Nordic Hydrology, 31 (4/5), 2000, 267-286

Model Sophistication in Relation to Scales in Snowmelt Runoff Modeling

Paper presented at the 12th Northern Res. Basins/Workshop
(Reykjavik, Iceland – Aug.23rd -27th 1999)

Lars Bengtsson
Water Resources Eng., Lund University, SE-22100, Sweden

Vijay P. Singh
Lousiana State University, Baton Rouge, LA 70803, U.S.A.

Snowmelt induced runoff from river basins is usually successfully simulated using a simple degree-day approach and conceptual rainfall-runoff models. Fluctuations within the day can not be described by such crude approaches. In the present paper, it is investigated which degree of sophistication is required in snow models and runoff models to resolve the basin runoff from basins of different character, and also how snow models and runoff models must adapt to each other. Models of different degree of sophistication are tested on basins ranging from 6,000 km^2 down to less than 1 km^2. It is found that for large basins it is sufficient to use a very simple runoff module and a degree day approach, but that the snow model has to be distributed related to land cover and topography. Also for small forested basins, where most of the stream flow is of groundwater origin, the degree-day method combined with a conceptual runoff model reproduces the snowmelt induced runoff well. Where overland flow takes place, a high resolution snow model is required for resolving the runoff fluctuations at the basin outlet.

Introduction

The snow is unevenly distributed within a basin and also the atmospheric heat exchange between snow and atmosphere is spatially variable within a basin. Different types of heat exchange processes are important in different environments. To which extent the uneven spatial snow distribution and snowmelt should be accounted for, when coupling snow models and runoff models, depends on which spatial and temporal resolution that is required for a certain study, and over how large area the output in the form of runoff, evaporation and radiation balance is interpreted. The relevant hydrological temporal and spatial scales are not necessarily the same as for rain produced runoff; nor are the best model structure and required degree of model sophistication the same for snowmelt induced runoff as for rain.

In regions with much snow, as in northern Sweden, much of the annual runoff occurs in relation to snowmelt. In large rivers there is usually a steady rise of the flow through the melt period. There is a relation between snow storage and total runoff, and in large river basins also a relation, although not a very clear one, between the amount of snow stored in a basin and the peak river flow, indicating that the response time of melt conditions is quite long and the time and spatial scales representative of the peak flows are quite large. In small basins the response time is much shorter and there are fluctuations in the stream flow during the melt period. In agricultural basins, and even more so in urban basins, overland flow is common and the melt peak at noon is felt later in the afternoon as pronounced peak flow.

This paper starts with a discussion on winter snow distribution in a basin due to climatic variations, topography, vegetation, wind drift, and, in urban basins, snow removal. It continues with the different timing of melt within a basin, and the temporal distribution of melt at small scale. These introductory sections lead to definition of temporal and spatial scales in snow hydrology. Thereafter, different snow models and runoff models are described with emphasize on resolution in time and space depending on input. Models are combined in different ways and applied on small and large basins of different character. It is concluded which model combinations that are required for resolving observed stream flow fluctuations; it is also concluded which snow model should be used to match the resolution of existing hydrological models.

Snow Distribution and Spatial Scales

The snow is unevenly distributed in a river basin. The snow distribution is a function of variation of climate, topography and land use within the river basin, which all can be related to spatial scales. Spatial scales, the length of an element or the distance over which a process is considered, in snow hydrology span from the mm-size of the snow grain to the region scale of 100 km up to 1,000 km. On the regional scale the snow conditions are determined by the climate. An example is the large northern Swedish rivers, where it is colder and where there is much more precipitation in the western upstream mountains and highlands than in the forested downstream lowlands to the east. The difference in snow cover is even more pronounced in spring, since melts occurs earlier in the lowland areas than in the upstream parts of the basins.

The topography within a basin changes within a scale smaller than the basin scale. Uneven snow distribution is attributed to variations in the topography. Precipitation increases with height. Temperature decreases with height, which means that precipitation falling as rain in valleys may fall as snow higher up on hills and in mountains. The mid-winter snow distribution in the Kultsjön river basin was demonstrated by Brandt (1986). The snow water equivalent varied from less than 200 mm to more than 500 mm. The large variation is because the basin is rather mountainous with elevation ranging from 500 to 1500 m. The trend is consistent with a precipitation increase of 10-15% per 100 m as suggested by The Swedish Committee on Spillway Design (1989). Temperature and precipitation variations can be pronounced within a few km.

In low land areas the snow distribution is much more homogeneous, but land cover and land use influence the snow distribution. Snow that falls in forested areas is intercepted on the trees and some of this intercepted snow evaporates, often resulting in less snow in forested areas than on open fields. Because of wind turbulence during snowfall, more snow falls in forest openings than in the forests themselves. From data from Norway, Finland and Sweden, summarized by Brandt (1986), it seems that the amount of snow accumulated in coniferous forest openings is slightly more than that on open fields, but the snow accumulated in forests is less, 90-95%, than on open fields. One of the most thorough studies is probably the one by Mustonen (1965). He found little difference between the snow cover in forests and on open land in Finland. However, in forest openings there was about 10% more snow than in the forests.

The river response is much related to the drainage density and thus to the hill slope scale; the runoff response of small streams is quickly transferred to the rivers, if there are no lakes in the river system. Snow melts at different rates on open fields and in forests; and differently on north and south facing slopes. Woo (1998) has recently, much based on measurements, discussed various spatial scales in snow hydrology. He stresses the importance of dividing even small basins into elements, which experience similar snow depositional and scouring conditions.

Open fields are of the order maybe 100 m -1 km in scale; the length of mild slopes only of the order 100 m. The flow-paths towards draining streams may be even shorter. Within a field, the snow is redistributed by wind (Pomeroy *et al.* 1997) and in urban environments also by man (Bengtsson and Westerström 1992). Snow within urban areas is removed from asphalted surfaces and piled up near-by, or the snow is removed completely from downtown areas and dumped at snow deposits. Man is influencing the character of the snow by walking on it, putting dust and salt on it; fall-outs from cars and activities within the city accumulate in the snow; all of it resulting in compacted snow, accelerated metamorphosis and decreased albedo. The within field scale is 10-100 m, or less within urban areas (Semádeni-Davies 1999).

Before all the snow has disappeared from fairly homogeneous areas as a forest or an open field, there are patches with bare ground. These patches affect the energy

Table 1 – Spatial scales and related variables and snow processes.

Length scale	Variable condition	Dependent process
100 –1000 km	Climate	Snow accumulation / ablation
10 – 100 km	Topography	Snow and melt distribution
1 – 10 km	Topography, Land cover	Snow and melt distribution
100 m – 1 km	Land cover	Melt distribution in time
10 – 100 m	Hill slope, radiation exposure	Radiation, Melt distribution
1 – 10 m	Snow patchiness	Radiation
0.1 – 1 m	Snow depth	Duration of melt
1 – 10 cm	Ice lenses	Percolation through the snow
0.1 – 1cm	Snow grains	Radiation

balance on a small scale basis (Liston 1995; Marsh and Pomeroy 1996; Shook and Gray 1996), and also the overland flow-paths (Woo 1998). In nature such conditions exist only for a few days, when the mean snow cover is some cm, but are common through-out the melt period in urban areas. Such a micro scale or patch scale is of the order meters. The snow depth as such, of scale 1 m, is a scale determining the amount of snow and the delay between surface melt and meltwater appearance at the base of the snow pack.

The scales to be considered concerning snow accumulation and daily snowmelt related to basin runoff are in the range 1 m – 100 km. In Table 1 spatial scales are related to variable conditions and to snow processes in a synthesized way. Also shorter scales than m is included in the table, since shorter spatial scales are relevant for the surface melt and detailed resolution of meltwater percolation through the snow pack. After some melt-refreezing cycles ice lenses develop in the pack, and so do zones with high permeability (Marsh and Woo 1984). The meltwater tends to percolate through the snow within these high permeability zones (Colbeck 1975). These zones may extend over some 10 cm. The smallest spatial scale is the mm snow grain size. The grains grow in the course of the winter because of metamorphous effects. The grain size affects the snow albedo, and thus the absorption of solar radiation; it also affects the capability of the snow pack to hold liquid water.

Snow Processes and Temporal Scales

Once snowmelt has begun, there is a large difference in snow cover between different parts of a basin. The weather may still be cold and full winter in high altitude areas, when melt begins in the lowlands of a large basin. Melt proceeds at a faster rate on areas exposed to solar radiation than on shaded areas. Thus, melt begins earlier and proceeds faster in open areas as compared to forested areas. The snow melts faster on south facing slopes than on north facing ones. In an urban environment, the

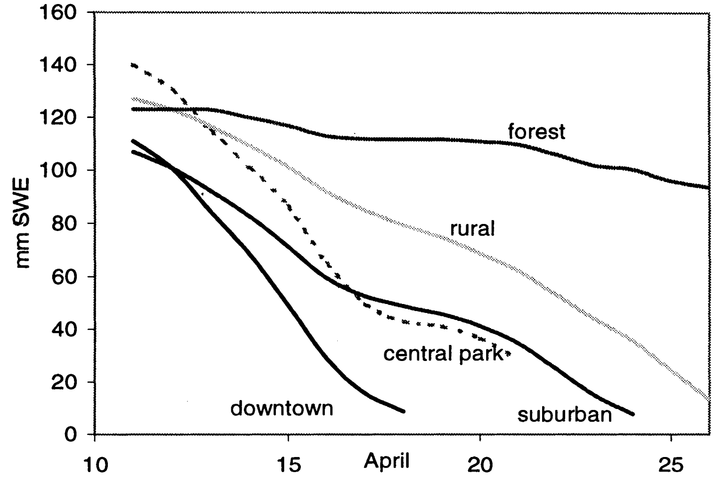

Fig. 1. Snow water equivalent versus time at nearby sites, Luleå, Sweden, 1980.

snow cover is, first of all, extremely heterogeneous; within a couple of meters, there are snow piles, almost untouched snow and snow free parts. Secondly, also the exposure to radiation is heterogeneous; nearby parts are shaded by buildings or receives extra radiation from buildings (Xu and Buttle 1987; Semådeni-Davies and Bengtsson 1998). In small basins the timing of melt is very different within the basin. Some examples from Luleå, Sweden, are shown in Fig. 1; all the sites shown in the figure are within 4 km distance; while the snowmelt on open rural field was distributed over more than 20 days, the urban snow disappeared within a 10-day period. The snow in a forest outside the city remained for a long time, since this snow was not exposed to solar radiation and because there was a cold spell soon after the open fields were free from snow.

Processes over different time scales are of importance for the generation of melt induced runoff, ranging from the short times over which the radiation at the snow surface varies to the duration of the snow cover. According to Blöschl and Sivapalan (1995) the characteristic time of a process may be considered a time scale. The energy exchange between the snow surface and the atmosphere varies over the day. When melt occurs, the surface melt rate is highest during mid-day and recedes towards the evening, often so that the snow surface temperature drops below 0°C during night. The surface melt is different from one hour to another; in fact it can vary from minute to minute, when the sun is temporary shaded by clouds.

Before runoff can take place, the meltwater must percolate through the snow pack. Water can not leave the pack until the liquid content is above the water holding capacity of the snow (at least in vertical drains). When the snow is dry, the percolation through the pack may take considerable time. When the snow is ripe, the melt flux moves much faster. After some warm days and cold nights and repeated refreezing of the liquid water within the snow, ice lenses and vertical high permeable drains develop. The meltwater preferably moves in these vertical drains at a rather high progression rate (Bengtsson 1982a; Marsh and Woo 1985). Increased surface melt is usually recognized at the bottom of the snow pack within an hour. Martinec (1985) reported 2-hour lag time between surface melt and runoff from a rather deep alpine snow cover. Thus, for the meltwater input at the bottom of a snow pack, the characteristic time of the process including the surface melt process is one or a few hours.

When meltwater reaches the bottom of the snow pack, it may infiltrate or run off along the base of the snow pack. The saturated flow at the bottom of the snow pack is either Darcian flow or faster flow in horizontal high permeable zones or even in channels, where the meltwater flow has eroded the snow. A theoretical approach is given by Singh *et al.* (1997). For surfaces extending over some 10's of m, it seems that the time from high surface melt to high runoff including short overland flowpaths should be one or a few hours.

The meltwater, which reaches the bottom of the snow pack, moves towards the basin outlet as overland flow or, after infiltration and percolation to the groundwater, as groundwater; or partly as overland flow and partly as groundwater. The runoff from a river basin is generated by melt integrated over time and space. The runoff processes are not directly related to snow, but the conditions during snowmelt are different from the conditions during rain storms (*e.g.* Espeby 1990; and Barry *et al.* 1990). Kane and Stein (1983) pointed out the influence of the moisture content prior to ablation on the storage capacity of the frozen soil and thus on the infiltration capacity and production of overland flow. The ground is frozen, so the infiltration capacity is to a higher or lesser extent reduced. There is a rapid hydraulic coupling of groundwater recharge in upland areas and downhill groundwater discharge and consequent stream response (Thorne *et al.* 1998). Groundwater levels are usually very high during the latter periods of snowmelt; saturated overland flow near streams is common.

The distribution of melt induced runoff at the basin outlet over the melt period and within individual days depends much on the size and character of the basin and on how the stream flow is generated. Where overland flow occurs, there are pronounced stream flow peaks in the late afternoon and low flows in the early mornings, as is shown in Fig. 2 for the Hovi agricultural field in southern Finland, as described by Bengtsson *et al.* (1992). The runoff from Hovi was initiated on 1 April, peaked every afternoon with a highest value of almost 20 mm/d on 6 April. Most of the field was free of snow on 10 April. The overland flow travel time is about 16 hours, Bengtsson *et al.* (1992). The characteristic time of overland flow in the snow may thus be fractions of a day. Rain in summer and autumn does not produce overland flow from the Hovi basin. Thus, the flow generation process is different in summer and in spring.

In urban areas, the runoff response to convective storms, when rain falls on im-

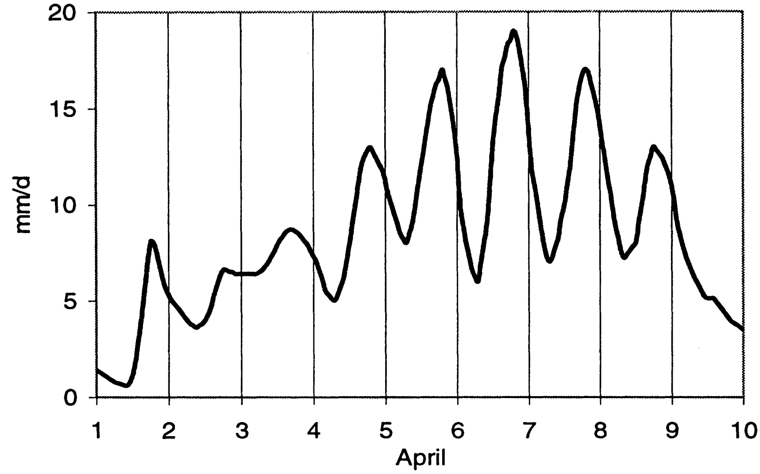

Fig. 2. Observed runoff from Hovi Agricultural Watershed, 0.12 km^2, Finland, April 1988.

permeable surfaces, is high and quick. Even in basins extending over several km^2, the peak flow lasts only a few minutes and the flow declines to low values within less than an hour. When the runoff is caused by snowmelt, and also when the melt runoff is increased by rain on snow, the percolation process within the snow pack delays and attenuates the runoff peak and distributes it over a longer time. Bengtsson and Westerström (1992) did not find much variation within the hour of the runoff from a 25 m long asphalted plot. The characteristic time of melt induced urban runoff is the same as for the meltwater input to the base of the snow pack, of the order hour.

In forested basins the melt induced runoff water is dominated by pre-event water existing as sub-surface water prior to the melt event, as found in many studies (*e.g.* Dincer *et al.* 1970; Rodhe 1981; Buttle and Sani 1992). The runoff produced is a consequence of melt over several previous days. The runoff variations within the day, even in small basins, are minor and masked on a rising hydrograph. In large rivers there are no clear discharge fluctuations between night and day during snowmelt, although the continuous discharge increase is somewhat interrupted during nights. When analyzing snowmelt induced hydrographs from Swedish rivers having basins of 1,000 km^2 to 6,000 km^2, disregarding all processes behind the river flow generation, the recession time coefficient was found to correspond to a recession time coefficient of several days. Therefore the melt of individual days is not clearly recognized in the river flow, Bengtsson (1986). In a similar way Robinson and Sivapalan (1997) related time scales to hydrological regimes by comparing the recession time coefficient with rain storm duration.

To shortly summarize the discussion on relevant time scales in snow hydrology: the runoff at a point varies depending on snowmelt variations within the day. In small basins where overland flow takes place snowmelt variations within the day cause fluctuations in the basin runoff suggesting that the melt-runoff process must be studied on a time scale shorter than a day. In small as well as large river basins, where the runoff production is mainly as sub-surface flow, the characteristic time of the melt induced runoff process is days. In the next part of the paper, after a short review of snow models and runoff models, the effect of knowing the snowmelt with different time resolution on computed basin runoff is investigated; computed and observed runoff are compared.

Snow Models

Snow models include parts for determining 1) surface melt, 2) internal processes, 3) infiltration into frozen soils and, if overland flow takes part, 4) horizontal flow in the bottom of the snow pack along short pathways leading to nearby streams. The surface melt can be determined from energy balance (*e.g.* Anderson 1968) in which the balance either is determined for the whole pack, or, in more advanced full snow models for the snow surface only (*e.g.* Morris 1983). Morris (1982) compared a full snow model with vertical resolution with a bulk snow pack model and a temperature index approach. Given correct radiation data she found that energy balance computations were superior to temperature index computations when determining snowmelt at a point, but that the vertically distributed snow model did not perform better than a bulk snow pack model. Blöschl and Kirnbauer (1991) reached similar conclusions. However, since the radiative part of the energy balance is the difference between large fluxes, which are not known with accuracy, a non-critical use of the energy approach for determining surface melt may result in large errors. A comparison between different models organized by the World Meteorological Organization (1986) showed that in large basins, simple models based on temperature index for snowmelt and storage relations for runoff performed better than more advanced mathematical models. Therefore, a simplified but robust approach, a temperature index model, is often used; the snowmelt is simply calculated as proportional to the air temperature. When there is little wind and little sunshine, as in a coniferous forest, the temperature index approach is theoretically sound (Bengtsson 1976).

The global radiation is correlated with the air temperature. The snowmelt takes place during similar atmospheric conditions each year. Therefore, the melt can be estimated as proportional to the air temperature with a proportionality factor, degree-day factor, which does not vary so much from year to year for the same location and the same period of the year. However, the degree-day approach does not give the correct melt of individual days. Determination of the mean surface melt over single 24-hour periods requires that radiation is included in the computations as shown by Bengtsson (1986), Sand (1990), and Granger and Gray (1990). Surface melt varies with the solar radiation, and thus varies much during the day. Hourly variation of snowmelt can be determined only if the radiation is explicitly considered.

The meltwater produced at the snow surface percolates through the snow pack. The percolation process is a slow process, when the snow is initially dry, but when the snow is ripe and when ice lenses and high-permeable zones have developed in the snow, the meltwater travels from the snow surface through a not very thick snow cover to the ground within an hour or so, as was discussed in the section on time scales. This means that unless runoff variations shorter than an hour are described in the runoff model, it is sufficient to account for the internal process of refreezing of meltwater in a snow model used in combination with a runoff model, and not considering the percolation dynamics in detail.

The third part of snow models, the distribution of the meltwater, which reaches the base of the snow pack, between infiltration and overland flow, is included in the runoff module of hydrological models, but this distribution is different for snowmelt and rain. The infiltration is reduced or completely hindered in frozen soils (Kane and Stein 1983; Granger *et al.* 1984; Engelmark 1984), and the overland flow is slower along the ground at the base of the snow pack than on snow free areas.

The conclusion of this short section on snow models is that when the characteristic time of the runoff process is some hours to a day, the snow model must be such that energy balance computations are performed and the refreezing process within the snow pack accounted for. When it is sufficient to know the mean snowmelt over a period of some days, the melt can be calculated with a degree-day approach.

Runoff Models

Meltwater reaching the ground is the input to runoff models. Rainfall-runoff models may be physically based models, conceptual models, or empirical ones. The rational method is an example of the latter. Since physically based models, as well as conceptual models, need to be calibrated, the difference between physically based models and conceptual models is in practice minor, at least when the conceptual models are distributed. Rainfall-runoff models consist of a module for the hydrological processes within a basin including infiltration, overland flow, groundwater recharge, sub-surface flow to small streams, and of a module for the hydro-dynamics of rivers and lakes. In urban runoff models and also in some physically based rural models, the hydro-dynamic part of the model is very advanced including back-water effects and flow variations within minutes or even seconds. However, when applied in large basins, the storage effect of the rivers and the lakes is the most essential feature of the river system hydraulics module.

The hydrological part of a physically based runoff model is based on a groundwater model using the Darcy law, on percolation computations using the Richards equation, and overland flow computations using the kinematic wave. Differentiating these equations using only few spatial steps, the physically based hydrological models are similar to conceptual models, in which output usually is related to the stor-

	Snow process	Runoff process	Type of basin	Snow model	Runoff model
Month	Accumulation				Water balance
	Snow drift Melt period				
Week					
		Groundwater	Very large	Temp-index	Conceptual model
Day	Melt intensity variation	Overland flow	Small forested Agricultural	Energy balance	Physically based model
Hour	Meltwater percolation	Urban overland flow	Urban	Percolation	
Minute	Solar radiation variation	Conduit flow			Dynamic wave model

Fig. 3. Relations between temporal scales, type of river basin, processes and model resolution.

ages within the model, *i.e.* the state of the model variables. In these conceptual models, the basin or sub-basin values are integrated and given as point values. The conceptual hydrological models can be distributed down to small sub-basins, which are connected with river and lake routing in a hydro-dynamic module.

There are many processes described in a hydrological model and many parameters controlling the processes. The degree of sophistication in a snow model should correspond to the degree of sophistication in the hydrological model. In Fig. 3 time scales, snow and runoff processes are related to each other and compared with the resolution of different models, summarizing the discussion above. In a hydrological model with coarse time resolution, a snow model with high time resolution is not required. However, if the river flow from a basin varies much over the day, a model with high resolution of snowmelt as well as of the runoff generation is required to reproduce these variations. Still, depending on the basin conditions during snowmelt, some hydrological processes do not need to be considered and the hydrological model can be simplified. In the urban runoff models of today, the hydrological and hydro-dynamic modules operate on a minute time base, for example the MOUSE-

NAM model, Hernebring *et al.* (1997), but the snow models used in these urban models only give a mean daily melt. The snow model is not adapted to the runoff model in such a way that the resolution of the input matches the output of the urban runoff model. In fact the hydrological and hydraulic processes in urban basins are so much faster than the snowmelt related processes that they do not need to be considered: over the time scales relevant for the snowmelt process, all meltwater which reaches impermeable ground also reaches the basin outlet.

Effect of Snowmelt Resolution on Runoff Computations

So far the discussion on the effect of snowmelt resolution on runoff has been restricted to runoff observations and on travel time analysis. In this section, snowmelt with different time resolution is used as input to runoff models to see the effect on the output. The objectives are to determine which snowmelt time resolution is required to obtain optimal output resolution from differently sophisticated runoff models and to find out which spatial and temporal resolution is required in snow models and runoff models to reproduce snowmelt induced runoff from different kinds of river basins. When daily values are sufficient, it remains to determine how well this daily melt must be known for the runoff to be computed with good correspondence with observations. In a previous study (Bengtsson 1986), it has been shown that a snowmelt resolution of some days is sufficient to reproduce the melt induced runoff, when the basin recession time coefficient is more than a few days.

A physically based hill slope model including sub-surface flow and overland flow and a conceptual model are applied to basins of different character. The conceptual model used is the rainfall-runoff model of the Department of Water Resources Engineering, Lund University, the FLAPS-model (Bengtsson *et al.* 1995), which is similar to the in the Nordic countries frequently used HBV-model (Bergström 1992). It includes kinematic river and lake routing.

It has been found that the characteristic time of the meltwater input to the base of the snow pack is hours. In urban basins the runoff process is fast. The runoff distribution in time is close to the distribution of melt at the base of the snow pack. Thus, the urban runoff is well reproduced if the melt is. To resolve the melt from hour to hour an advanced snow model is required, but once the melt is known, the urban runoff model can be extremely simple. Spatial distribution is required when computing the melt, since the snow cover and melt intensity vary over short distances (Buttle and Xu 1988; Bengtsson and Westerström 1992). However, since the runoff process does not need to be considered in detail, the spatial distribution can be restricted to determining the fractional area of different snow distributions, snow character and sun exposure (Semádeni-Davies 1999).

Except for urban areas, runoff processes seldom need to be considered over so short time as hours, when describing the basin runoff. Storage on the ground, in the soil and in the groundwater as well as in small ponds and lakes attenuates the effect of temporary and spatially high precipitation and snowmelt input. If there are only small flow variations within a day, it might be that the flow can be reproduced without detailed resolution of the snow processes. In fact, with most runoff models it is not possible to reproduce the runoff process on an hourly basis.

The stream flow generated from small (up to a few km^2) rural catchments dominated by open fields as meadows and agricultural land show variations over the day. Bengtsson (1982b) and Engelmark (1984) showed that after infiltration of the early meltwater, most of the meltwater from an open field at Bensbyn Research Watershed in Luleå, Sweden, moves as overland flow at the base of the snow pack or in cracks between the vegetation just below or above the ground surface. Melt intensity variations over the day influence the basin runoff.

The runoff from the Bensbyn open field was computed using a physically based hill slope model with routines for groundwater flow and overland flow, and with one hour resolution output. The meltwater input was computed from energy balance computations based on measurements at a site 4 km from the Bensbyn field. When the computed runoff with 1-hour resolution also on the melt input is compared with the observed stream flow from the Bensbyn watershed there is good agreement, as shown in Fig. 4, except for April 23. This day the observed runoff was high relative the computed melt and the high flow lasted well into the night. It seems that the melt was not well simulated for that day.

To test the importance of the resolution of meltwater input in the runoff model, 12-hour mean meltwater input to the base of the snow pack was used as model input. As seen in Fig. 4 the peaks and lows during the day are not reproduced very well with the 12-hour mean input. The comparison shows that a runoff model, which resolves fast hydrological processes, also requires meltwater input with high resolution. Runoff fluctuations within the day can of course not be simulated with daily melt as input.

Most often conceptual models based on relations between storage and runoff are used for computing runoff from rural basins. In principle the groundwater flow is well described with these models, but the percolation process is simplified and does not account for the time for the moisture flux to reach the groundwater. This means that conceptual runoff models are not likely to be able to reproduce stream flows, which vary over such short periods as hours. Thus, the runoff from the Bensbyn open field is not expected to be reproduced well with a conceptual model. Still, to test how well the runoff from the Bensbyn field can be described with a conceptual model, the FLAPS model was applied to Bensbyn using 1-12 hour resolution and treating the field as a point. It was found that it did not influence the output from the runoff model, whether the meltwater input was the hourly values or melt evenly distributed over 12 hours. The peaks and lows were flattened out compared to the observed ones, but the model performed as well as the physically based model with coarse meltwater input.

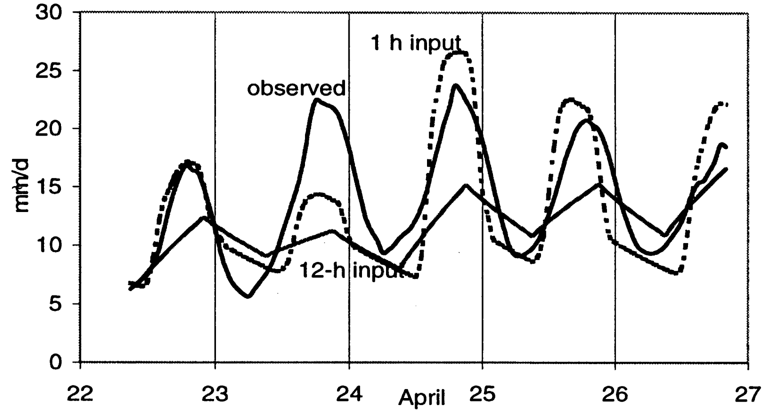

Fig. 4 Melt induced runoff from the Bensbyn field, April 1980, simulated with differently high temporal resolution on the meltwater input to a hill slope model, and observed runoff.

The stream flow in Scandinavian forests mainly originates from groundwater (Lundin 1982; Bengtsson *et al.* 1989). The groundwater moves towards ditches and small streams when entering the river drainage system. When the soil is frozen, percolation is much restricted to macro-pores (Roberge and Plamondon 1987; Espeby 1990), which means that the percolating water reaches the groundwater fast. There is a hydraulic connection between upstream groundwater and downstream in a small stream. The response in the stream to groundwater recharge depends on the groundwater level. Conceptual models are suitable for reproducing the runoff during such conditions. Model tests for forested basins were done on the Teereseunoja Research Watershed in southern Finland (Bengtsson *et al.* 1989). The watershed is 69 ha. Most of the area, 90%, is covered by forest dominated by Norwegian spruce. The observed basin runoff, including small evening peaks and low flows in the mornings, was well reproduced using the FLAPS-conceptual model with 12 hour resolution on the snowmelt input. Data were not available for energy balance computations of snowmelt. Instead a degree-day approach was used based on air temperature and calibrated *versus* melt from a snow lysimeter in an open area and *versus* snow surveys in the forest. The snow model included a routine for computing release and refreezing of liquid water. Although almost the entire basin is covered with forest, the timing of melt and runoff were reproduced better when the basin was differentiated into elements distinguishing between forest and open land than when it was not. The necessity of land cover separation even in very small basins has been discussed by Woo (1998).

High and low flows during a day can not be reproduced with input of longer resolution than 12 hours, but since the stream flow fluctuations during the day in the Teereseunoja stream are minor, also the daily flows give a representative picture of

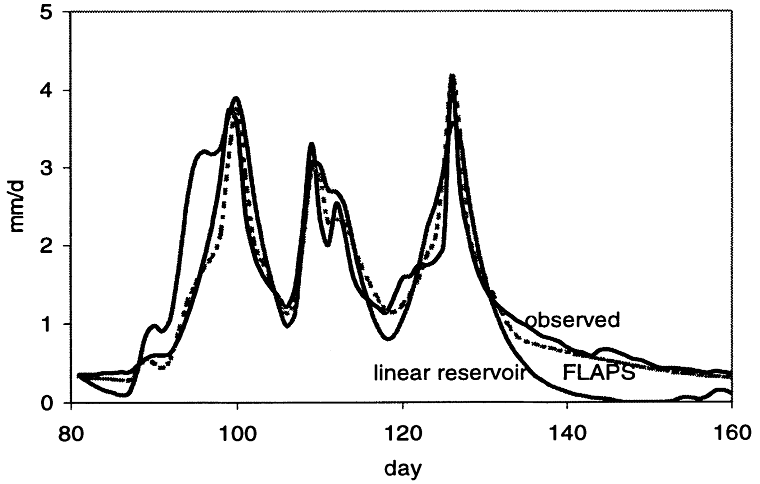

Fig. 5 Observed and simulated runoff using a complete conceptual model, FLAPS, and a single linear reservoir without soil routine, from Teereseunoja Research Watershed, 0.69 km^2, in Finland, 1986, with daily melt as input.

the Teereseunoja basin response to snowmelt. Using the daily melt as input into the conceptual model, the daily runoff was well reproduced, which is shown in Fig. 5 for a simulation extending over many melt episodes dividing the basin into open land and forest.

During snowmelt, the soil is frozen. There is little storage available in the unsaturated zone. The groundwater level is high and groundwater in the high-permeable layers near the ground surface dominates the contribution to the basin runoff. There is only little, if any, evapotranspiration. Such a snowmelt situation is much easier to describe in a model than a summer storm situation is. The meltwater more or less recharge the groundwater directly. The runoff model can be simplified to a single box relating runoff (q) to groundwater storage (h): $q = h / T$, T being a time recession coefficient. The snowmelt takes place at different rate from different sub-areas, but the meltwater is in the most simple model approach transferred directly into the single box representing mainly the groundwater storage in the basin. When such a simple runoff model was applied on Teereseunoja, the melt induced runoff, except for the very early melt, was computed almost as well at the runoff computed with a full conceptual model including soil water routines and base flow, as can be seen in Fig. 5, which as discussed earlier also shows the result of the complete conceptual FLAPS-model. The conclusion of this exercise is that very simple runoff models can be used to simulate snowmelt induced runoff from forested basins.

For the small Teereseunoja basin, melt resolution over a whole day was sufficient for modeling the runoff, and a crude simple runoff model could be used. Therefore, intuitively, single days of intense melt should not affect the river discharge from a

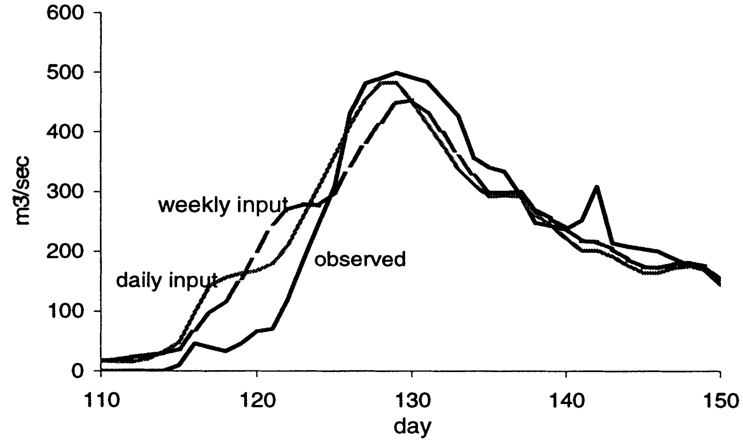

Fig. 6 Melt induced river discharge from the Vattudal basin, 3,851 km^2, 1986: observed river discharge, computed discharge with daily input, and with weekly averaged melt.

river basin of thousands of km^2 very much. In this paper the degree-day method approach over the full day and the FLAPS model as runoff model were applied on the large Swedish river basin Vattudal, about 4,000 km^2. When account was taken of the land cover and topography so that the whole basin was divided into sub-areas but still not connecting the different areas in a river drainage system, the basin runoff was well reproduced, as shown in Fig. 6. Input distributed over two days also gave good agreement with observations. Even when the mean melt over a full week was used as mean daily input, the simulated peak did not, as seen in Fig. 6, deviate much from the observed one, although it was too low.

For model simulations of high time resolution to be justified, good spatial resolution is also required. To see the difference in the model simulations when the land cover and topography distribution over the basin is accounted for or not, the spatial distribution of the Vattudal Basin in Sweden was eliminated and all zones were lumped; and the conceptual model was run as a point model. The output was a peaked hydrograph, since melt took place with the same rate from all the zones and the contribution to runoff occurred at the same time from all sub-areas. The peak runoff was 750 m^3/s as compared to the observed 500 m^3/s. It is more important to divide a large basin into zones of different melt conditions than to know the day to day weather variations.

It seems that the melt induced runoff from large river basin easily can be simulated with good result. It has been found in many studies over the years (*e.g.* World Meteorological Organization 1986, and Lindström *et al.* 1997) that snowmelt induced runoff from large river basins is well estimated using simple conceptual models with temperature index as the only snow model approach. Bergström (1996) concluded from vast experience of the HBV-model that simple snow and runoff models

Table 2 – Combination of models suggested to be used in river basins of different character.

Basin character	Time scale	Length scale	Snow model	Runoff model
Very large	Several days	> 100 km	Temp-index	Distributed conceptual
Small forested	Day	1 – 100 km	Temp-index	Distributed conceptual
Small agricultural	Hours	0.1 – few km	Energy balance	Physically based
Urban	Hour	< 1 km	Energy balance	Spatial melt distribution

can be used to model snowmelt induced floods in rivers, if the topography and vegetation are accounted for by separating the snow model into classes, and also accounting for refreezing within the snow pack.

An attempt to synthesize the discussion on scales related to different processes in different kinds of river basins and resolution of different models is made in Table 2. When overland flow occurs in snow, as may be the case on agricultural fields and on urban surfaces, the time scale for the runoff process is of the order hour. To model the melt with this resolution, the energy balance for the snow pack must be considered, and a physically based model including overland flow and groundwater dynamics must be used to describe the runoff process. When the runoff is mainly as groundwater as in forested basins or if the river basins are very large, a time resolution of day is sufficient allowing simple models to be used. In principle, it is sufficient to describe the slowest one of the snow and the runoff process in a model, since this process determines the runoff variations in time.

Conclusion

Models and data relevant for the scale over which the hydrological processes vary and influence the basin runoff should be used in runoff simulations. Depending on the hydrological regime, data and also models with different time and spatial resolution are required for simulating and analyzing melt induced basin runoff. The snow and runoff models should have the same resolution in time. In urban basins the characteristic time of the snowmelt and the meltwater percolation through the snow pack is longer than that of the overland and conduit flow. The snowmelt from different surfaces can be modeled separately and added together to form the basin runoff. The flow-paths on agricultural fields and meadows may be as overland flow. The snowmelt induced runoff from open fields varies during the day, which means that melt intensity variations during the day must be accounted for when modeling the basin runoff. This can be done with an energy balance approach for the snow pack accounting for refreezing of liquid water in the snow pack. Consequently, also the runoff model must have high resolution.

In forested catchments, where most of the stream flow is groundwater, knowing the snowmelt as average over a day is sufficient for good estimates of the stream flow. The snow model can be the degree-day method. A conventional conceptual model with runoff related to storage is well suited for simulating snowmelt induced runoff. Since the timing of the melt depends on the land cover, the fractional distribution of the land must be included in the modeling approach in order to simulate the peaks correctly. In large basins of thousands of km^2, a degree-day approach can be used as snow model provided the basin is distributed into elements depending on land use and topography The structure of the river system and the way the sub-areas are connected only has a minor influence on the basin runoff.

References

Anderson, E.A. (1968) Development and testing of snow pack energy balance equations, *Water. Resour. Res.*, *Vol. 4*, pp.19-37.

Barry, R., Prévost, M., Stein, J., and Plamondon, A.P. (1990) Simulation of snowmelt runoff pathways on the Lac Laflamme Watershed, *J Hydrology*, *Vol.113*, pp103-121.

Bengtsson, L. (1976) Snowmelt estimated from energy budget studies, *Nordic Hydrology, Vol. 7*, pp,3-18.

Bengtsson, L. (1982 a) Percolation of meltwater through a snow pack, *Cold Regions Sciences and Technology, Vol. 6*, pp.73-81.

Bengtsson, L. (1982) Ground- and meltwater in the snowmelt induced runoff, *Hydrol. Sci J.*, *Vol.27*, pp.147-158.

Bengtsson, L. (1986) Snowmelt simulation models in relation to space and time, *IAHS Publ. No.155*, *Modelling snowmelt-induced processes*, pp.51-59.

Bengtsson, L., and Westerström, G. (1992) Urban snowmelt and runoff in northern Sweden, *Hydrological Sciences J.*, *Vol.37*, pp.263-275.

Bengtsson, L., Iritz, L., Zhang, T., Berndtsson, R., and Niemczynovocz, J. (1995) A decision support system for watershed flood design purposes in China, Proc. Second Int. Conf. Hydro-Science and -Engineering, Vol. II. A, Tsinghua University Press, Beijing, pp.982-989.

Bengtsson, L., Lepistö, A., Saxena, R.K., and Seuna, P. (1989) Mixing of acid meltwater with groundwater in a forested basin in Finland. - *IAHS Publ.*179, *Atmospheric Deposition*, pp. 251-258.

Bengtsson, L., Seuna, P., Lepistö, A., and Saxena, R.K. (1992) Particle movement of meltwater in a subdrained agricultural basin, *J. Hydrology*, *Vol.135*, pp.383-398.

Bergström, S. (1992) The HBV model – its structure and applications, Swedish Meteorological and Hydrological Institute, Rep. Hydrology No. 4, Norrköping, Sweden.

Bergström, S. (1996) Modelling snowmelt induced flooding. Contribution to *EU RIBAMOD Concerted Action*, First Expert Meeting, Copenhagen, October 10-11.

Blöschl, G., and Kirnbauer, R. (1991) Point snow models with different degrees of complexity – internal processes. *J. Hydrology* , *Vol.* 129, pp. 127-147.

Blöschl, G., and Sivapalan, M. (1995) Scale issues in hydrological modeling: a review, *Hydrol. Proc., Vol.9*, pp, 251-290.

Brandt, M. (1986) Spatial snow studies, SMHI-Hydrology 7, pp. 52, Norrköping, Sweden (in Swedish).

Buttle, J. M.,and Xu, F. (1988) Snowmelt runoff in suburban environments, *Nordic Hydrology, Vol.18,* pp.19-40.

Buttle, J.M., and Sami, K. (1992) Testing the groundwater ridging hypothesis of stream flow generation during snowmelt, *J. Hydrology, Vol.135*, pp.53-72.

Colbeck, S.C. (1972) A theory of water percolation in snow, *J. Glaciol. Vol.11*, pp.369-385.

Dincer, T., Payne, B.R., Florkowski, T., Marinec, J., and Tongiorgi, E. (1970) Snowmelt runoff from measurements of tritium and oxygen 18, *Water Resour. Res., Vol.6*, pp.110-124.

Engelmark, H. (1984) Infiltration in unsaturated frozen soil, *Nordic Hydrology, Vol.15*, pp.243-252.

Espeby, B. (1990) Tracing the origin of natural waters in a glacial till slope during snowmelt, *J. Hydrology, Vol.118*, pp.107-127.

Granger, R.J., and Gray, D.M. (1990) A net radiation model for calculating daily snowmelt in open environments, *Nordic Hydrology, Vol.*21, pp.217-234

Granger, R.J., Gray, D.M., and Dyck, G.E. (1984), Snowmelt infiltration to frozen Praire soils, *Can. J. Earth Sci., Vol.21*, pp.669-677.

Hernebring, C., Marklund, S., and Gustafsson, L-G. (1997) Modeling of flow components in sewer systems influenced by snowmelt processes using the MouseNAM concept, Second DHI Software User Conf., Helsingör, Denmark.

Kane, D.L., and Stein, J. (1983) Water movement into seasonally frozen soils, *Water Resour. Res., Vol.19*, pp.1547-1557.

Lindström, G., Gardelin, M., Johansson, B., Persson, M., and Bergström, S. (1997) Development and test of the distributed HBV-96 hydrological model, *J. Hydrology, Vol.10*, pp.272-288.

Liston, G.E. (1995) Local advection of momentum, heat, and moisture during the melt of patchy snow covers, *J. Appl. Meteorol., Vol.17*, pp.1833-1842.

Lundin, L. (1982) Soil moisture and groundwater in till soil and the significance of soil type for runoff. UNGI rep. 56, Uppsala University, Dept. Phys. Geogr., 216 pp (in Swedish).

Marsh, P., and Woo, M-k. (1984) Wetting front advance and freezing of meltwater within a snow cover, 1. Observations in the Canadian Arctic, *Water Resour. Res., Vol.20*, pp.1853-1864.

Marsh, P., and Woo, M-k. (1985) Meltwater movement in natural heterogeneous snow covers, *Water Resour. Res., Vol.21*, pp.1710-1716.

Marsh, P., and Pomeroy, J.W. (1996) Meltwater fluxes at an Arctic forest-tundra site, *Hydrological Processes, Vol.10*, pp.1383-1400.

Martinec, J. (1985) Time in hydrology. In: *Facts of Hydrology,* ed.: J.C. Rodda, pp.249-290. John Wiley & Sons.

Morris, E.M. (1982) Sensitivity of the European Hydrological System snow models, *IAHS Publ. 138, Hydrological Aspects of Alpine and High Mountain Areas*, pp.221-231.

Morris, E.M. (1983) Modelling the flow of mass and energy within a snow pack for hydrological forecasting, *Annals Glaciol., Vol.4,* pp.198-203.

Mustonen, S. (1965) Effect of meteorological and terrain factors on water equivalent of snow cover and on frost depth, *Acta Forestalia Fennica, Vol.79*, pp.5-40, Helsinki (in Finnish).

Pomeroy, J.W., Marsh, P., and Gray, D.M. (1997) Application of a distributed blowing snow model to the Arctic, *Hydrol. Proc., Vol.11*, pp.1451-1464.

Roberge, J., and Plamondon, A.P. (1987) Snowmelt runoff pathways in a boreal forest hill slope, the role of pipe throughflow, *J. Hydrol., Vol. 95*, pp.39-54.

Robinson, J.S., and Sivapalan, M. (1997) Temporal scales and hydrological regimes: implications for flood frequency scaling, *Water Resour. Res., Vol.33*, pp.2981-2999.

Rodhe, A. (1981) Spring flood – meltwater or groundwater? *Nordic Hydrology, Vol.12*, pp.21-30.

Sand, K. (1990) Modeling snowmelt runoff processes in temperate and arctic environments, Norway Technical University, Dept. Hydraulic and Environmental Engineering, IVB rep. B-2-1990-1, Trondheim.

Semádeni-Davies, A.F. (1999) Snow heterogeneity in Luleå, Sweden, *Urban Water, Vol.1*, pp.39-47.

Semádeni-Davies, A.F., and Bengtsson, L. (1998) Snowmelt sensitivity to radiation in the urban environment, *Hydrol. Sci. J., Vol.43*, pp.67-89.

Shook, K., and Gray, D.M. (1966) Small-scale spatial structure of shallow snow covers, *Hydrol. Proc., Vol.10*,. pp.1283-1292.

Singh, V.P., Bengtsson, L., and Westerström, G. (1997) Kinematic wave modeling of saturated basal flow in a snow pack, *Hydrol. Processes, Vol.11*, pp.177-187.

The Swedish Committee on Spillway Design (1990) Principles for estimation of design flows for reservoir system in Sweden, Swedish Meteorological and Hydrological Institute, SMHI, Norrköping, Sweden 113 pp. (in Swedish).

Thorne, G.A., Laporte, J., and Clerke, D. (1998) The effects of frozen soils on groundwater recharge and discharge in granite rock terrane of the Canadian Shield, *Nordic Hydrology, Vol. 29*, pp.371-384.

Woo, M-k. (1998) Arctic snow cover information for hydrological investigations at various scales, *Nordic Hydrology, Vol. 29*, pp.245-266.

World Meteorological Organization (1986) *Intercomparison of models of snowmelt-runoff.* Operational Report No. 23, World Meteorological Organization, Geneva.

Xu, F., and Buttle, J.M. (1987) Patterns of net radiation over urban snowpacks, Proc. East. Snow Conf. 43, pp.173-184.

Received: December, 1999
Revised: April, 2000
Accepted: September, 2000

Addresses:
Lars Bengtsson,
Water Resources Engineering,
Lund University,
P.O.Box 118 LTH,
SE-22100 Lund,
Sweden.
Email: Lars.Bengtsson@tvrl.lth.se

Vijay P. Singh,
Civil & Environ.Engineering,
Louisianna State University,
Baton Rouge, LA 70803,
U.S.A.