the water vapour tension and T_a refers to the air temperature. Considering the sky without clouds, air temperature 2 °C, snowpack temperature 0 °C, snowpack emissivity 1 and relative air moisture 50% and using formulas 3–5, we obtain the longwave radiation balance of the snowpack equal −92 W m^{-2} (DeWalle and Rango 2008).

2.2 Sensible and latent heats

A convective sensible heat transfer is running between the air and the snow in case of different air and snowpack temperatures due to turbulent atmospheric circulation. The direction and the amount of energy flux depend on temperature differences, wind speed, snow surface roughness and atmospheric stability (stability of dry air). Large snow and air temperature differences in spring and early summer caused energy gain of snowpack. However, the snowpack temperature is often higher than the air temperature in the spring nights. This situation represents snowpack energy loss (DeWalle and Rango 2008).

The latent heat transfer represents the transfer of water vapour between the atmosphere and the snowpack. The water vapour flux and its direction are determined by the vapour pressure gradient and the intensity of the turbulence. The snowpack energy loss represents the evaporation and the sublimation from the snow, the snowpack energy gain represents the condensation and the desublimation of the water vapour on the snow surface (DeWalle and Rango 2008).

2.3 Energy supplied by precipitation

In the case of liquid precipitation on snow (so called rain-on-snow events) the precipitation is cooled to the snow temperature. The heat supplied to the snowpack equals the difference of snow energy before the precipitation and after the achievement of temperature stability. Formula 6 expresses the daily amount of transferred energy Q_p [kJ m^{-2} d^{-1}] depending on the precipitation P_r [mm d^{-1}] and the temperature difference (Singh and Singh 2001).

$$Q_p = \frac{(\rho_w c_p \cdot (T_r - T_s) \cdot P_r)}{1000} \quad (6)$$

where ρ_w refers to the water density, c_p refers to the specific heat capacity of water (4.20 kJ kg^{-1} °C^{-1}), T_r is the temperature of liquid precipitation and T_s is the temperature of snowpack.

2.4 Ground heat flux

This part of energy balance usually plays a minor role because of the small heat conductivity of the soil and, in case of higher snow depth, the lack of solar radiation supplied to the soil. The soil heat causes the slow ripening of snowpack during the winter and may cause slow snowmelt.

2.5 Change of internal energy of the snowpack

The internal energy of the snowpack depends on snow temperature and can be expressed as the sum of internal energy of three snowpack parts – ice, water and water vapour. The internal energy of the snow equals the internal energy of the ice in case of snow temperature below zero. The change of internal energy doesn't play a significant role in the whole energy balance and thus it is possible to neglect this part of energy balance (Singh and Singh 2001).

The advantage of methods based on energy budget is their broad possibilities of use in several climatic conditions. It is possible to describe precisely the physical nature of snow accumulation, transformation and melting processes using the energy budget method. The main disadvantage of the energy budget method is the difficulty to obtain the input data necessary for parameterization, calibration and validation of the model (Ohmura 2001). Selected research areas of snow processes made by several authors using the energy balance method are listed in the table 1.

3. Models based on temperature index method

The task of determining individual items of energy balance is quite complex, therefore the so called index methods are often used. These methods make use of the connection between the snow melting and an easily available magnitude that is related to energy balance (air temperature, precipitation). The temperature index method and its various modifications are most widely used. The air temperature is a broadly used magnitude in hydrological modelling, as it provides a simplified description of energy exchange at the atmosphere – snowpack – ground surface interface. The SWE decrease M [mm d^{-1}] is calculated according to formula 7:

$$M = a \cdot (T - T_C) \quad (7)$$

where a [mm °C^{-1} d^{-1}] is the melt factor, which is the SWE decrease in a day caused by the air temperature change T of 1 °C compared to the critical value T_c in which the melting process begins. T is the mean air temperature. The value of T_c varies mainly from 0 to 2 °C. The value of temperature factor is not constant and varies depending on physical characteristics of snow or global radiation. Values vary from 1 to 7 mm °C^{-1} d^{-1}. The temperature index method was defined for the daily time step, but it is possible to use a shorter or longer temporal resolution.

Tab. 1 Research areas solved by the means of energy balance method

Research area	*Reference*
Snow accumulation and snowmelt modelling, model development and calibration, operational hydrology	Garen and Marks (2005); Herrero et al. (2009); Ohara and Kavvas (2006); Walter et al. (2005); Fierz et al. (2003); Franz et al. (2008); Sensoy et al. (2006)
Effect of vegetation	Jost et al. (2009); Koivusalo and Kokkonen (2002); Burles and Boon (2011)
Glacier melt modelling	Magnusson et al. (2011)
Climate change modelling	Dankers and Christensen (2005); Özdogan (2011); Strasser and Marke (2010); Uhlmann et al. (2009)

The temperature index method was derived and firstly used by Finsterwalder and Schunk (1887) in the glaciological study in the Alps; nowadays it is still the most widely used snowmelt method among hydrologists. Hock (2003) summarises the advantages of the temperature index method as 1) the good availability of air temperature data, 2) the relatively simple spatial distribution of the air temperature and its predictability, 3) the simplicity of computation procedure, and 4) satisfactory results of the model despite of its simplicity. Beven (2001) formulated shortcomings of this method as 1) the accuracy of the method decreases with the increasing temporal resolution, 2) the intensity of the snowmelt has a big spatial variability depending on topographic conditions, mainly the slope, aspect and land cover. This variability is very difficult to express using temperature index method.

The temperature index method comes from the linear dependence of snowmelt on the air temperature. Incoming longwave radiation and sensible heat represent 3/4 of total energy balance of the snowpack (Hock, 2003). Shortcomings of the models based on the temperature index method with daily time step are apparent mainly in cases of air temperature fluctuations near zero (Jeníček and Taufmannová 2010; Kutláková and Jeníček 2012). The mean daily air temperature indicates no snowmelt; however, the positive air temperature which occurs during day could cause the snowmelt (Hock 2003). In mountain areas it is necessary to consider the change of the air temperature depending on the altitude; therefor the basin is usually divided into several elevation zones.

The melt factor a is the key parameter in the formula 7. Martinec (1977) derived an empirical relation between the melt factor and the snow density as (formula 8):

$$a = 1.1 \cdot \frac{\rho_s}{\rho_w} \tag{8}$$

where a is the melt factor (mm °C^{-1} d^{-1}), ρ_s is the snow density (mass of snow divided the bulk volume of the snowpack) and ρ_w is the water density. Formula 8 points to the increasing tendency of the melt factor in the spring together with the increasing snow density due to the ripening process. Kuusisto (1980) derived a relation between the melt factor and the snow density separately for forest and open areas (formulas 9, 10):

$$\text{forest: } a = 0.0104 \cdot \rho_s - 0.7 \tag{9}$$

$$\text{open areas: } a = 0.0196 \cdot \rho_s - 2.3 \tag{10}$$

Forests cause the decreasing of the direct solar radiation and therefore the snowmelt in the periods without precipitation. Different melt factors derived Federer et al. (1972) for the northwest of the USA. He experimentally derived the melt factor 4.5–7.5 mm °C^{-1} d^{-1} for open areas, 2.7–4.5 mm °C^{-1} d^{-1} for deciduous forests and 1.4–2.7 mm °C^{-1} d^{-1} for coniferous forests (approximate ratio 3 : 2 : 1). Kuusisto (1980) expressed varying of the melt factor in dependence on the cover density of coniferous forests (formula 11):

$$a = 2.92 - 0.0164 \cdot C_c \tag{11}$$

where C_c is the rate of treetop cover. Coniferous forests reach the values from 0.1 to 0.7. Table 2 contains the list of factors which have an influence on melt factor.

Tab. 2 Factors influencing melt factor (DeWalle and Rango 2008)

Factor	*Cause*	*Influence on melt factor*
Seasonal influence	Decrease of cold content and albedo, increasing of shortwave radiation and snow density	Melt factor increases
Open area vs. forest	Shading and wind protection	Lower melt factor and its spatial variability in the forest
Topography (slope, aspect)	Variability of shortwave radiation and wind exposure	Higher melt factor in south hill slopes
Snow cover area	Spatial snowmelt variability	Melt factor decreases in the basin with larger snow cover area
Snowpack pollution	Dust and other pollution causes lower albedo	Higher melt factor
Precipitation	Rainfall supplies sensible heat, clouds decrease solar radiation	Generally lower melt factor in rainy days due to lower radiation. But precipitation itself cause higher melt factor.
Snow vs. ice	Glacial ice has lower albedo than snow	Higher melt factor in glaciated basins
Meteorological conditions for certain air temperature	Higher snowmelt in case of higher wind speed, higher radiation or higher moisture by the same temperature	Higher melt factor

Tab. 3 Research areas solved be the means of temperature-index method

Research areas	*Reference*
Snow accumulation and snowmelt modelling, model development and calibration, operational hydrology	Walter et al. (2005); Brubaker et al. (1996); Hu et al. (2006); Kutláková and Jeníček (2012)
Effect of vegetation	Federer et al. (1972); Kuusisto (1980); Jeníček and Taufmannová (2010); Pobříslová and Kulasová (2000)
Glacier melt modelling	Finsterwalder and Schunk (1887); Magnusson et al. (2011)
Climate change modelling	Steiger and Mayer (2008)
DDF derivation	Federer et al (1972); Kuusisto (1980); Martinec (1977); Hasa (2010)

Table 2 shows the increasing melt factor due to the precipitation, which could be described by formula 12 (Singh a Singh 2001):

$$a = a' + 0.0126 \cdot P \qquad (12)$$

where a is the melt factor with precipitation influence, a' is the melt factor without precipitation influence and P (mm) is the precipitation.

The temperature index method is widely used in numerous variants for modelling snow accumulation and melting. Selected research areas of snow processes made by several authors using temperature-index method are listed in the table 3.

4. Summary of results from the Krušné Mountains

Periodical measuring of the snow depth and SWE in the Department of Physical Geography and Geoecology (Faculty of Science, Charles University in Prague) started in 2006 in the Šumava Mts. (Jeníček et al. 2008). Snow measurements are more intensively made in the spring time in order to describe the relation between snow melting and stream hydrographs in the experimental catchments.

The snow monitoring in the Bystřice River basin (stream gauge Abertamy, 9.0 km^2) and the Zlatý Brook basin (stream gauge Zlatý kopec, 5.5 km^2) in the Krušné Mts. began in the winter 2008/2009 (Fig. 1). In the regional scale the research is carried out in the Bystřice River basin upwards stream gauge Ostrov (127.6 km^2). The aim of the research is 1) to assess the effect of selected physical-geographic factors on snow accumulation and snowmelt runoff dynamics; 2) to develop methods for snow measuring and spatial interpretation of snow data and; 3) to model the snow accumulation and the snowmelt including snowmelt runoff scenarios under changing conditions (effect of vegetation, effect of climate change).

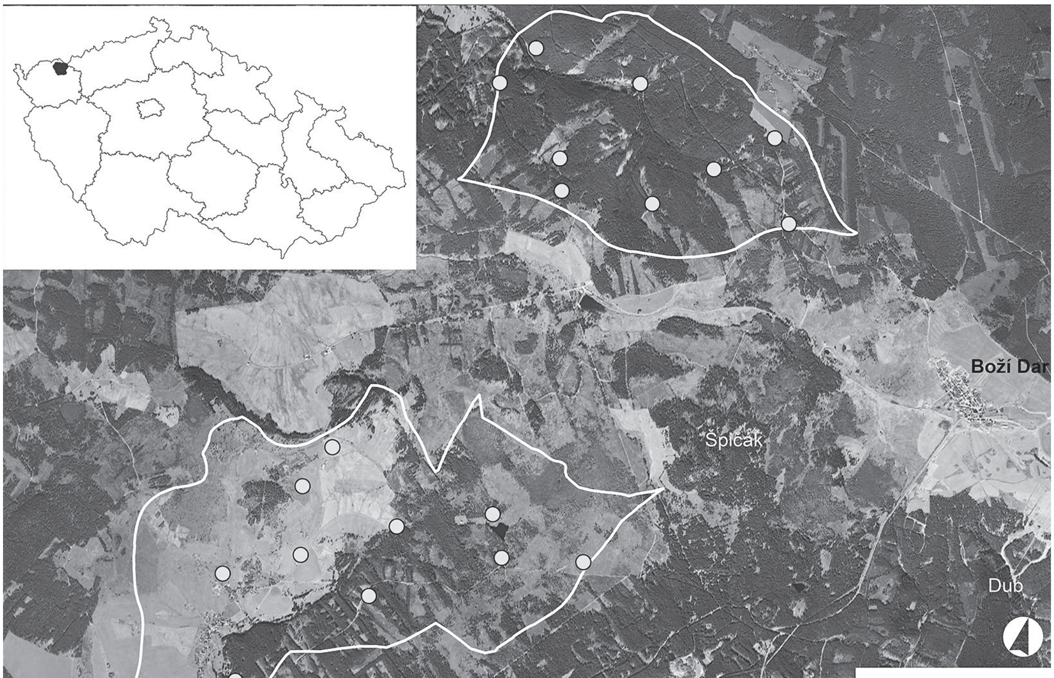

Fig. 1 Location of experimental catchments Bystřice River (south, stream gauge Abertamy) and Zlatý Brook (north, stream gauge Zlatý kopec) with measured points and the location of Bystřice River basin (stream gauge Ostrov, overview map of the Czech Republic)

Both of the studied headwater river basins belong to the highest elevated basins in the Krušné Mountains (750–1050 m a.s.l.) and there is an assumption of high amount of the water accumulated in the snowpack and thus a spring flood risk. The Bystřice River basin and the Zlatý Brook basin represent not only two different basins in terms of vegetation (95% of forest cover in the Zlatý Brook basin, 46.5% of forest cover in the Bystřice River basin), but they differ also in terms of orography, especially the windward and leeward effects. Differences between the snow accumulation and the snowmelt in the forest and open areas documented with the average of point measurements in both catchments are evident from the figure 2.

The main aim of the study performed by Kučerová and Jeníček (2011) was to asses selected interpolation methods in terms of their prediction capability of snow water equivalent (SWE) in unknown points. The study was carried out in the Bystřice River basin (127.6 km^2, figure 1). Point data measured in the basin were transferred into grids with resolution of 60 × 60 metres using nine interpolation methods which belong to deterministic, geostatistical and global methods.

The prediction of SWE in unknown points was assessed by means of cross validation. Differences between estimated values and measured values were evaluated by several parameters (figure 3). The best prediction quality of SWE in unknown points was achieved by means of geostatistical methods (ordinary kriging, cokriging and residual kriging) and global methods (linear regresion) compared to deterministic methods (Thiessen polygons, inverse distance weighting, global and local polynomial and radial basis functions). The linear regression using the altitude as an independent variable was the best evaluated interpolation method according to the most parameters (Kučerová and Jeníček 2011).

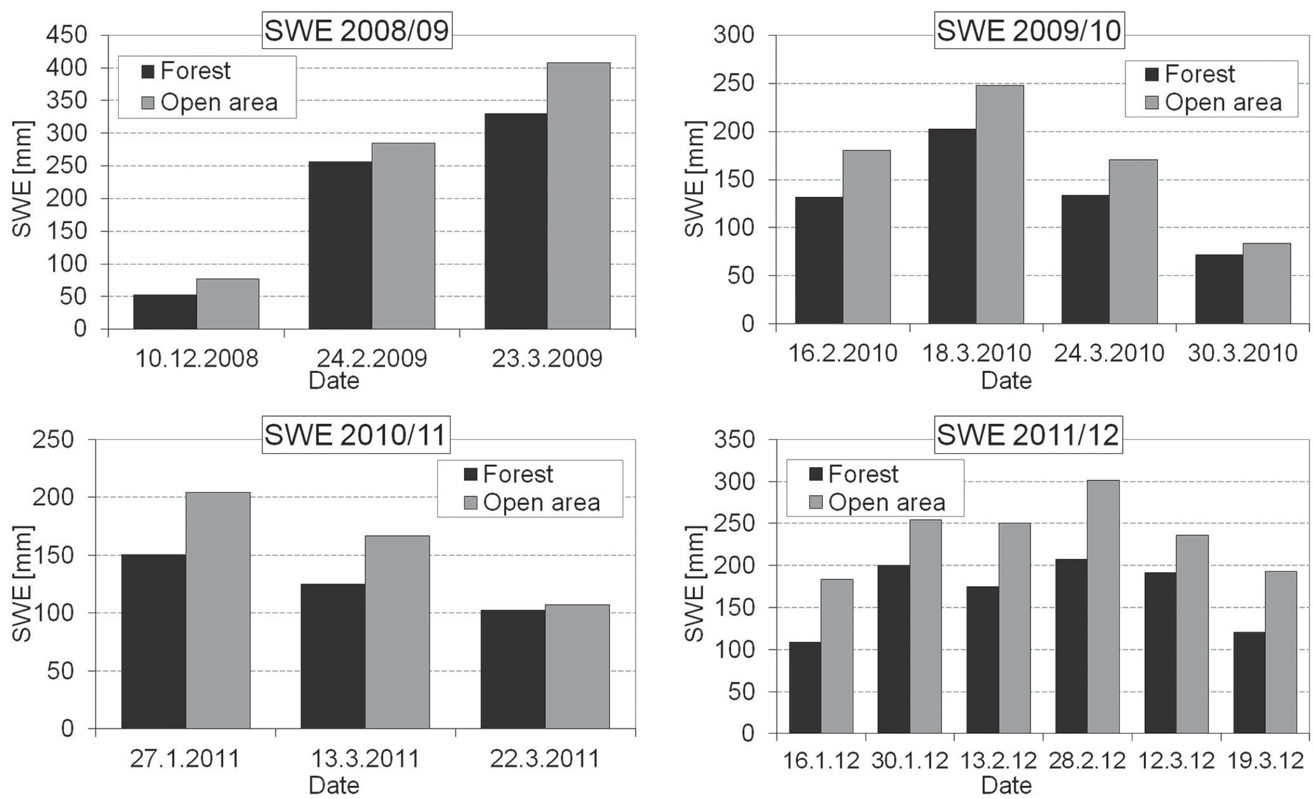

Fig. 2 Snow water equivalent (SWE) in the forest and in open areas during the winters 2008/09, 2009/10, 2010/11, and 2011/12 in the Bystřice River basin and in the Zlatý Brook basin

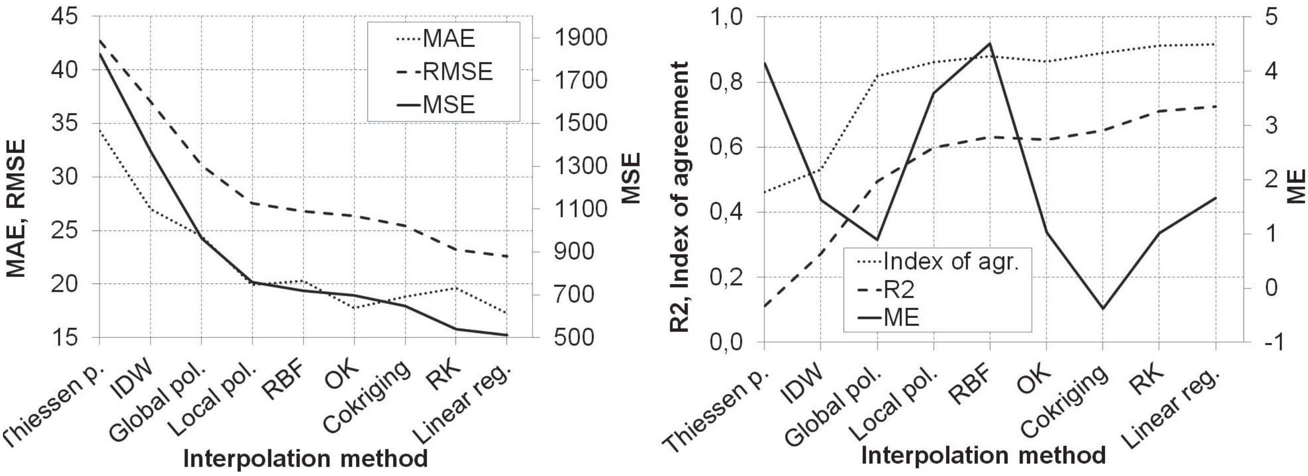

Fig. 3 Parameter values evaluating interpolation methods in terms of quality of their prediction of dependent variable (SWE) in unknown points (MAE – mean absolute error, RMSE – root mean square error, MSE – mean square error, Index of agreement, ME – mean absolute error, R2 – coefficient of determination)

The findings of the study are in agreement with the results achieved by others authors. According to the study, the global and geostatistical methods give a generally better prediction of climatic variables in unknown points compared to deterministic methods (e.g. Carrera-Hernández and Gaskin 2007; Haberlandt 2007; Lloyd 2005). Moreover, the prediction quality may be enhanced by including one or more independent variables in the calculation of the dependent one – slope, aspect, altitude, vegetation (Erxleben et al. 2002; López-Moreno and Nogués-Bravo 2006).

The temperature index approach was used in order to derive degree-day factors (DDF) in the headwater part of the Bystřice River basin (Hasa 2010). Degree-day factors were calculated by using the air temperature and flow rates in the stream gauge Abertamy for spring snowmelt periods 2009 and 2010. The snow water equivalent of melted snow was simplified to the runoff depth derived from the continual monitoring of the flow rates in the stream gauge Abertamy. The air temperature was measured at the climatological station Hřebečná located in the middle of the catchment. The degree day factor was calculated for each snowmelt period; 1.40 mm $°C^{-1}$ d^{-1} and 1.48 mm $°C^{-1}$ d^{-1} for years 2009 and 2010, respectively. Derived values are markedly smaller comparing to those derived by Martinec (1977), Kuusisto (1980) and Federer et al. (1972) using the land cover data and measured snow density (table 4). The DDF analysis was also made in the Ptačí Brook experimental catchment (Bohemian Forest, southwest of the Czech Republic). This analysis was based on point measured data of SWE and the monitoring of precipitation, air temperature and discharge during winters 2010/11 and 2011/12. The derived DDF varies between 2.77 mm $°C^{-1}$ d^{-1} and 3.28 mm $°C^{-1}$ d^{-1} according to assessed year and season. The results correspond better with the results of above mentioned studies. Computed degree day factors are applicable for modelling the runoff from the melting snow in the experimental catchment.

Jeníček and Taufmannová (2010) have used the HEC-HMS model with the temperature-index snowmelt model for the runoff simulation from different vegetation covers (forest and open areas). Table 5 shows all methods used for simulation the runoff from the Bystřice River basin in the HEC-HMS model.

Tab. 4 Empirical degree day factors (DDF) in the Bystřice River basin

Method	*DDF (mm °C^{-1} day^{-1})*	*Required data*
Derived from point measured data (Hasa 2010)	1.40–1.48	SWE, discharge, air temperature
Martinec (1977)	3.85	snow density
Kuusisto (1980)	4.12	snow density + land cover
Federer et al. (1972)	4.92	land cover

Tab. 5 The list of HEC-HMS components used for runoff simulation

HEC-HMS Component	*Applied method*
Snow accumulation and snowmelt model	Temperature index (DeWalle and Rango 2001)
Runoff-volume model	SCS CN - Soil Conservation Service Curve Number (Feldman 2000)
Direct-runoff model	Clark Unit Hydrograph (Feldman 2000)
Baseflow model	Exponential Recession (Feldman 2000)
Channel model	Muskingum-Cunge (Feldman 2000)

The results indicate an applicability of temperature-index method for scenario simulations of the rainfall-runoff process, especially for "rain-on-snow" events. Similar results were achieved by Kutláková and Jeníček (2012) by the application of the lumped HEC-HMS model in the Bystřice River basin (stream gauge Ostrov, 127.6 km^2). The model provided good results in the higher elevations with high amount of the snow without partial thawing (above 800 m a.s.l.). However, problems were recognized in the lower parts of the basin (300–600 m a.s.l.), where air temperatures fluctuated near zero. In those parts the model sometimes wrongly differentiated between the rain and snow. One of the conclusions of the study was to use better spatial description of the basin (more elevation bands) or to use the energy balance method in order to accurate the air temperature representation and thereafter the simulation of the snow accumulation and snowmelt.

5. Discussion and conclusions

Main hydrological uncertainties are associated with the collecting and analysis of input data and with the modelling of rainfall-runoff process using the chosen snowmelt runoff method. The hydrological uncertainty is the result of 1) the uncertainty of inputs (measured data), 2) the uncertainty of the hydrological model and 3) the uncertainty of model parameters.

Regular measurements of the snow depth and SWE in experimental river basins in the Šumava Mountains and the Krušné Mountains have been going on at the Department of Physical Geography and Geoecology since the winter 2006/2007. Performed measurements have confirmed a number of important facts. In particular, the spatial characteristics of the snowpack exhibit a high variability level in mountainous conditions, while this variability can be described only with difficulty by point measurements at meteorological stations (Egli et al. 2009). In general, the biggest differences may be found when comparing snow amounts deposited in forests and in open areas (Burles and Boon 2011; Jeníček and Taufmannová 2010; Jost et al. 2009; Pobříslová and Kulasová 2000). The above mentioned conclusion works for explorations done within the framework of a local scale of the selected area. Globally, the amount and nature of snowpack accumulation is most likely decided upon the elevation above sea level (Kutláková and Jeníček 2012; Essery 2003).

The snow model that uses the temperature index method for calculations is the main component of the HEC-HMS model applied in author's studies. The method is relatively sophisticated; nevertheless, it applies several simplifications of real processes that occur during the snow accumulation and melting. The processes still remain largely unknown and their satisfactory description has not been developed yet. One of the factors, that are difficult to involve in modelling, is the frozen ground. In the model applied in the studies by Jeníček and Taufmannová (2010) and Kutláková and Jeníček (2012) it is not possible to take into account the spatial variability of frozen ground, which is the problem especially on the regional scale in the areas with higher variability of altitude. The complex nature of infiltration in the frozen ground is also apparent in other studies (Bayard et al. 2005; Weigert and Schmidt 2005).

The undertaken simulations (Jeníček and Taufmannová 2010; Kutláková and Jeníček 2012) are associated with the disadvantage of impossibility to take into account the diurnal process of snow melting due to an insufficient number of time series of air temperature with the step of 1 hour. The above mentioned fact did not cause any significant deviations in case of rain-on-snow events. However, in the case of high flows caused by the high short-wave radiation without the liquid precipitation, the simulations didn't capture diurnal deviations of the observed flow. In this case, the use of the energy balance model is advisable instead of the temperature-index method.

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REFERENCES

BAYARD, D., STÄHLI, M., PARRIAUX, A., FLÜHLER, H. (2005): The influence of seasonally frozen soil on the snowmelt runoff at two Alpine sites in southern Switzerland. Journal of Hydrology, 309 (1–4), 66–84.

BEVEN, K., J. (2001): Rainfall-runoff Modelling, The Primer. Chichester, John Wiley & Sons Ltd.

BRUBAKER, K., RANGO, A. KUSTAS, W. (1996): Incorporating radiation inputs into the Snowmelt Runoff Model. Hydrological Processes, 10, 1329–1343.

Burles, K., Boon, S. (2011): Snowmelt energy balance in a burned forest plot, Crowsnest Pass, Alberta, Canada. Hydrological Processes, 25 (19), 3012–3029.

CARRERA-HERNÁNDEZ, J. J., GASKIN, S. J. (2007): Spatio temporal analysis of daily precipitation and temperature in the Basin of Mexico. Journal of Hydrology, 336, 231–249.

ČURDA, J., JANSKÝ, B., KOCUM, J. (2011): Vliv fyzickogeografických faktorů na extremitu povodní v povodí Vydry. Geografie, 116 (3), 335–353.

DANKERS, R., CHRISTENSEN, O. B. (2005): Climate change impact on snow cover evaporation and river discharge in the sub-arctic Tana basin, Northern Fennoscandia. Climate Change, 69, 367–392.

DEWALLE, D. R, RANGO, A. (2008): Principles of snow hydrology. Cambridge, Cambridge University Press.

EGLI, L., JONAS, T., MEISTNER, R (2009): Comparison of different automatic methods for estimating snow water equivalent. Cold Regions Science and Technology, 57, 107–115.

ERXLEBEN, J., ELDER, K., DAVIS, R. (2002): Comparison of spatial interpolation methods for estimating snow distribution in the Colorado Rocky Mountains. Hydrological Processes, 16, 3627–3649.

ESSERY, R. (2003): Aggregated and distributed modelling of snow cover for a high-latitude basin. Global and Planetary Change, 38, 115–120.

FEDERER, C. A., PIERCE, R. S., HORNBECK, J. W. (1972): Snow management seems unlikely in the Northeast. In: Proceedings symposium on watersheds in transition. American Water Resources Association, 212–219.

FIERZ, C., RIBER, P., ADAMS, E. E., CURRAN, A. R., FÖHN, P. M. B., LEHNING, M., PLÜSS, C. (2003): Evaluation of snow-surface energy balance models in alpine terrain. Journal of Hydrology, 282, 76–94.

FELDMAN, A. D. (2000): Hydrologic Modeling System HEC-HMS. Technical Reference Manual. USACE, Davis.

FRANZ, K. J., HOGUE, T. S., SOROOSHIAN, S. (2008): Operational snow modeling: Addressing the challenges of an energy balance model for National Weather Service forecasts. Journal of Hydrology, 360, 48–66.

FINSTERWALDER, S., SCHUNK, H. (1887): Der Suldenferner. Zeitschrift des Deutschen und Österreichischen Alpenvereins 18, 72–89.

HABERLANDT, V. (2007): Geostatistical interpolation of hourly precipitation from rain gauges and radar for a large-scale extreme rainfall event. Journal of Hydrology, 332, 144–157.

HASA, M. (2010): Metody modelování odtoku ze sněhové pokrývky: Případová studie pramenné oblasti ostrovské Bystřice. Bachelor thesis, Prague, Faculty of Science, Charles Univ. in Prague.

HERRERO, J., POLO, M. J., MOÑINO, A., LOSADA, M. A. (2009): An energy balance snowmelt model in a Mediterranean site. Journal of Hydrology, 371 (1–4), 98–107.

HOCK, R. (2003): Temperature index melt modelling in mountain areas. Journal of Hydrology, 282, 104–115.

HOLKO, L., ŠKVARENINA, J., KOSTKA, Z., FRIČ, M., STAROŇ, J. (2009): Impact of spruce forest on rainfall interception and seasonal snow cover evolution in the Western Tatra Mountains, Slovakia. Biologia, 64 (3), 594–599.

HU, H. H., KREYMBORG, L. R., DOEING, B. J., BARON, K. S., JUTILA, S. A. (2006): Gridded Snowmelt and Rainfall-Runoff CWMS Hydrologic Modeling of the Red River of the North Basin. Journal of Hydrology Engineering, 11, 91–100.

GAREN, D. C., MARKS, D. (2005): Spatially distributed energy balance snowmelt modelling in a mountainous river basin: estimation of meteorological inputs and verification of model results. Journal of Hydrology, 315, 126–153.

JAIN, S. K., THAKURAL, L. N., SINGH, R. D., LOHANI, A. K., MISHRA, S. K. (2011): Snow cover depletion under changed climate with the help of remote sensing and temperature data. Natural Hazards, 58 (3), 891–904.

JANSKÝ, B., ENGEL, Z., KOCUM, J., ŠEFRNA, L., ČESÁK, J. (2011): The Amazon River headstream area in the Cordillera Chila, Peru: hydrographic, hydrological and glaciological conditions. Hydrological Science Journal, 56 (1), 138–151.

JENÍČEK, M., TAUFMANNOVÁ, A. (2010): Vliv vegetace na akumulaci a tání sněhu – výběr z výsledků výzkumu Katedry fyzické geografie a geoekologie PřF UK v letech 2009 a 2010. In: JIRÁK, J. et al. (eds.): XV. Medzinárodné stretnutie snehárov (43–51). Praha, ČHMÚ.

JENÍČEK, M. (2009): Runoff changes in areas differing in land-use in the Blanice River basin – application of the deterministic model. Journal of Hydrology and Hydromechanics, 57 (3), 154–161.

JENÍČEK, M., KOCUM, J., JELÍNEK, J. (2008): Monitoring sněhové pokrývky v povodí Rokytky v letech 2007 a 2008. In: BROŽA, V., SZOLGAY, J., FOŠUMPAUR, P. (Eds.): Extrémní hydrologické jevy v povodích (227–236). Praha, ČVÚT.

JOST, G., MOORE, R. D., WEILER, M., GLUNS, D. R., ALILA, Y. (2009): Use of distributed snow measurements to test and improve a snowmelt model for predicting the effect of forest clear-cutting. Journal of Hydrology, 376 (1–2), 94–106.

JOST, G., WEILER, M., GLUNS, D. R., ALILA, Y. (2007): The influence of forest and topography on snow accumulation and melt at the watershed-scale. Journal of Hydrology, 347 (1–2), 101–115.

KLIMENT, Z., MATOUŠKOVÁ, M., LEDVINKA, O., KRÁLOVEC, V. (2011): Trend analysis of rainfall-runoff regimes in selected headwater areas of the Czech Republic. Journal of Hydrology and Hydromechanics, 59 (1), 36–50.

KOIVUSALO, H., KOKKONEN, T. (2002): Snow processes in a forest clearing and in a coniferous forest. Journal of Hydrology, 262, 145–164.

KUČEROVÁ D., JENÍČEK, M. (2011): Srovnání interpolačních metod z hlediska kvality předpovědi výšky a vodní hodnoty sněhu v neznámých bodech. In: HOLKO, L. (Ed.): XVI. Medzinárodné stretnutie snehárov (48–55). Liptovský Mikuláš, Ústav hydrológie SAV.

KUTLÁKOVÁ, L., JENÍČEK, M. (2012): Modelování akumulace a tání sněhu v povodí Bystřice v Krušných horách. Geografie, 117 (1), 110–125.

KUUSISTO, E. (1980): On the values and variability of degree-day melting factor in Finland. Nordic Hydrology, 11, 235–242.

LLOYD, C. D. (2005): Assessing the effect of integrating elevation data into the estimation of monthly precipitation in Great Britain. Journal of Hydrology, 308, 128–150.

LÓPEZ-MORENO, J. I., NOGUÉS-BRAVO, D. (2006): Interpolating local snow depth data: an elevation of methods. Hydrological Processes, 20, 2217–2232.

MAGNUSSON, J., FARINOTTI, D., JONAS, T., BAVAY, M. (2011): Quantitative evaluation of different hydrological modelling approaches in a partly glacierized Swiss watershed. Hydrological Processes, 25, 2071–2084.
MARTINEC, J. (1977): Snowmelt hydrography from spatially varied input. In: Proc. Third International Hydrology Symposium (110–111). Fort Collins.
OHARA, N., KAVVAS, M. L. (2006): Field observations and numerical model experiments for the snowmelt process at a field site. Advances in Water Resources, 29, 194–211.
OHMURA, A. (2001): Physical Basis for the Temperature-Based Melt-Index Method. Journal of Applied Meteorology, 40 (4), 753–761.
ÖZDOGAN, M. (2011): Climate change impacts on snow water availability in the Euphrates-Tigris basin. Hydrol. Earth Syst. Sci., 15, 2789–2803.
POBŘÍSLOVÁ, J., KULASOVÁ, A. (2000): Ukládání a tání sněhu v lese a na odlesněných partiích Jizerských hor. Opera Corcontica, 37, 113–119.
RANGO, A., MARTINEC, J. (1995): Revisiting the degree-day method for snowmelt computations. Water Resources Bulletin, 31, 657–669.
SENSOY, A., SORMAN, A. A., TEKELI, A. E., SORMAN, A. U., GAREN, D. C. (2006): Point-scale energy and mass balance snowpack simulations in the upper Karasu basin, Turkey. Hydrological Processes 20, 899–922.
SINGH, P., SINGH V. P. (2001): Snow and Glacier Hydrology. London, Kluwer Academic Publisher.
STEIGER, R. MAYER, M. (2008): Snowmaking and Climate Change. Mountain Research and Development, 28 (3), 292–298.
STRASSER, U. MARKE, T. (2010): ESCIMO. Spread – a spreadsheet-based point snow surface energy balance model to calculate hourly snow water equivalent and melt rates for historical and changing climate conditions. Geosci. Model Dev., 3, 643–652.
TANASIENKO, A. A., YAKUTINA, O. P., CHUMBAEV, A. S. (2009): Snowmelt runoff parameters and geochemical migration of elements in the dissected forest-steppe of West Siberia. Catena, 78 (2), 122–128.
TEUTSCHBEIN, C., WETTERHALL, F., SEIBERT, J. (2011). Evaluation of different downscaling techniques for hydrological climate-change impact studies at the catchment scale. Climate Dynamics, 37, 2087–2105.
UHLMANN, B., GOYETTE, S. BENISTON, M. (2009): Sensitivity analysis of snow patterns in Swiss ski resorts to shifts in temperature, precipitation and humidity under conditions of climate change. International Journal of Climatology, 29, 1048–1055.
VÁŇOVÁ, V., LANGHAMMER, J. (2011): Modeling the impact of land cover changes in flood mitigation in the upper Lužnice basin. Journal of Hydrology and Hydromechanics, 59 (4), 262–274.
VERBUNT, M., GURTZ, J., JASPER, K., LANG, H., WARMERDAM, P., ZAPPA, M. (2003): The hydrological role of snow and glaciers in alpine river basins and their distributed modeling. Journal of Hydrology, 282, 36–55.
WALTER, M. T., BROOKS, E. S., MCCOOL, D. K., KING, L. G., MOLNAU, M., BOLL, J. (2005): Process-based snowmelt mode-ling: does it require more input data than temperature-index modeling? Journal of Hydrology, 300 (1–4), 65–75.
WEIGERT, A., SCHMIDT, J. (2005): Water transport under winter conditions. Catena, 54, 193–208.

RÉSUMÉ

Modelování akumulace a tání sněhu – přehled současných přístupů a výsledků

Tání sněhu způsobené vysokými teplotami vzduchu doprovázené kapalnými srážkami je v České republice častou příčinnou vzniku povodně. Cílem příspěvku je popsat hlavní procesy, které ovlivňují odtok ze sněhové pokrývky a poskytnout přehled o metodách, které jsou v současné době používány pro modelování akumulace a tání sněhu. Pokrok v pochopení procesů probíhajících ve sněhové pokrývce dokumentují jednak četné zahraniční výzkumy, jednak výzkumy, které byly provedeny autory textu na malých experimentálních povodích v horských oblastech České republiky a jejichž souhrn je v textu uveden. Výzkum autorů se zaměřuje na 1) měření charakteristik sněhové pokrývky a analýzu její prostorové a časové distribuce, 2) hodnocení role vybraných fyzicko-geografických faktorů na akumulaci a tání sněhu (sklon a expozice svahu, typ vegetace), 3) testování metod prostorové interpolace bodově měřených charakteristik sněhu a 4) modelování akumulace a tání sněhu v lokálním a regionálním měřítku.

Ve sledovaných povodích ukazují hlavní výsledky výzkumu na rozdílné chování sněhové pokrývky akumulované v lese a na otevřených plochách, kdy v lese dochází k nižší akumulaci sněhové pokrývky a také k jejímu pozvolnějšímu tání než v případě otevřených ploch. Pro interpolaci bodových měření sněhu se ukazuje jako nejvhodnější použití metod, které do výpočtu neznámé závislé proměnné (výšky sněhu, vodní hodnoty sněhu) zahrnují jednu nebo více nezávislých proměnných, například nadmořskou výšku. Dosažené výsledky také potvrdily známé problémy s aplikací metody teplotního indexu (degree-day), především obtížnost modelování vodní hodnoty sněhu a odtoku ze sněhu v podmínkách teploty vzduchu kolísající kolem 0 °C. Problémy také nastávají při modelování denního chodu tání sněhu způsobeným vysokým úhrnem krátkovlnného záření a výrazným denním chodem teploty vzduchu (přes den kladná teplota vzduchu, v noci záporná teplota vzduchu). V těchto případech by bylo přínosnější použití metod založených na výpočtu energetické bilance sněhu. Metoda degree-day naopak uspokojivě modeluje situace s pozvolným přibýváním sněhu bez častých oblev a následné tání sněhu způsobené vysokými teplotami vzduchu a kapalnými srážkami.

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