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Urban snowmelt processes – current research and modelling needs

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Abstract

Despite the water balances of cold region towns being dominated by low intensity, long duration snowmelt events, urban drainage systems continue to be designed according to standards developed for short, high intensity rain storms. During the 1980s and early 1990s, work in Scandinavia (Bengtsson, 1983, 1984, 1986; Westerström, 1984; Bengtsson and Westerström, 1992; Thorolfsson, 1990) and Canada (Xu and Buttle, 1987; Buttle and Xu, 1988, Todhunter *et al.*, 1992) identified fundamental differences between rural and urban snowmelt processes. They found that snow properties such as density and albedo varied both between town and country and within the town depending on land-use. Moreover, both the longwave and shortwave radiation balances are heavily modified by buildings. Thus melt and runoff generation occurs at different times and rates. Town centres can have melt rates almost double that of residential areas. Despite snow removal policies, snowmelt in town centres is extremely important as these areas are the most likely to have combined sewer systems. These revelations will come as no surprise to practitioners working in cold regions, however, there is a lack of published material. This paper documents urban snow research from the last decade, it is both a summary and continuation of the state-of-the-art review found in a UNESCO special report on urban drainage in cold regions (Chapter 2, Semadeni-Davies and Bengtsson, 2000). Topics discussed include snow distribution, snow energy balance, frozen soil and runoff generation and modelling approaches - water quality issues are outside the scope. How to improve temporal and spatial resolution with limited budgets and limited data availability are ongoing problems, however, recent coupling between major urban drainage models such as SWMM and MOUSE and Geographic Information Systems offers a glimmer of hope. While full physically-based snow melt modelling is still out of the question, GIS could allow improved representations of snow distribution and energetics.

Keywords: hydrology, distribution, SWE, SCA, energy balance, runoff generation, scale, modelling

1 Introduction

Snowmelt is a poor cousin to rainfall when it comes to urban stormwater research. However, the last decade has seen increased interest in winter urban drainage largely due to the recognition that drainage systems, including BMPs (best management practices), in cold region towns are often unable to function during snowmelt (Marsalek, 1991). This point was recently reiterated by various authors in a UNESCO sponsored report on urban drainage in cold climates (UNESCO, 2000) and was taken up as a theme of a conference earlier this year in Sweden (1st International Conference on Urban Drainage and Highway Runoff in Cold Climate, 25-27 March 2003, Riksgränsen). Frozen soil, ice blockages and full storage facilities make inundation and combined sewer overflow (CSO) a recurrent spring event in some towns, particularly after rain-on-snow when the entire catchment can be contributing to runoff. The fact that urban snow, particularly near busy roads, can have water quality poorer than rainwater (see Viklander, 1997) makes matters worse. The problems incurred reflect a lack of knowledge of urban snow dynamics and winter runoff pathways. Symptomatic of this are simplistic snow and soil routines within drainage models used for design and operation For instance, the degree-day melt algorithm, which relates melt conceptually to average daily air temperature, does not have a spatial or temporal scale fine enough for simulating snowmelt driven CSO.

The dynamics of stormwater generation during winter and spring are very different from summer and autumn largely due to the fundamental differences between rain and snowmelt. While most urban drainage systems in cold regions have been designed for summer storm bursts, snowmelt can be the most hydrologically important event of the year (Bengtsson and Semadeni-Davies, 2000). Intense rain storms are short-lived whereas snowmelt persists slowly from days to weeks without drama. Yet the longevity of low intensity snowmelt means that drainage systems are often at full capacity during thaw. Moreover, pipes may be ice blocked and valves and gates frozen. Add to this the fact that winter flow pathways may differ from summer leading to increased water volumes and you have the potential for drainage systems that fail to function. Winter and spring flow pathways are largely related to land-use. High density land-use is associated with loss of vegetation and compacted soils and high percentages of impervious surfaces such as roads and roofs, all of which favour overland flow. Land-use also affects snow properties; the more intense the activity, the more compacted and polluted the snow and the more likely that that snow has been subject to some form of handling (Semadeni-Davies and Bengtsson, 2000). Impervious surfaces which ensure rapid response to even minor rainfalls tend to be snow-free prior to melt; snow is instead ploughed onto vacant lots or roadside verges. High spring flows thus lend credence to the idea that melt water may flow over normally permeable soil rather than infiltrate (Bengtsson and Westerström, 1992; Buttle and Xu, 1988). Rain-on-snow represents the worst of both worlds as it can spark rapid snowmelt leading to extreme flow events out of proportion to rain volumes with the entire catchment contributing to runoff. A combination of snowmelt, frozen soil and rain-on-snow can lead to high flood risks in winter and early spring. Floods during March 1997 and February 1999 in the Norwegian city of Trondheim are good examples (see, Milina, 2000). While the first had a rainfall return period of 15 years, runoff represented the 50-year flood. The snowpack acted as a dam storing initial rain water which was rapidly released upon melt. Frozen soil and ice in drainage dikes ensured rapid flow to the stormwater network.

Snow hydrology is common theme running through the topics of discussion at this meeting – after all snowmelt is the driving force behind winter and spring urban runoff. This paper can be seen as a follow up to the UNESCO Report chapter on snow hydrology (Semadeni-Davies and Bengtsson, 2000) which, due to a lack of urban research, was adapted to the urban context from mainstream snow hydrology . The origins of urban snow research can be traced to two main groups, one in Sweden (Bengtsson, 1983, 1984, 1986; Westerström, 1981, 1984; Bengtsson and Westerström, 1992) and the other in Canada (Xu and Buttle, 1987; Buttle and Xu, 1988, Todhunter *et al.*, 1992) working during the 1980s and early 1990s. Together they identified fundamental differences between rural and urban snowmelt processes and runoff generation. The contemporary establishment of the Risvollan urban research catchment in 1986, Trondheim, Norway (Thorolfsson, 1990) has provided a valuable long-term data set which is at the centre of current snow research. Today, the main players in urban snow hydrology in Scandinavia are located at Lund University in Sweden and the Norwegian University of Science and Technology in Norway. This paper outlines the work at these institutes with respect to snow distribution, energy balance and modelling needs. Particular attention is paid to improving the performance of snowmelt runoff models by increasing resolution. The key seems to be observing and describing small scale heterogeneity in snowmelt and runoff processes. Snow handling and water quality, which are the main thrust of research at the Luleå University of Technology, Sweden (see Viklander 1997), are outside the scope of this paper.

2 Observations of snow and snowmelt

2.1 Snow Distribution

The water balance by of a snow pack can be written as:

SWE = (P + C) – (E + S + R) Equation 1

Where SWE is the snow water equivalent (the depth of liquid water held in the snowpack if the pack were melted), P is precipitation, C, E and S are condensation, evaporation and sublimation respectively and R is runoff from the base of the snowpack including rain which has percolated through the snowpack. SWE is the most important input of any snowmelt runoff model as this variable determines the volume of water available for runoff generation (WMO, 1986; WMO, 1992). The use of rain gauges to measure snowfall is notorious due to under-catch (*ibid*), even so, SWE prior to melt is often internally calculated, albeit with some correction, as a function of precipitation and temperature in operational snow models. The urban drainage models MOUSE (DHI, 1994) and US EPA SWMM (see Huber and Dickinson, 1988) take this approach. "Snowfall" is then accumulated in a snow storage magazine which keeps a running tab of SWE. Alternatively, SWE can be calculated as the product of snow depth and density – this method requires accurate snow surveying to provide an estimate of catchment wide or areal snow volume. The total volume of snow in a catchment and is usually defined in terms of point SWE and the snow covered area (SCA) which is a substitute for snow distribution; that is, melt water runoff is only simulated for snow-covered areas and only while there is snow in the SWE magazine.

At its simplest, a snow model can assume an even blanket of snow with no spatial variation of SWE, that is, SCA is said to be 100 % and total snow volume is the product of average point SWE and catchment area. In urban hydrology, where melt processes and flow pathways are dictated by land-use, and where the interest is in small scale processes, more complex representations of snow distribution are needed. The town can be seen as a mosaic where land-use not only influences snow distribution but also melt rates and flow paths. Snow location with respect to buildings and roads largely determines the energy available for melt and the flow pathways of the melt water. Thus compacted and dirty snow ploughed on an asphalted verge of a main road will have very different melt behaviour and runoff generation from undisturbed clean snow in a suburban backyard. A common proxy method for heterogeneity is to use a snow depletion curve where SCA is linked to some snow characteristic such as average SWE or days since melt began. SWMM takes this approach based on snow depth. Snow is redistributed into four "sub-catchments" relating to different surface types (e.g., roads, bare soil). The shape of depletion curves for each surface type where snow is present is user defined. Depletion curves are recommended for rural snowmelt models (e.g., Rango and Martinec, 1982), but their application to urban areas is not well documented. One foreseeable problem is that SCA is a weighting that does not link snow to a certain location and therefore cannot give any spatial information on melt processes or flow paths.

Perhaps direct measurement offers a better alternative. Westerström (1984) was able to model daily stormwater flows from a residential catchment according to SCA which was manually updated from a daily snow survey. While accurate, the method was time consuming and of little practical value outside research. Determining point SWE and SCA for urban applications is problematic, especially where land-use is intense. Traditionally, snow distribution is mapped on the basis of straight snow courses of between 100 and 500 m in length where point measurements of depth and density are made at regular intervals – say every 10-20 metres. Snow courses are not suited to most urban environments as they depend on some degree of autocorrelation between adjacent sites. Ploughing and urban barriers such as roads and property boundaries make obtaining unbiased snow surveys extremely difficult if not impossible. Unlike rural catchments, local topography is second to snow handling when it comes to snow distribution. Buttle and Xu (1988) and Bengtsson and Westerström (1992) presented snow surveys in residential and inner city areas and found that many impervious surfaces were snow-free at the time of snowmelt. Ploughing and application of de-icing salt removes snow from roads while roofs are exposed to sun and wind causing rapid evaporation and melt. Snow is also blown free and is able to slide off sloped roofs. Semadeni-Davies (1999 b) largely confirmed Bengtsson and Westerström's (1992) findings for Luleå, Sweden, and extended their survey to include a industrial zone and a high density housing estate. Measurements of SWE were made at representative points (e.g., roadside snow pile and undisturbed snow) which were used to build a set of rules for snow location and characteristics (depth, density, albedo). For instance, piles of ploughed snow are likely to be on impervious surfaces in the inner city but on soil in residential areas. Aerial photographs were used to map snow cover on the basis of ground photographs of snow location. A summary of the survey is shown in Table 1 for three land uses. While snow piles had the greatest point SWE in the residential area, undisturbed snow made the accounted for up to 70% of the areal SWE. The opposite was true for inner-city areas.

Landuse	Snow cover (%)		Density (kg m^{-3})		Depth (m)		Albedo	
	Piles	Other	Piles	Other	Piles	Other	Piles	Other
Houses	10	55	250 - 550	240-310	1-4	0.6 - 1.5	0.3 - 0.5	0.4-0.6
Apartments	7	0	400 - 700	-	1-3	-	<0.3	-
Shopping street	5	0	>500	-	0.5-2	-	<0.3	-

Table 1 Summary of snow cover and properties for three land use types, Luleå, 1998

Matheussen and Thorolfsson (2001) presented snow survey data for the Risvollan urban catchment in Trondheim showing the same general pattern as in Canada and Sweden. Measurements were made each winter from 1994 to 2000 over three snow courses, two in parkland and a third which includes residential land, a kindergarten playground and a road. Noting the inter-annual similarity in snow distribution, Matheussen and Thorolfsson (2002 a) later used a web-cam set on an apartment block rooftop to monitor snow accumulation and depletion around an area of single family houses. The camera has now been in operation for three snow seasons. The rationale was to determine SCA to simulate catchment wide runoff with some link to physical location. The method is similar to snow mapping from remotely sensed satellite images of snow distributions in the wider environment, and offers a means of mapping snow accumulation and depletion over time. Similar photo-based methods have been set up in remote rural catchments such as on Greenland (Hinkler *et al.*, 2003). With an eye to providing a tool for calibrating and validating SCA and snow depletion curves within urban snow models, Matheussen and Thorolfsson (2003 a) trained a neural network to recognise snow cover within the digital images (Fig. 1). The network was trained using a series of pictures taken at 15 minute intervals during spring 2001 and the winter and spring of 2001-2002; a total of 165 images. The neural network was validated with good agreement against 72 aerial photographs taken on four separate occasions during the spring of 2002. The method is particularly powerful as it can be inputted into a GIS platform. Indeed, the areas analysed (black in Fig. 1) are categorised according to land-use which forms an attribute of a grid snowmelt model (see below). However, the web-cam image is oblique and therefore sensitive to obstructions. Moreover, snow in shadows and low albedo snow can be falsely registered as bare ground.

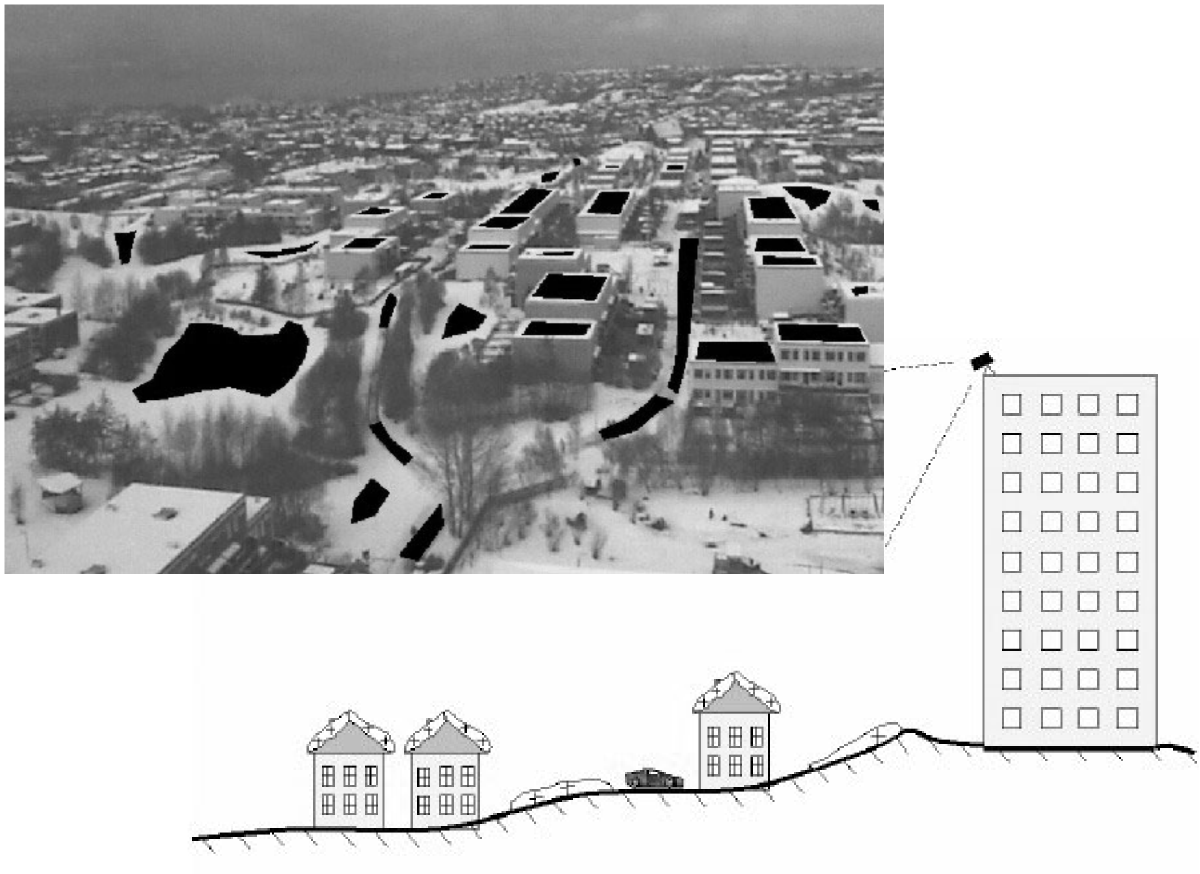

Figure 1 Set-up of digital camera and sample photograph showing ground truth areas for the Risvollan catchment. (modified from Matheussen and Thorolfsson, 2003 a)

2.2 Snow Energy Balance

There are three main phases of snowmelt in urban areas:

- melt from roads due to snow handling including application of de-icing salts and direct heating
- early melt where snow is in transition from cold to warm and liquid water can be re-frozen or stored in the snowpack
- late melt, possibly in combination with rainfall, where melt at the surface is quickly released as runoff from the base of the snowpack.

The timing of each phase is very dependant on snow location. This discussion does not consider melt due to snow handling and concentrates on urban micro–climates. Figure 2 shows the inwards and outwards energy fluxes for a generalised snowpack; alternatively, the energy balance can be written as:

$$Q_M = Q^* + Q_H + Q_E + Q_P + Q_G - \frac{dE}{dt} \qquad \text{Equation 2}$$

where Q_M is heat flux density available for melt; Q^* is net allwave radiation flux density (net longwave plus net shortwave or solar radiation), Q_H is turbulent sensible heat flux density; Q_E is turbulent latent heat flux density, Q_P is heat advected flux from rain; Q_G is the conductive heat flux density to or from the ground – often considered negligible - and dE/dt is the change in the internal energy held in the snow volume. All fluxes are measured in Wm^{-2}. Energy reaching cold (below freezing) snow is initially used to overcome the snow cold content so that dE/dt=0, that is, the snow is isothermal at 0˚C.

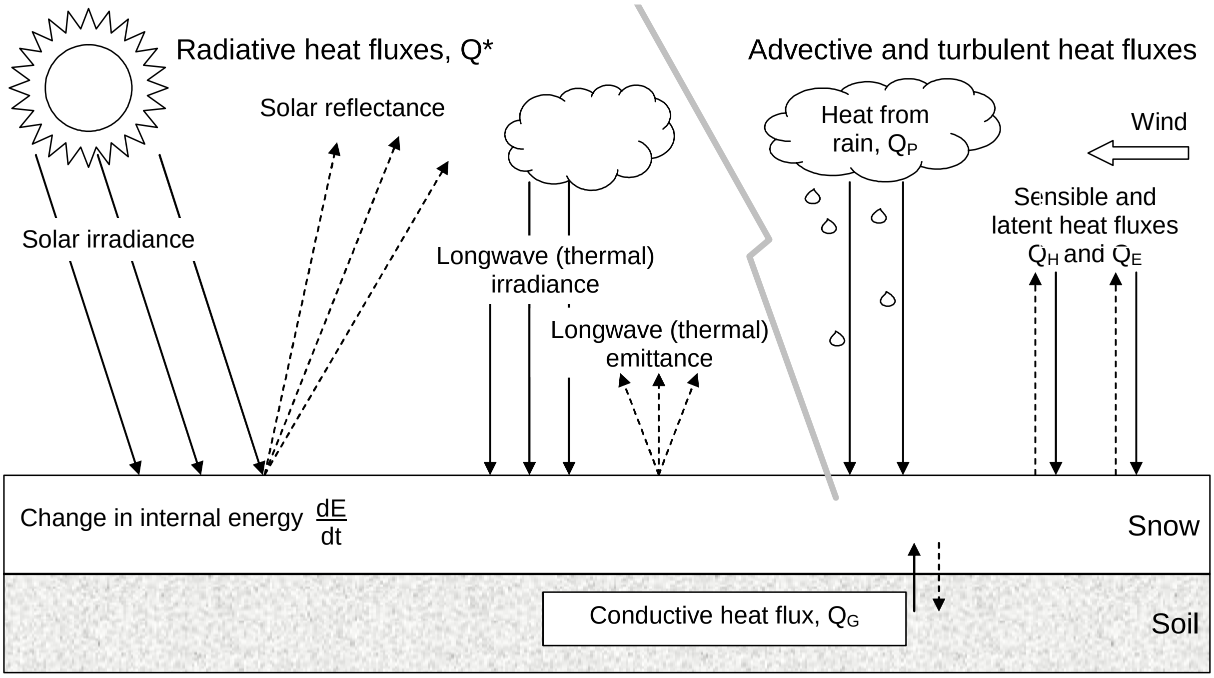

Figure 2 Heat fluxes for a snow volume, solid lines represent heat gains and dashed losses.

Once isothermal, energy available is able for melt calculated as:

$$M_s = \frac{Q_M}{\rho_w L_f} \qquad \text{Equation 3}$$

where M_s is the rate of melt (mm s^{-1}), ρ_w is the density of liquid water at 0ºC (kg m^{-3}) and L_f is the latent heat of fusion (J kg^{-1}).

Liquid water entering subfreezing snow will freeze releasing latent heat which warms the snowpack. If the snowpack is isothermal, liquid water will be held until the irreducible water content of the snow pack is reached. The amount of water that can be held by the snowpack is approximately 3-4% by volume and depends on snow texture, grain size and grain shape. Liquid water in the snowpack is subject to freezing if the temperature drops, cold nights cause water to freeze downwards from the surface and once again the cold content and irreducible water content must be overcome for melt water to be released. Assuming no preferential flow, for an initially cold snowpack the time delay between first melt at the snow surface and runoff generation can be several days. Anderson (1976) provided a comprehensive description of snowmelt which has been incorporated into complex snow models such as CROCUS (Brun *et al*, 1989; 1992) and SNTHERM (Jordan, 1991).

Snow research is rare in urban climatology; a recent comprehensive review of urban climate research mentioned snow only with respect to its impact on surface reflectance (Arnfield, 2003). The effect of high albedo of new fallen snow on net solar radiation is arguably the most dramatic example of snow modifying urban climates (Oke, 1987). Albedo (reflectivity or the ratio of outgoing to incoming shortwave radiation) is dependant on grain structure, density, the presence of impurities, sun elevation, cloud cover, snow depth and the albedo of the underlying surface. The dependence on density means that albedo is often related to snow age (USACE, 1956), however, pollutants and ploughing – which compacts and mixes the snow – lower albedo dramatically in urban areas. Near roads, urban snow is known for its dirty black appearance due to pollutants. Indeed, Novotny *et al* (1999) found that 90% of pollution from traffic was found in a 5m swath of snow adjacent to roads. This snow can melt quickly due to the low albedo although piling snow beside roads complicates matters by changing depth and density. Fresh snow acts like a mirror, albedo normally ranges from 0.8 for fresh dry snow to 0.6 at the onset of melt and 0.4 for wet melting snow (USACE, 1956). By comparison, the snow albedo measured in downtown Luleå by Bengtsson and Westerström (1992) and Semadeni-Davies (1998) almost 20 years later was around 0.2, which is equivalent to bare soil.

Snow also effects urban net longwave radiation, at below freezing, clean snow is almost a perfect blackbody (emissivity = 1) which emits more radiation than surfaces with similar temperatures (e.g., concrete, emissivity = 0.7-0.9). The atmosphere has an emissivity of around 0.8 under clear skies and approaches unity with increasing cloud cover. Outgoing radiation is proportional to the fourth power of snow surface temperature, emissivity is a scalar. At below freezing air temperatures, the snow surface is able to emit more longwave radiation that it receives as the snow emissivity is greater than the emissivity of air. However, for air temperatures above freezing, the incoming longwave radiation can be greater than the outgoing longwave radiation which has an upper limit (about 312 W m^{-2}) corresponding to a maximum temperature of 0°C with ε=1. For old dirty snow ($\varepsilon \leq 0.82$, see Oke 1987), the upwards emissions would be even less ($\leq$ 258 W m^{-2}) at a time when absorption of shortwave radiation is high. This means that urban snow has a potential for much greater net allwave radiation absorption than rural snow of the same age even before one considers the affect of urban topography..

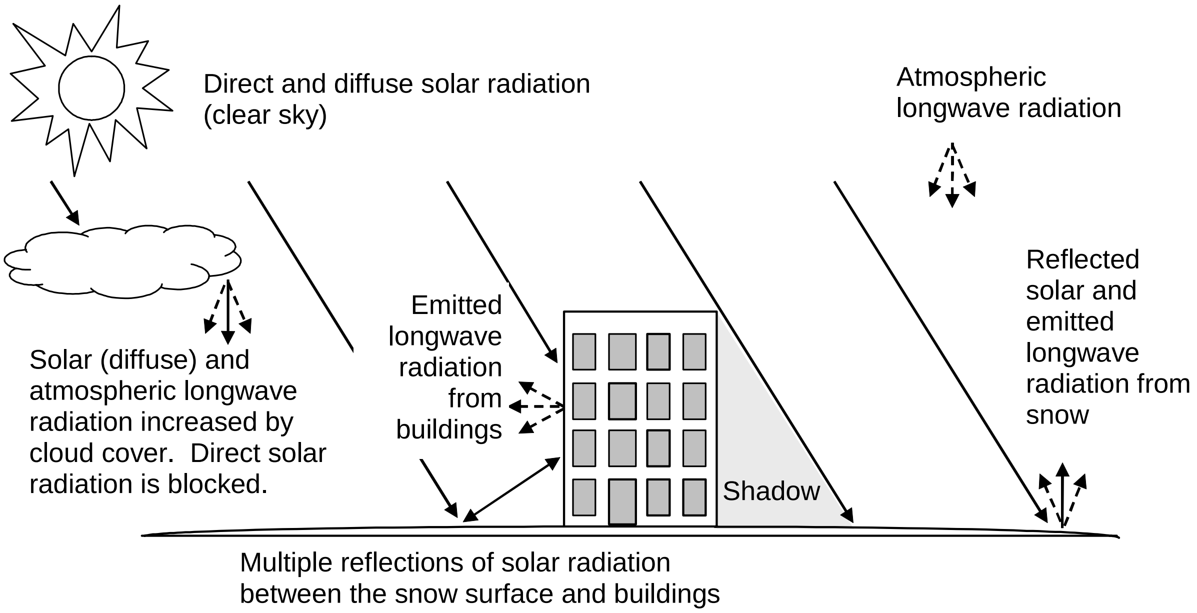

Figure 3 The effect of a building on radiation exchanges between the snowpack and atmosphere. Solid and dashed lines refer to solar and longwave radiation respectively. (From Semadeni-Davies *et al.,* 2001)

Aside from snow properties, energy availability is tied to snow location with respect to urban structures. Urban topography, with its canyon configuration of road valleys and buildings peaks, most resembles alpine conditions in that there is a vertical plane (Fig. 3). In both situations the turbulent and radiative fluxes are heavily modified by surrounding terrain / buildings. Obled and Harder (1979) found that slope angle and aspect to the solar beam are of prime importance for determining irradiance in alpine areas as these dictate the amount of shortwave radiation possible at a point. This "raw" irradiance is then altered by shading, multiple reflections and longwave emissions from neighbouring topographical features. Xu and Buttle (1987) reported that net radiation over clean snow in a Canadian suburb could be enhanced by between 67 and 435 %. This snow had albedo and emissivity similar to surrounding rural snow, so enhancement was due to interaction with local houses. Measurements from Sweden show a similar pattern with increases of 10 to 100 W m^{-2} up to 10 m from buildings (Bengtsson and Westerström, 1992).

Todhunter *et al. (*1992) presented a net radiation model over an urban snowpack which was validated using data collected by Xu and Buttle (1987). Semadeni-Davies and Bengtsson (1998) coupled a similar radiation model to a physically based snowmelt model to demonstrate how melt could be affected by buildings. The radiation model assumptions were simplified from Todhunter *et al.* (1992), for instance the building was said to be infinitely long and lying from west to east with an north (shaded) and a south (exposed) face. Radiation and runoff simulations were tested against data collected from an open site with very good fit (R^2 was 0.90 and 0.71 for net allwave radiation and runoff respectively). The coupled model was used to drive a sensitivity analysis to investigate the effects of cloud cover; snow albedo; building outer wall temperature and the distance between snow and the building. It was found that buildings have the greatest influence over the radiation budget on sunny days when there is obvious shading to the north and heating of walls to the south. Under cloudy skies, shortwave radiation is restricted to diffuse radiation and incoming radiation is similar both to the north and south of buildings. Incoming longwave radiation and diffuse solar radiation near walls are limited by sky-view, this is the proportion of the sky dome that is not obstructed by buildings. The corollary is wall-view, indeed there may still be longwave enhancement on cloudy days due to the lower emissivity of building materials compared to snow. The effect of buildings on the local radiation balance can be likened to a forest canopy which screens the underlying snowpack and absorbs much incoming shortwave radiation which enhances incoming longwave radiation at the forest floor (e.g., Pomeroy and Dion, 1996).

The simulated changes in melt rates due to increased albedo were consistent with field observations of artificially blackened snow (Conway *et al.,* 1996). Shortwave radiation absorbed by dirty snow caused dramatic increases in melt – a drop in albedo from 0.8 to 0.4 , which is consistent with urban snow, means an increase in melt of at least 6 mm day^{-1} in full sunlight.

Semadeni-Davies *et al. (*2001 a) validated the radiation model used by Semadeni-Davies and Bengtsson (1998) with purpose collected data from a black plastic clad wall and showed how sensitive the snow radiation budget is to urban structures. An example of their findings is given in Fig 4. The model and measurements show differences in net radiation over snow near walls compared to open sites that range from –120 to +150 W m^{-2}. To give an indication of the effect on snowmelt, consider that a loss or gain of 100 W m^{-2} represents around 1 mm h^{-1}, while this is not as dramatic as a summer storm, given the persistence of melt over time and possibility of soil frost, the volume of water soon mounts up. Semadeni-Davies *et al. (*2001 a) noted that the complexity of a full radiation model precludes its inclusion within an urban drainage model.

Aside from redistribution, snow ploughing effects melt by changing snow characteristics and pack geometry. In ploughed snow the natural layering is broken and snow becomes compacted. The bulk energy balance of a melting snow pile was discussed and modelled by Sundin *et al. (*1999) and Sundin (1998). Pile shape greatly influences melt due to high surface exposure to radiation and wind per unit area; furthermore, steep sides change the local sun angle from that of surrounding horizontal snowpack. This means that in the mornings and afternoon part of the snow pile is in full sunlight, while the opposite side will be in shadow.

However, the water concentration of piled snow in comparison to unploughed snow means that piles have considerably less exposure per unit volume water than undisturbed snowpacks. The effect of gravel particles was not discussed except to say that the surface albedo was generally low. Translated into degree day coefficients (see discussion below), Sundin (1998) estimates a melt rate as high as 11 mm/˚C/day compared with the 8 mm/˚C/day maximum quoted by Bengtsson and Westerström (1992) for inner city snow. Even so, the high SWE for a snow-pile and later insulation by a crusting of friction material where present can mean that snow piles persist well after the snowpack proper is melted.

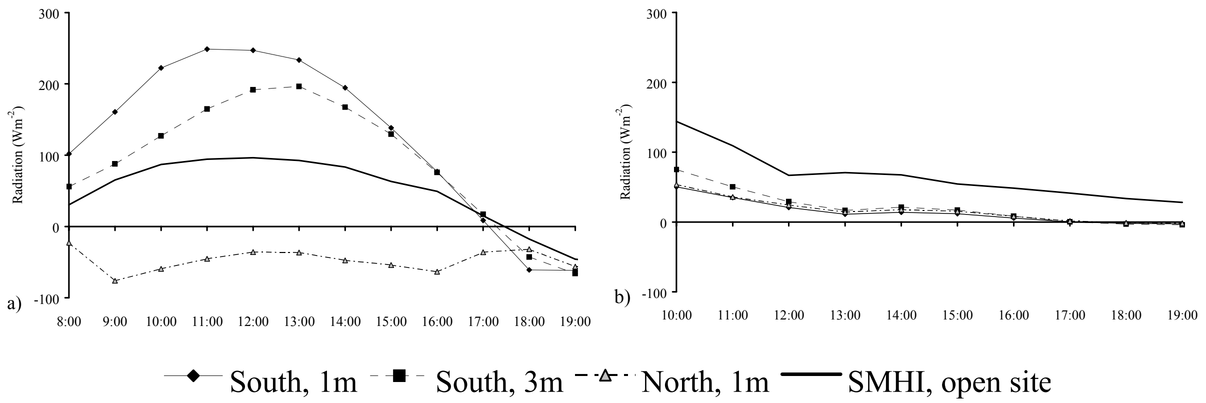

Figure 4 Net allwave radiation measured over a snow pack on two sides of a purpose built wall for a) clear sky (16 April, 1998) and b) cloudy sky (18 April, 1998) compared to an open site at the Swedish Meteorological and Hydrological Institute metrological station 5 km distant. Background example from Semadeni-Davies *et al.* (2001 a) published in Semadeni-Davies *et al.* (2001 b)

A remaining challenge for urban snow hydrologists is to determine the impact of urban infrastructure on the turbulent fluxes of latent and sensible heat. Wind tunnelling or sheltering are obvious candidates for investigation. The effect of wind is easily seen at street level when it comes to snow drifts, but how does wind flow influence melt? Certainly notions of a logarithmic wind profile are not applicable to the urban environment. Investigations over natural snow covers shows that uneven snow accumulation favours the development snow-free patches (e.g., Shook and Gray, 1996) which has particular importance for small scale processes by altering the near-surface wind, air temperature and humidity. Liston (1995) developed a numerical atmospheric boundary layer model to simulate local exchanges of momentum, heat and moisture for patchy snow. With a snow-free, bare-soil fetch of 4 km followed by a 4 km stretch of snow, Q_E is estimated to around -30 W m^{-2} at the snow leading-edge whereas 1 km distant it is slightly positive. The total energy available to the snow increases by up to 30% as snow-free and snow-covered patches decrease in size to 100 m. Observations by Neumann and Marsh (1998) largely confirmed the local advection modelled by Liston (1995). In towns where roads are cleared and roofs are exposed, local advection is likely to be even more pronounced. Semadeni-Davies (unpublished background data to Semadeni-Davies *et al* 2001 a) found that the difference in temperature between roadside snow and asphalt could be as great as 30˚C.

2.3 Runoff

Once melt water has reached the base of the snowpack it is free to either flow overland or infiltrate soil. The soil (thermal conductivity, heat capacity) and hydraulic (hydraulic conductivity, water storage capacity) properties dictate whether infiltration can take place. Ice grains in soil serve to decrease porosity and thus hydraulic conductivity, and horizontal ice lenses limit vertical flow. The hydrology of frozen soils has been investigated by a number of researchers over the last 30 years (see Dunne and Black, 1971; Kane and Stein, 1983; Kane & Chacho, 1990; Williams and Smith, 1991). Generally, infiltration capacity depends largely on soil moisture before freeze-up. Initially dry frozen soil has open pore spaces that allow infiltration. The soil will warm to isothermal at 0ºC as latent heat is released by water freezing on contact with soil. If the soil is very cold, ice lenses can form at soil interfaces as water freezes before it can infiltrate. In initially saturated frozen soil, hydraulic conductivity is controlled by the grain-size of the soil and the pore ice content; pores become blocked with ice-grains leaving only a thin film of water around the soil particle. The colder the soil the less liquid water storage possible and the lower hydraulic conductivity - which can drop as much as 10^{-6} m s^{-1} between 0 and –1°C – due to ice build-up in pore spaces. Even so, in rural soils ground frost that allows unhindered infiltration at the catchment scale is most usual (e.g. Buttle & Sami, 1992). While there may be small scale overland flow (Dunne and Black, 1971), liquid water is generally able to infiltrate via adjacent unfrozen soil patches or macropores (Shanley and Chalmers, 1999) and spring stream flow peaks contain pre-event rather than snowmelt water. In urban catchments, soils are compacted by heavy machinery, topsoil is removed, horizons mixed and vegetation changed. Moreover, surface water flows short distances to stormwater inlets (reducing opportunities for infiltration) and the response time to rain or melt events is rapid. With this in mind, Bengtsson and Westerström (1992) suggest that the contributing area increases progressively over spring as soils become saturated or frozen or both.

Buttle and Xu (1988) found that the spring infiltration capacity of soils in a residential suburb ranged from 0-200 mm/h compared to up to 900 mm/h for nearby wooded areas. They also reported spring runoff coefficients for the catchment that had been partially urbanised in the 1970s. Upon urbanisation coefficients rose from around 0.06 to 0.23. Hydrographs for the rural and residential sub-catchments during 1984 and 1985 show that total spring runoff response to melt and precipitation are up to twice those of the rural subcatchment (e.g., 0.65 vs. 0.94; 0.30 vs. 0.49; etc). Runoff coefficients for individual melt generated quick-flow events are usually greater in the residential area. Both land-uses experience increases in overland flow as soil becomes wetter, by the end of the melt season the rural values can exceed the residential. More intense melt rates mean that the urban snowpack is depleted causing the residential runoff response to snowmelt to decline while that of the rural subcatchment is increasing. Higher coefficients in residential areas have several causes such as impervious surfaces, higher intensity melt rates leading to quicker saturation, compaction, channelisation and, finally, increased exposure leading to greater frost penetration and lowered infiltration capacity.

A similar illustration of seasonal changes to flow pathways comes from the Twin Cities, Minnesota (Brown, 1987) where runoff coefficients for 20 catchments were monitored for one year. Catchments ranged both in size (0.34 to 215 km^2) and land-use (rural to fully urbanised). The coefficients were based on the total annual volumes of surface runoff derived from rainfall and snowmelt respectively. For all the catchments, the runoff coefficient increased with urbanisation and greatest for snowmelt; the gap was greatest in partly urbanised areas (e.g., 0.3 vs. 0.1 at 50 % urbanised) and the two sets started to converge as urbanisation (imperviousness) tended to 100 % (e.g., 0.38 vs. 0.36). Urbanisation causes a decrease in the relative importance of soil properties as surfaces become increasingly covered in impervious materials such as asphalt and concrete. There was no breakdown for antecedent conditions for individual events. Thus a late spring melt event may have a high runoff coefficient due to soil saturation or ice build-up. Also, whether different rates of melt in urban and rural areas affect the coefficient was not discussed, nor were the possible effects of snow handling. If losses due to snow removal were not accounted for, the snowmelt volume expected would be overestimated leading to lower calculated coefficients than actual values.

3 Modelling needs

The extreme heterogeneity of urban environments currently precludes physically-based modelling to solve equations 1 and 2 in all but research applications. However, Buttle *et al.* (1990) expressed concern that assumptions surrounding urban snowmelt models built into commercially available drainage modes are largely untested in urban areas. Poor temporal resolution and lack of snow physics were named as possible problems. Spatial and temporal scales of hydrological processes are linked (Fig. 5). In urban areas the dominant runoff process is overland flow which has scales of <1 km, minutes (e.g., Schilling, 1991), but the recognition that the two go hand in hand does not translate into urban snowmelt modelling as is. Semadeni-Davies (2000) discussed urban snowmelt models largely with respect to the energy balance and scale issues and found that modelling techniques are determined both by standard available data and tradition rather than need. Yet the lack of data and costs associated with modelling means that the *status quo* will be with us for some years to come. This section looks at simulating snow distribution, snowmelt and runoff generation with emphasis on the need to incorporate spatial snowmelt patterns and finer time steps. Much of the work is still to be done and - without apologies - the discussion raises more questions than it answers.

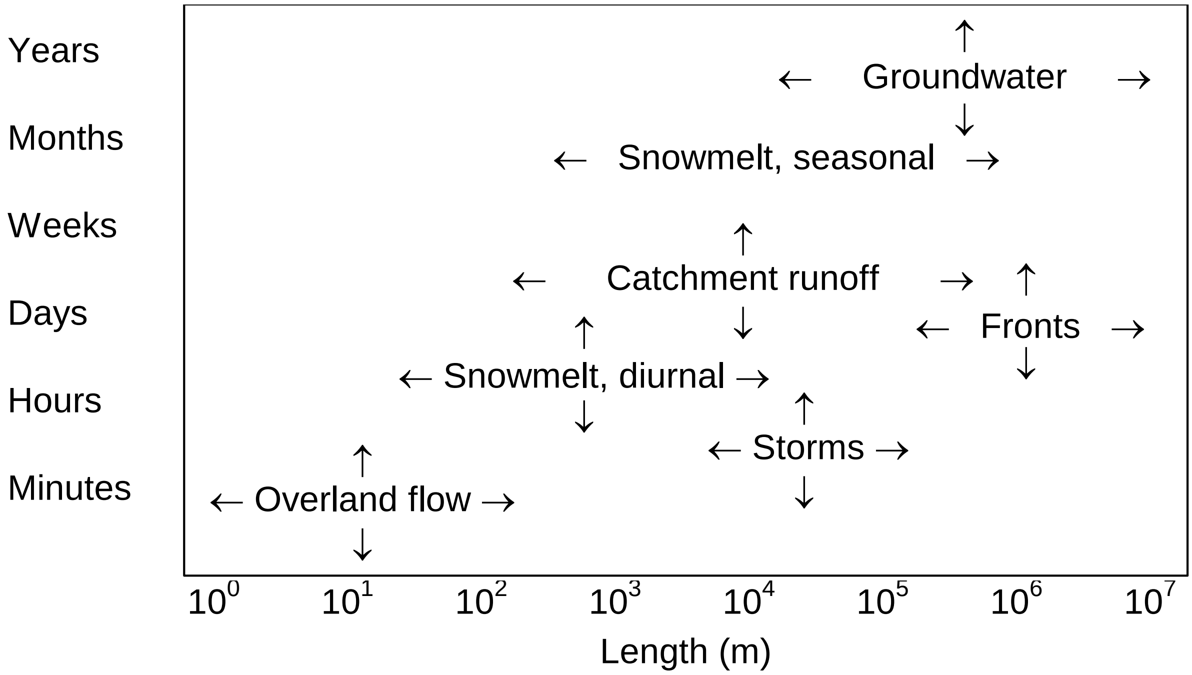

Figure 5 Link between spatial and temporal scales of hydrological and meteorological processes (after Semadeni-Davies and Bengtsson, 2000)

3.1 Spatial resolution

Snow distribution influences both melt processes and runoff generation. In the absence of snow removal out of the catchment, simulated snow accumulation can give an approximation of the total snow volume based on point SWE, however, assumptions on SCA and snow properties do affect runoff simulations (Semadeni-Davies, 1998). Of commercially available snowmelt models, SWMM has arguably the most realistic simulation of SCA which allows snow ploughing and redistribution according to surface cover and *a priori* user knowledge.

Mapping SCA with imaging tools (Semadeni-Davies, 1999 b; Matheussen and Thorolfsson, 2002 a) was discussed above. Matheussen and Thorolfsson (2002 b; 2003 b) extended their three land covers specified in the image analysis to five covers in order to develop a more physically-based urban snowmelt model. The break-down into different covers - road, shoulder, wall, roof and park - was motivated by both energy balance considerations and runoff generation with each having different routines for snow accumulation, distribution and ablation processes. For instance, roads only accumulate snow up to a SWE of 30mm, excess snow is "ploughed" and added to the shoulder land cover (a zone of 1-3 m adjacent to roads) in much the same way as SWMM. Their modelling work will be discussed further below in terms of GIS.

3.2 Temporal resolution: snow energetics

Snowmelt within commercial urban drainage models is simulated with some variant of the degree-day or temperature index method where melt is related to the average daily temperature:

$$\begin{aligned} M &= 0 && T_a \leq T_m \\ M &= C_m\left(T_a - T_m\right) && T_a > T_m \end{aligned} \qquad \text{Equation 4}$$

where, M is daily snowmelt (mm day^{-1}), C_m is the melt-rate factor (mm/°C/day), T_a is the average daily ambient air temperature (°C) and T_m is the threshold melt temperature (°C). T_m, is often intuitively set to 0 ºC. The melt-rate factor, C_m, varies with location and snow characteristics. For open areas C_m ranges from 2 to 10 mm/°C/day, for forests and exposed windy slopes respectively. In urban areas, the coefficient can range from 1.5 to 11 mm/°C/day with the value highly dependant on location and timing in the melt season (Westerström, 1984; Sundin, 1998, see Table 2). At the catchment scale, this method can provide reliable estimates of daily runoff (WMO, 1986) often outperforming physically-based models (e.g. Ferguson and Morris 1987). Westerström (1984) found the degree-day method in tandem with detailed observations of SCA adequate for daily stormwater generation, however Bengtsson (1986) states that the temporal and spatial scale is unsuited to most urban applications. While some applications, such as simulating sewer infiltration, can be carried out with daily time-steps (Semadeni-Davies and Bengtsson, 2000), Matheussen and Thorolfsson (1999) found that a temporal resolution of at least one hour is essential to model combined sewer overflows. This is important in regions such as coastal Norway where CSO is associated with snowmelt (e.g., Thorolfsson and Brandt, 1996). Semadeni-Davies (2000) overviewed the method with respect to urban drainage and found it lacking in theoretical validity. She also pointed out the discordant scale between melt calculations and pipe routines driven by surface runoff (e.g., kinematic wave and St-Vennant equations) found in commercial drainage models.

Landuse	Melt rate, C_m, (mm/ºC/day)
Suburban housing	1.5-7
Inner-city (park and apartment yards)	1.5 - 8
Ploughed snow in piles	5-11

Table 2 Degree-day melt rate coefficients measured in Luleå (derived from Westerström, 1984; Bengtsson and Westerström, 1992; Sundin, 1998)

Sand (1990) and Buttle *et al.* (1990) suggested incorporating energy balance components to improve the temporal resolution of runoff simulations. Sand (1990) compared an hourly net solar radiation index to lysimeter data, set in open sites, and found good agreement. Thorolfsson and Killingtveit (1991) incorporated this routine into the conceptual HBV model structure (Lindström *et al.*, 1997) which was then applied successfully to several Norwegian urban or partly urbanised catchments with varying land-uses (commercial, industrial and residential) and sizes (0.2 – 6.9 km^2). Similarly, Buttle *et al.* (1990) found a simple point energy balance gave good results for infiltration into residential yards. Aside from runoff modelling, Bartošová and Novotny (1999) showed the value of an hourly energy balance snowmelt model when simulating stormwater quality, this model also incorporated the effect of de-icing salt on melt. None of these models included the shading or enhancement by buildings, however, Buttle *et al.* (1990) noted the problem.

Semadeni-Davies *et al.* (2001 b) compared several melt indices and showed that hourly runoff was better simulated for an open site with an hourly temperature index (R^2=0.61) than with a net radiation index ((R^2=0.53; Figure 6). The advantage of including the radiation term is that the effect of buildings could be represented spatially without resorting to a full energy balance. A daily temperature index is included in Figure 6 for completeness and shows how much detail is lost at this timescale.

3.3 Ground frost: a new challenge

Commercial urban drainage models assume that runoff processes are the same year round with overland flow governed by water content according to algorithms such as Horton or Green-Amt. Observations of urban spring runoff cited above, however, suggest that ground frost could be responsible for high stormwater flows from normally permeable surfaces. Indeed, soil frost was implicated as one of the causes of the floods in Trondheim noted earlier.

There are a number of physically-based models which solve governing energy and mass balance equations for soil layers at the point or plot scale. However, in urban catchments, there is a need for simple ground-frost routines due to heterogeneity and data requirements. Unlike snowmelt, ground-frost build-up and thaw are slow processes, so a daily time-step is probably adequate. Buttle and Xu (1988) compared runoff coefficients to pre-snow air temperatures and antecedent rainfall. They found that the number of days with air temperature below freezing prior to snowfall could account for much of the variance in spring quickflow. This suggests that an index could be used define infiltration capacity. Alternatively, spring infiltration simulations could have a form such as the empirical model derived by Zhao and Gray (1998). They determined statistical relationships between infiltration calculated with a coupled soil energy and mass balance model and various parameters such as pre-melt soil moisture storage. For instance, they suggest that infiltration into a clay - sandy loam drops as much as 25 to 41% with a temperature decrease from –4 to –8°C.

A third idea is to calibrate an existing infiltration model separately for summer and winter. This approach was used by Sand and Kane (1986) who parameterised the HBV model soil variables seasonally for an Alaskan catchment. Similarly Semadeni-Davies *et al* (2001 b) reduced field capacity in a conceptual runoff model to mimic loss of pore space due to ice-grain build-up. The model was compared to measurements from a runoff plot over gravel (Bengtsson, unpublished 1980). Changing the storage better simulated the total runoff volume although the correlation coefficient was the same as assuming that overland flow is due to saturation alone.

3.4 Geographic Information Systems

The above discussion shows the need to better describe spatial variations in snow distribution, energetics and surface permeability in order to improve temporal resolutions. The coupling of urban drainage models to Geographic Information Systems (GIS) offers a way forward, especially as the commonly used SWMM and MOUSE urban drainage models both have GIS plug-ins available. GIS provides a means for manipulating inter-related attributes stored as a series of layers; spatial information is held either as pixels which refer to a grid (raster mode) or polygons which group continuous points with similar attributes such as soil type or vegetation (vector mode). Each layer is thus analogous to a map and new maps can be formed by combining layers or performing some transformation function. GIS has proved valuable to hydrological science (e.g., Singh and Fiorentino, 1996) in general as well as for alpine snowmelt modelling (e.g. Cazorzi and Dalla Fontana, 1996) and urban hydrology (e.g., Coroza *et al.*, 1997; Zech *et al.*, 1993).

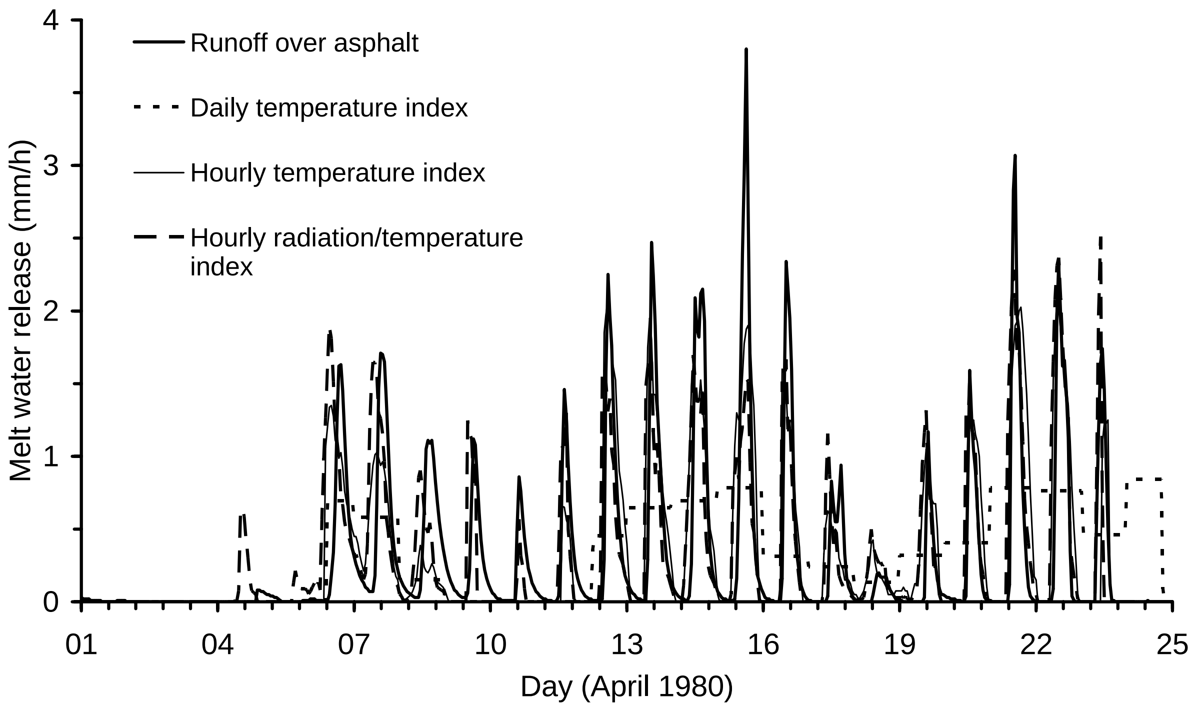

Figure 6 Measured runoff against snowmelt modelled with different melt indices (constructed from background data to Semadeni-Davies *et al.*, 2001 b)

The following example shows how a standard GIS package was used to model energy fields over a melting snowpack in a heterogeneous environment. Giesbrecht and Woo (2000) and Woo and Giesbrecht (2000) used overlays of forest attributes within Arc/INFO to determine the effect of trees on melt within a sub-Arctic woodland. A variant of their method, which emphasised the effect uneven radiation inputs, could prove useful for urban snow modelling. The aim was to show how the choice of spatial scale influences the simulated pattern of snow-free patch growth. Snow depth (and melt) was approximated hourly with respect to distance from tree trunks. Longwave radiation emitted by trees was simulated according to canopy and trunk temperatures derived from the ambient air temperature by regression analysis. Of particular interest given the extreme heterogeneity of urban radiation fields was the use of the inbuilt shadow function (HILLSHADE). Within the shaded zones, direct shortwave radiation was removed form energy balance calculations and air temperature was dropped 1ºC. The model gave good fit, but the discrete time-step did leave bands of unmelted snow projecting from trunks which related to the simulated position of the tree shadow. They did not consider sky-view for radiation simulations, however, the effect in this sparsely wooded area is probably so low as too not warrant the added effort of modelling the complex geometry. In an urban area, use of a buffer could enable longwave enhancement to be calculated according to building height and the snowpack distance from buildings, albeit with different parameters for full-sun and shade when the sky is clear.

The effect of snow cover heterogeneity and spatial variability of energy and snow properties on urban melt rates and runoff generation were demonstrated by Semadeni-Davies and Bengtsson (1998) and Semadeni-Davies (1998). Semadeni-Davies (1999 b) postulated that photo-derived maps of SCA could be of benefit to urban snowmelt modelling linked to locations, especially if snow depth, density and albedo are also mapped according to snow surveys or even some expert system. Matheussen and Thorolfsson (2002 b; 2003 b) used home-grown GIS techniques to run a simplified physically-based snowmelt / runoff model for the Risvollan catchment. The model was run continuously over three winter seasons. Snow was assumed to be ripe and the observations of SWE and SCA cited above were used for calibration of the snow model. Stream flow measurements were available to assess the runoff routines. The catchment was separated into 2x2 metre grid cells each of which was assigned a land-use as defined above. Roofs for instance have high exposure (i.e., rapid melt) and impervious surfaces so that all melt water is routed to stormwater drains. The wall land cover is a buffer zone around buildings that empirically simulates longwave radiation enhancement due to solar loading of walls. At this stage, sky-view is not included. Shading is handled by a digital elevation model (DEM) which also allows solar angles to be determined for each cell. Buildings are treated as hills within the DEM. Separate simulation of albedo is a new innovation that recognises the huge spatial variation of this snow property (see Bengtsson and Westerström, 1992; Semadeni-Davies, 1999 b). As in most albedo algorithms (e.g., USACE, 1956), the curve decays with age, however there are separate curves for each land cover with roads and shoulder having the greatest decay rates. While grid-based, the overall contribution of each land cover to runoff was combined to give an area weighted value. The model was able to approximate snow cover with limited success with snow depletion being too rapid. Runoff was adequately modelled (R^2 = 0.65) with an hourly time-step. Although the technique requires time for set-up and extensive data collection, it does illustrate how a physically-based model can be applied to urban areas.

4 Conclusion

Snow hydrology is at the heart of research into cold region urban drainage yet there are few researchers working in this field. This paper has presented an overview of research from the last decade with reference to pioneering work undertaken in the 1980s and early 1990s. Out of necessity, the review has concentrated on research at Lund University in Sweden and the Norwegian University of Science and Technology in Trondheim, Norway. If the 1980s were a time of observation, the 1990s were for consolidation and translation of this knowledge into models – work which is ongoing. The new challenges are to investigate turbulent fluxes and the role of frozen soil. There is a relationship between spatial and temporal resolution, both need to be improved to better simulate drainage phenomena such as CSO. However, modelling is restricted by extreme heterogeneity, high costs and lack of data. GIS is mentioned as a possible tool for resolving scale issues, but improved modelling requires an improved knowledge base and improved data collection for model inputs, calibration and testing.

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