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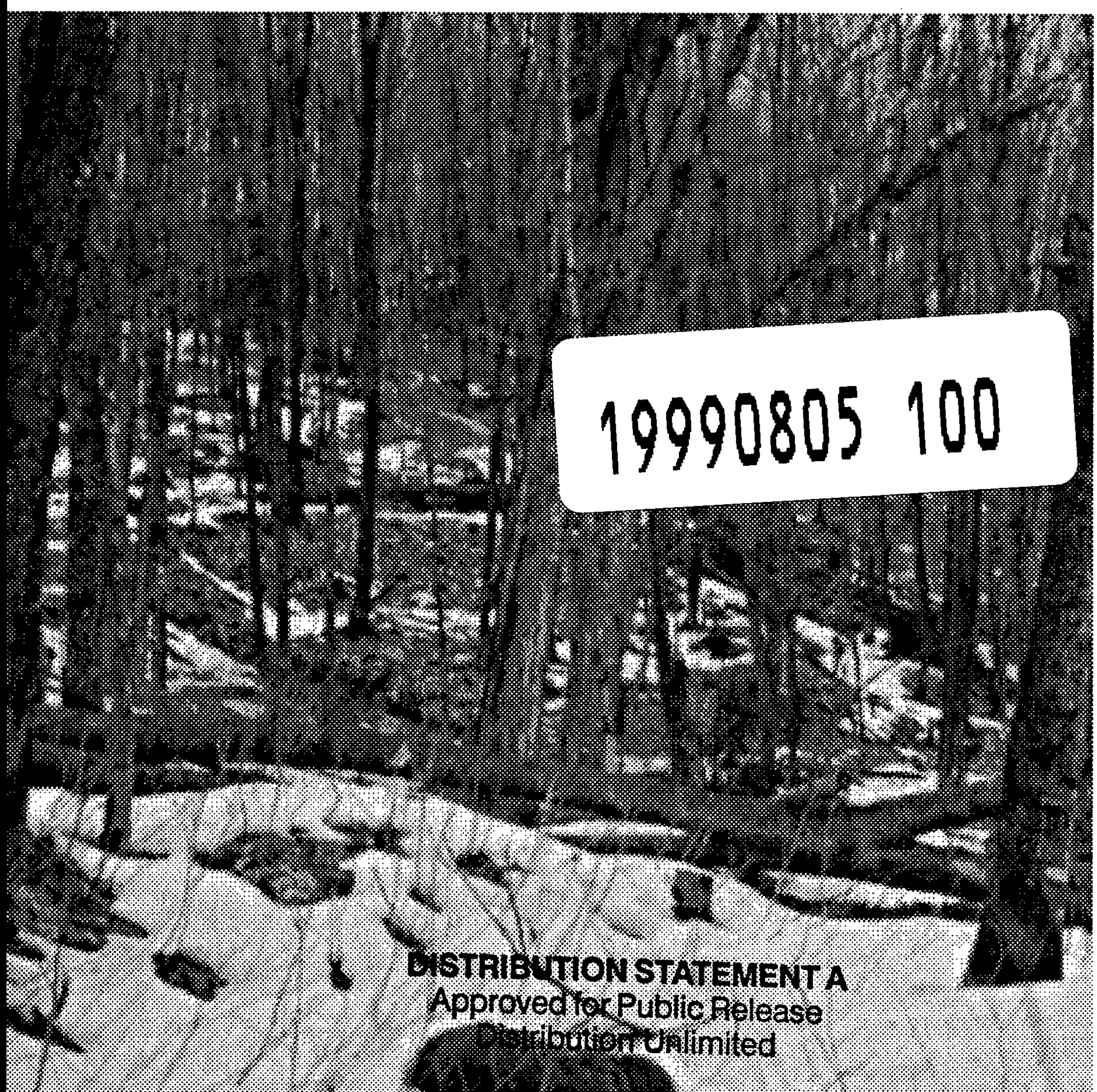
A Synopsis and Comparison of Selected Snowmelt Algorithms

Rae A. Melloh

July 1999



Abstract: One-dimensional snowpack algorithms in major operational snowmelt models used in the United States (HEC-1, SSARR, NWSRFS, SRM, and PRMS) are reviewed and contrasted with two U.S. Army Cold Regions Research and Engineering Laboratory (CRREL) algorithms (SNTHERM and SNAP) that are candidates for use in distributed operational models. In contrast to current operational models, the CRREL algorithms provide more detail in snowpack processes and require no calibration. The CRREL algorithms also include a full surface energy balance that requires more meteorological data than most operational models. Simpler surface energy balances could be used with the CRREL models. In future modeling systems, it would be preferable for the surface energy balance algorithms to be made independent of the internal snowpack process algorithms, so that available meteorological data can be used to drive a snowpack model of choice. Improvements are needed in the way that forest canopies and other groundcovers are accounted for in the surface energy balances of the CRREL models.

Cover: Snowmelt in the Sleepers River Research Watershed in Danville, Vermont. (Photo by R. Melloh.)

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Prepared for
OFFICE OF THE CHIEF OF ENGINEERS

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PREFACE

This report was prepared by Rae A. Melloh, Research Physical Scientist, Geological Sciences Division, U.S. Army Cold Regions Research and Engineering Laboratory (CRREL), Hanover, New Hampshire. Funding for this work was provided by the U.S. Army Corps of Engineers Hydrologic Engineer Center (HEC-96-030), and Civil Works, Cold Regions Engineering Program, Work Unit 31578, *Snowmelt Processes and Simulation Methodologies in Corps Hydrologic Models*.

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INTRODUCTION

Estimating watershed runoff in areas with seasonal snow cover requires a snowmelt algorithm be part of the modeling system. In recent years, advances in computing speed have made possible the implementation of more detailed one-dimensional snowpack models and distributed model techniques. A review of old and new algorithms is needed to determine which model elements should be chosen for use in current modeling systems, and what additional work is needed to adapt these new tools to water resource uses.

Previous reviews of operational snowmelt runoff models include a comparison of models by the World Meteorological Organization (WMO) (1986), the U.S. Army Corps of Engineers (1998), and a book on computer models of watershed hydrology (Singh 1995). These studies compared the structure and approach of snowmelt models, but did not attempt to judge whether one model performed better than another. The WMO comparison did not separate snowmelt algorithms from the hydrograph transformation and routing aspects of the models, but looked at the net result rather than the performance of individual parts of their modeling systems. In this review we look specifically at the snowmelt algorithms used in current operational model systems in the United States: their developmental setting, equations, strengths, and limitations.

All snowmelt algorithms consist of a surface energy balance and most also include some accounting of internal snowpack processes. These two parts of snowmelt algorithms follow simple to detailed approaches in the models reviewed here. Models requiring only minimal meteorological data input rely on air temperature data to define the surface energy balance, with precipitation amounts sometimes used to further define cloudy or clear weather. More-detailed surface energy balance models require air temperature, relative humidity, wind speed, and either cloud cover or radiation data. Internal snowpack processes may be simply ignored and the surface energy balance equated to snowmelt, or processes within a melting snowpack may be considered in detail. Which type of model is chosen for a particular basin or application will vary with data availability and the detail or type of model results needed.

Historically, operational hydrological models have been of limited detail both in the surface energy balance and in the internal snowpack processes. All of the models reviewed were developed initially from snowmelt studies at a point in snow laboratories or research watersheds. The point models were then extended to lumped basin scales, or more recently to hydrologic response units (Leavesley 1983) by using coefficients varied to capture the impact of winds or solar exposure, as dictated by landscape position or vegetative cover. With availability of fast computers, and digital elevation and land cover data, there is now ability to describe the surface energy balance in some detail across a landscape. This ability has led to the development of distributed snowmelt models. Today's computational efficiency also allows simulation of internal snowpack processes that have been mostly ignored in hydrologic models. A preliminary hydrologic model that accounts for both a varied surface energy balance across a landscape and details within a layered snowpack

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has been developed by Melloh et al. (1997). Over time we can expect to improve our ability to simulate and forecast snowmelt runoff with these new tools. Here we review the major operational models of past years and contrast these with two snowmelt algorithms developed at CRREL that are candidates for use in distributed operational models.

MODEL DESCRIPTIONS

HEC-1

HEC-1, a single-event model, uses either a degree-day temperature index method or a simplified energy balance approach. The equations used are a subset of those set forth in EM-1110-2-1406 (U.S. Army Corps of Engineers 1960), a document which summarizes equations derived from snow hydrology studies by the North Pacific Division, Corps of Engineers, and the U.S. Weather Bureau at the Central Sierra Snow Laboratory (CSSL) (latitude 39°22′N), near Soda Springs, California (U.S. Army Corps of Engineers 1956).

Temperature index degree-day method

The temperature index degree-day method uses the following equation:

$$M = C_d (T_a - T_b) \tag{1}$$

where M = snowmelt (in. day^{-1}),
C_d = degree-day melt coefficient (in °F^{-1} day^{-1}),
T_a = air temperature (°F),
T_b = base temperature at which melt will occur, normally 32°F.

EM-1110-2-1406 gives degree-day melt coefficients determined from mean values of snowpack ablation related to air temperatures at nearby stations in the CSSL. The temperature indexes of snowmelt were found to be more reliable for forested areas than open areas. The term "forest" pertains to the coniferous forests of the CSSL. Basins of significant open, deciduous, or sparse coniferous forest would be modeled less reliably with the index method. This is because temperature is a better sole indicator of the surface energy balance in the forest where the canopy diminishes the direct solar radiation and wind.

Energy balance method

The energy balance method in HEC-1 follows two equations, one for cloudy weather, and another for clear weather. Cloudy conditions are inferred from meteorological records, by whether or not rainfall was reported. When there is rain, cloudy conditions are presumed and melt is calculated as

$$Melt = C_1 [(0.029 + 0.0084kv + 0.007P_r)(T_a - T_b) + 0.09] \tag{2}$$

where $Melt$ = snowmelt (in. day^{-1})
k = basin convection–condensation constant (0 to 1, dimensionless),
v = mean wind speed at 50-ft height (mi hr^{-1}),
P_r = precipitation rate (in. day^{-1}),
T_a = temperature of saturated air at 10 ft (°F),
T_b = base temperature at which melt will occur, usually 32°F,
C_1 = coefficient used in HEC-1 to account for variation from the generalized snowmelt equation and is dimensionless.

This equation is from U.S. Army Corps of Engineers (1960) with coefficients fixed to assume 100% cloud cover and a canopy intermediate between open and dense forest. The convection–condensation constant (k) represents the mean exposure of the subbasin to wind, considering topographic and forest effects, and is fixed at a value of 0.6 in HEC-1. This coefficient should vary from 1 to 0.3 for unforested plains and dense forest, respectively. Corrections to wind and temperature measurement heights must be made externally to the program input. The first term (0.029) represents longwave radiation melt with complete cloud cover. The second term includes wind speed (v) and represents convection-condensation melt in open or partly forested areas. Term 3 is melt due to rain (P_r), and term 4 is the sum of 0.07 and 0.02 in./day melt due to shortwave radiation in the open and to ground heat, respectively.

The equation used when there is no rainfall, assumes no clouds, and 50% forest canopy cover. This is equation is also from U.S. Army Corps of Engineers (1960):

$$M = C_2 \begin{pmatrix} [k'(1-F)(0.004I_i)(1-\alpha)] + k(0.008v)(0.22T_a') \\ +k\,(0.008v)(0.78T_d') + F(0.029T_a') \end{pmatrix} \tag{3}$$

where M = snowmelt (in. day^{-1})

k' = shortwave radiation melt factor (dimensionless)

F = average basin forest canopy coverage (dimensionless),

I_i = solar radiation incident on a horizontal surface (langley day^{-1}),

α = albedo (dimensionless),

k = condensation-convection coefficient (dimensionless),

T_a' = difference between the air (10 ft) and snow surface temperatures (°F),

T_d' = difference between the dew point and surface snow temperatures (°F),

C_2 = coefficient used in HEC-1 to account for variation from the generalized snowmelt equation (dimensionless).

Term 1 represents melt due to direct solar radiation in the open, term 2 represents convection melt, term 3 is condensation melt, and term 4 is longwave radiation melt in the forest. The shortwave radiation melt factor (k') depends on average exposure of an open area in comparison to a horizontal surface and is assumed equal to 1.0 in HEC-1, implying a horizontal surface. The forest canopy coverage (F) is fixed at 0.5. Albedo (α) is reduced from 0.75 to a minimum of 0.4, using the inverse square of days since the last snowfall to account for factors that reduce albedo as the snow ages, such as increased snow grain size. The condensation–convection coefficient (k) is taken as 1.0, and the snow surface temperature is taken as 32°F.

Limitations

- The HEC-1 methods do not allow for important snowpack processes such as snow ripening, pore water retention, and flow of water through the pack.
- The use of wind speeds measured at 50 ft (15.2 m) is awkward, since many weather stations measure at 2- or 10-m heights. It is uncertain whether methods developed for wind speeds at 50 ft convert well to other measurement heights.
- The HEC-1 snow methods have been used more often in planning studies than in forecasting, and in situations where snowmelt runoff is not a primary contributor.
- Radiation, convection, and condensation coefficients that vary to represent a range of watershed conditions in EM-1110-2-1406 are fixed to midrange values in HEC-1.
- Output of modeled snow information in HEC-1 is limited to melt contribution to "rainfall" excess. Temporal changes in the snowpack depth or snow water equivalent are not given.
- Temperature lapse rates are fixed and cannot be varied with actual weather conditions.

SSARR

SSARR (Streamflow Synthesis and Reservoir Regulation) was developed by the North Pacific Division (NPD) Corps of Engineers beginning in 1956 to provide hydrologic simulations on snowmelt-dominated river systems for planning, design, and operation of water control works (U.S. Army Corps of Engineers 1991). SSARR was later expanded to provide operational river forecasting and river management for the Columbia River. SSARR was developed during a time when electronic digital computers first made continuous stream flow hydrograph simulation practical. The philosophy of the model developers was that limitations in data quantity and quality, and in development of fundamental relationships, prevented the development of all-purpose, physically based models. SSARR, thus, was conceptually based and of sufficiently limited detail to allow operational application on a daily basis. SSARR was one of 11 models from eight countries evaluated in a worldwide comparison of snowmelt runoff models (WMO 1986), and has been used extensively in operational snowmelt modeling in the Pacific Northwest. The basic snowmelt equations are the same as those used in HEC-1, though with fewer restricted coefficients.

Temperature index method

The degree-day, temperature index method is the same equation as used in HEC-1 (eq 1), except that the degree-day melt coefficient (C_d) can be varied during the model run on as much as a daily basis.

Energy balance method

The SSARR energy balance method equations are the same as HEC-1 (eq 2 and 3) with the following important additions:

- Fractional forest cover canopy (F) is not fixed, but can be varied from 0 to 1,
- k' and k can be varied,
- Albedo (α) can be specified.

Conditioning of the snowpack follows the work of Anderson (1973). A "cold content" is continuously accounted for using an antecedent index. The "heat deficit" must be made up by above-freezing temperatures before snow is allowed to melt and enter the soil system. A liquid water holding capacity of snow can be specified, and is usually taken as 2–5%. Snowmelt resulting from ground heat can be specified as a constant melt rate in inches per day, for each month.

Strengths of SSARR

- A canopy interception factor may be modeled as a reservoir and reduced by evapotranspiration. Interception quantity can be varied monthly to account for seasonal changes in vegetative cover.
- Gage-catch deficiency can be accounted for.
- The surface energy balance equation is for partly forested areas (eq 3) and contains coefficients that can be used to account for forest cover and solar angle effects on shortwave radiation, wind environment effects on condensation and convection, and forest cover effects on longwave radiation melt.

Limitations of SSARR

The albedo can be varied over the melt season, but must be provided.

NWSRFS

The National Weather Service's centralized library of computerized forecasting techniques is known as the National Weather Service River Forecast System (NWSRFS). The Generalized Streamflow Simulation System (Burnash et al. 1973), frequently identified as the Sacramento Catchment Model is a major component of the NWSRFS (Peck 1976). The snow accumulation and ablation routines used in NWSRFS are based primarily on the efforts of Anderson (1973, 1976) and Anderson and Crawford (1964). Anderson developed a combined energy and temperature-index method that used only temperature and precipitation as input meteorology. A difference from other temperature index methods is that each physical process is represented separately, rather than using a single melt index. Processes are conceptually modeled and include snowpack accumulation, heat exchange at the air/snow interface, areal extent of snow cover, heat storage within the snowpack, liquid water retention, lagged transmission of melt through the pack, and heat exchange at the ground/snow interface. A conceptual simulation of "cold content" and liquid water characterizes the condition or "ripeness" of the snowpack. The model also includes an index approach to dealing with frozen ground (Anderson and Neuman 1984).

The energy balance of the snow cover is expressed as

$$M = Q_n + Q_e + Q_h + Q_{Px} \tag{4}$$

where M = snowmelt,

Q_n = net radiation transfer,

Q_e = latent heat transfer,

Q_h = sensible heat transfer,

Q_{Px} = heat transfer by rainwater.

The major assumptions are

- The ratio between sensible and latent heat is given by the Bowen ratio,
- Outgoing longwave radiation can be calculated using Stefan's law with a snow surface temperature of 0°C during melt,
- Incoming solar radiation is negligible during overcast conditions,
- Incoming longwave radiation is equal to blackbody radiation at the temperature of the bottom of the cloud cover, assumed equal to the air temperature,
- Relative humidity is 90%,
- Atmospheric pressure can be computed from elevation,
- Saturation vapor pressure can be estimated as a function of air temperature.

Snowmelt during rain periods

Snowmelt during a 6-hr rain-on-snow period is described by Anderson (1973) as

$$\begin{aligned} M = {} & 3.67\times10^{-9}(T_a+273)^4 \\ & -20.4+0.0125P_xT_a+8.5f(u_a) \\ & [(0.9e_{sat}-6.11)+0.00057P_aT_a] \end{aligned} \tag{5}$$

where T_a = air temperature (°C),

P_x = water equivalent of precipitation (mm),

$f(u_a)$ = wind speed function at a height (z_a) above the snow surface (mm mb^{-1} 6 hr^{-1}),

e_{sat} = saturation vapor pressure at the air temperature (T_a) at height z_a,

P_a = standard atmospheric pressure at a given elevation (mb).

Humid, overcast conditions are needed for the above equation to apply; therefore, rain must exceed 2.5 mm during a 6-hour period before this equation is put to use. Rain falling on snow is added to surface melt.

Non-rain periods

During non-rain periods, a wide range of meteorological conditions can occur from sunny to overcast, dry to humid, and calm to windy. Since these conditions are not well described by air temperature and precipitation data alone, Anderson's recourse was the empirical relationship:

$$M = M_F\,(T_a - T_b) \tag{6}$$

where M = snowmelt (mm 6 hr^{-1})
M_F = proportionality, or melt factor (mm°C^{-1} 6 hr^{-1}),
T_a = air temperature (°C),
T_b = base temperature (°C).

M_F varies seasonally due to the increase in incoming solar radiation and decrease in albedo through spring. The melt factor is important in determining the timing of snowmelt runoff. At the Central Sierra Snow Laboratory, the melt factor could be represented by a sine function of the form

$$M_F = \left(\frac{M_{FMAX} + M_{FMIN}}{2}\right) + \sin\left(\frac{n2\pi}{366}\right)\left(\frac{M_{FMAX} + M_{FMIN}}{2}\right) \tag{7}$$

where n = day number beginning with 21 March,
M_{FMAX} = M_F on 21 June,
M_{FMIN} = M_F on 21 December (mm °C^{-1} 6 hr^{-1}).

This M_F sinusoid recommended for the contiguous United States was based on a snowmelt season at the CSSL (Anderson 1968). A modified sinusoid was recommended for Alaska, based on studies of the Chena River basin, near Fairbanks. Typical values of M_{FMAX} and M_{FMIN} are given for coniferous, mixed, deciduous, and open sites (Anderson 1973).

Non-melt periods

Energy exchange between the snow cover and air is tracked, during non-melt periods, as proportional to a representation of the temperature gradient in the top of the snowpack. The surface snow temperature is assumed equal to the air temperature. The temperature within the snow cover is represented by a continuous antecedent temperature index, computed from the air temperature time series,

$$ATI_2 = ATI_1 + TIPM(T_a - ATI_1) \tag{8}$$

where ATI_1 = temperature indices at the beginning of a 6-hr period (°C),
ATI_2 = temperature indices at the end of a 6-hr period (°C),
T_a = air temperature (°C),
$TIPM$ = weighting multiplier (0.1 to 1.0) for previous 6-hr periods.

Gain or loss of heat from the snowpack is assumed proportional to the difference between the current air temperature and antecedent temperature index.

Heat exchange during non-melt periods is, then,

$$\Delta D = NM_F(ATI_1 - T_a) \tag{9}$$

where ΔD = change in snow cover heat deficit expressed in water equivalent (mm 6 hr^{-1}),
NM_F = proportionality factor, referred to as the negative melt factor (mm°C^{-1} 6 hr^{-1}), which represents the rate of heat gain or loss per degree Celsius and time period, represented by millimeters of water equivalent.

The model assumes a seasonal variation in the negative melt factor NM_F represented by

$$NM_F = \left(\frac{M_F}{M_{FMAX}}\right) MaxNM_F \tag{10}$$

where $MaxNM_F$ is a snow model parameter representing the maximum negative melt factor (mm°C^{-1} 6 hr^{-1}). The negative melt factor thus increases through the melt season as the melt factor increases toward its maximum. The idea behind this is that thermal conductivity increases as a function of snow density, and snow density increases as the snowmelt season progresses.

Snow cover heat storage

The model keeps a continuous accounting of the heat deficit (D_m) of the snow cover, defined as the amount of heat that must be added to return the snow from below 0°C to an isothermal state (0°C), and computed as

$$D_m = -\left(\frac{P_x T_p}{160}\right) \qquad (11)$$

where D_m = heat deficit of snow due to mass change (mm),
P_x = water equivalent of precipitation (mm),
T_p = temperature of precipitation assumed equal to air temperature (°C).

The heat deficit consists of liquid water that will refreeze and snow that is at temperatures below 0°C, including new snow. Melt and rain continue to refreeze within the snow cover until the heat deficit reaches zero.

Retention and transmission of liquid water

A constant liquid water holding capacity is defined as the amount of liquid water that the snow can hold against gravity and is expressed as a percentage by weight of the solid (ice) portion of the snow cover. Excess water not held within the snowpack is both delayed and damped as it moves through the snowpack. The relationship used was that observed in April and May 1954, from the Central Sierra Snow Laboratory (CSSL) lysimeter. Though applied to all snow types, the observed relationship was observed in well-aged snow at 0°C with a spherical crystalline structure. Flow through the finer, drier snow would need to be lagged and attenuated more. The equation for lag is

$$L = 5.33\left[1 - e^{\left(\frac{-0.03WE}{E}\right)}\right] \qquad (12)$$

where L = lag (hr),
WE = water-equivalent of the solid portion of the snow cover (mm),
E = excess liquid water (mm 6 hr^{-1}).

A rate of outflow equation was determined by curve matching to lysimeter data, measured in English units,

$$R_1 = \frac{-1.0}{5.0e^{\left(\frac{-500E_{ls}}{WE^{1.3}}\right)} + 1.0} \qquad (13)$$

where R_1 = one-hour withdrawal rate (hr^{-1}),
E_{ls} = amount of lagged excess liquid water for the period (in.),
WE = water-equivalent of the solid portion of the snow cover (in.).

The amount of snow cover outflow it computed on an hourly basis is

$$O_S = R_1(S_1 + E_l) \qquad (14)$$

where O_S = snow-cover outflow (mm hr^{-1}),
S_1 = excess water storage at the beginning of the period (mm),
E_l = lagged excess liquid water entering storage during the current period (mm).

The change in liquid-excess stored in the snow cover is computed as

$$S_2 = S_1 + E_l - tO_s \qquad (15)$$

where S_2 is excess water in storage at the end of the period (mm), and t is the length of the time period (hr).

Melt due to heat exchange at the snow/soil interface is assumed in the range of 0.3 to 0.15 mm day^{-1}, and is added to the snow cover outflow.

Limitations

Processes not considered include vapor exchange due to condensation and sublimation, snow interception due to forest canopy, and redistribution of snow due to wind. The equation for lag of meltwater through the snowpack is based singly on the Central Sierra Snow Laboratory lysimeter observation.

SRM

The simple degree-day method, SRM, was developed in small European basins by Martinec (1975) to simulate and forecast streamflow from mountainous basins. A more recent, restricted degree-day version can utilize radiation input (Martinec 1989, Kustas et al. 1994, Brubaker et al. 1996). SRM uses remote sensing derived input of snow cover distribution for operational use, and this gives it a diagnostic capability beyond the simple degree-day method for basins where such operational snow cover maps are available. The model has been tested in over 60 basins worldwide, including number of basins in the United States (Rango 1995). SRM (along with SSARR) was one of eleven models from eight countries evaluated in a worldwide comparison of snowmelt runoff models (WMO 1986).

Degree-day method

The degree-day method is based on eq 16:

$$M = a\,T_d \tag{16}$$

where M = snowmelt (cm),
a = degree-day factor (cm day^{-1} °C^{-1}),
T_d = degree-days (°C day), the mean daily temperature over 24 hr, or the average of the maximum and minimum temperature over 24 hr.

The degree-day coefficient (a) for a site varies greatly over time as it implicitly represents all terms of the energy budget that account for the mass balance of the snowpack. The degree-day coefficient (a) can be evaluated over time by comparing degree-day values with the daily decrease in snow water equivalent. This can be done along snow courses, or when lysimeter data are available. Where such data are unavailable, a can be estimated as a function of snow density (Martinec et al. 1994, Martinec 1960).

Restricted degree-day radiation balance approach

The restricted degree-day radiation balance approach is as follows:

$$M = r\,T_d + m_Q\,R \tag{17}$$

where M = snowmelt (cm day^{-1}),
r = constant restricted degree-day factor (cm day^{-1} °C^{-1}),
m_Q = physical constant converting radiation to snow water equivalent [0.026 cm day^{-1} (W m^{-2})$^{-1}$]
R = net radiation in W m^{-2}.

The term including the restricted degree-day factor (r) represents melt attributable to turbulent energy exchange, while the second term converts net surface radiation (R) to depth of melt in snow water equivalent. Low r values occur when low winds reduce sensible heat transfer, and when low relative humidity increases latent heat loss due to evaporation (Kustas et al. 1994). Martinec (1989) showed r values that have much smaller variation than the original degree-day factor, ranging between 0.20 and 0.25 cm °C^{-1} throughout the ablation period. Brubaker et al. (1996) provided a method to estimate r from representative meteorological characteristics of the basin that requires wind speed, relative humidity, and air temperature, and is based on a simplified energy balance equation for snowmelt. In contrast to the degree-day method, the restricted degree-day factor (r) for a site is held constant throughout the snowmelt season.

Limitations

Rain infiltration or refreezing within the snowpack is not physically modeled in SRM. Early in the season before the snowpack is ripe, rainfall is assumed to add to the snowpack water equivalent. Precipitation is added to snowmelt when the snowpack is ripe. The user must specify when the snowpack becomes ripe based on judgment.

PRMS

The Precipitation Runoff Modeling System (PRMS) (Leavesley et al. 1983) was developed by the U.S. Geological Survey and is an application of a conceptual two-layer snowpack model developed by Obled and Rosse (1977). Obled and Rosse used Anderson's model (Anderson 1968) as a starting point for their model development. Data from open and forested lysimeter sites (located at 1350-m elevation, 15 km from Grenoble in the north French Alps) were used to calibrate and test the model, and a second lysimeter site (in Davos, Switzerland) was used to verify that the model worked at another site.

The snowpack is modeled as a two-layered system with a surface layer of 3- to 5-cm thickness. The snowpack mass balance is computed once a day and energy balance each 12 hours, representing night and day.

When the surface layer temperature (T_S) is below freezing (< 0 °C), non-melt conditions prevail, and heat transfer between the surface and snowpack occurs by conduction. When the temperature of the surface snow is at freezing (T_S = 0°C), an energy balance (I) at the air/snow interface is computed for each 12-hr period. If the energy balance is negative (I<0), there is no melt and the heat transfer occurs as conduction between the surface and bottom layer of snow. If the energy balance is positive (I>0), the available energy is used to melt snow in the surface layer and conduction is ignored.

Conduction between the snow layers is computed by

$$I_{COND} = 2\rho_s c_i \sqrt{\frac{K_E}{\rho_s c_i}\frac{\Delta t}{\pi}}\,(T_S - T_P) \tag{18}$$

where ρ_s = snowpack density (g cm^{-3}),
c_i = specific heat of ice (cal g^{-1} °C^{-1}),
K_E = effective thermal conductivity of snow (cal cm^{-1} s^{-1} °C^{-1}),

Δt = time interval (s),
T_S = temperature of the surface snow layer (°C),
T_P = temperature of the lower layer of the snowpack (°C).

Density of the snow (ρ_s) is computed daily using the procedure of Riley et al. (1973). T_P is calculated as a function of the modeled snow water equivalent and the calories needed to bring the snow to an isothermal state of 0°C. Heat transfer due to rain occurs as mass transfer using the average 24-hr air temperature as the rain temperature. When T_P is < 0°C, the meltwater or rain is refrozen in the pack and decreases the cold content of the snowpack. When the snowpack becomes isothermal, the meltwater is first used to satisfy the irreducible water content of the pack, and the remainder leaves the bottom of the snowpack.

The energy balance (I, cal) for 12-hour periods is computed as

$$I = I_{Ns} + I_{Nir} + (I_{COND,CONV}) \tag{19}$$

where I = energy balance at the air/snow interface (cal)
I_{Ns} = net shortwave radiation (cal),
I_{Nir} = net longwave radiation (cal),
$I_{COND,CONV}$ = approximation of latent and sensible heat (cal).

Net shortwave radiation, I_{Ns} (cal), is computed as

$$I_{Ns} = I_S\downarrow (1.0-\alpha)\, V_T \tag{20}$$

where $I_S\downarrow$ = incoming shortwave radiation (cal),
V_T = transmission coefficient for winter cover density of the vegetation canopy,
α = albedo of the snowpack surface.

Incoming shortwave radiation ($I_S\downarrow$) is measured or estimated from cloud cover.

The vegetation transmission coefficient (V_T) is based on relations presented by Miller (1959) and Vezina and Pech (1964). Albedo (α) is computed as a function of the number of days since the last snowfall, and whether the snowpack is accumulating or melting.

Net longwave radiation (cal) is computed as

$$I_{Nir} = (1.0 - V_{DEN})\,(Em\, L_{IR}\downarrow - L_{IR}\uparrow) + V_{DEN}\,(L_{IR}\downarrow - L_{IR}\uparrow) \tag{21}$$

where V_{DEN} = winter vegetation cover density above the snowpack,
Em = emissivity of the air,
$Em\, L_{IR}\downarrow$ = longwave radiation emitted from the atmosphere, calculated at the surface air temperature (cal),
$L_{IR}\uparrow$ = longwave energy emitted from the snowpack surface for the 12-hr period (cal).

Emissivities varying from 0.757 to 1.0 are calculated as a function of air moisture content (U.S. Army 1956). If humidity data are unavailable, *Em* is estimated. On days with precipitation due to frontal storms, *Em* is assumed equal to 1.0. During convective storms, *Em* is computed separately for each 12-hr period as a function of precipitation and ratio of observed to potential solar radiation. $L_{IR}\uparrow$ and $L_{IR}\downarrow$ are calculated according to the Stefan-Boltzmann law using the temperature of the snowpack surface, and air temperature, respectively.

The latent and sensible heat loss (cal) is computed using a temperature-index approach,

$$I_{COND,CONV} = C_M T_{AVG} \tag{22}$$

where T_{AVG} is the mean air temperature for the 12-hr period (°C), and C_M is a monthly convection–condensation parameter (cal °C^{-1}). $I_{COND,CONV}$ is computed only on rainy days, or when clouds cause observed solar radiation to be less than one-third of potential. Forested areas assume a value of $I_{COND,CONV} = 0.5$ on these days.

A limitation of the PRMS is that heat conduction from the soil to the snowpack is always assumed negligible.

SNTHERM

SNTHERM (Jordan 1990, 1991) is a one-dimensional mass and energy balance model that considers, in more detail, the processes that occur within a multi-layered snowpack. These include snow accumulation, compaction, grain growth, melt, condensation melt, advection, pore water retention, and water flow through the pack. In addition to being the most comprehensive of the models reviewed, SNTHERM has the most diverse scientific validation. Originally written to predict snow surface temperatures (Jordan 1991), SNTHERM has been applied to a range of snow types and scientific inquiries, including prediction of the spectral signature of snow at an alpine site in California (Davis et al. 1993), snow layer evo-

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lution during summer snowmelt on the Greenland ice sheet (Rowe et al. 1995), an energy balance study of a continental, midlatitude alpine snowpack at Niwot Ridge, Colorado (Cline 1997), and beneath canopy energy balance studies in the boreal forest in Saskatchewan, Canada (Hardy et al. 1997). The model is currently being applied to snow cover mapping in the boreal forest of Canada (Davis et al. 1997), snowmelt forecasting in Bosnia (Melloh et al. 1997), integration with mesoscale meteorological data (Melloh et al., in prep), and in distributed snow model studies in Sleepers River Research Watershed in northern Vermont (Melloh and Jordan, in prep.). SNTHERM is applicable to a full range of meteorological conditions such as snowfall, rainfall, freeze-thaw cycles and transitions between bare and snow-covered ground.

The meteorological boundary conditions in SNTHERM require air temperature, dew point temperature, wind speed, precipitation, and either incoming values of solar and infrared radiation, or cloud cover and site information (solar aspect and inclination of the surface). The surface energy balance is

$$I_{\text{top}} = I_{s\downarrow}\left(1-\alpha_{\text{top}}\right) + I_{\text{ir}\downarrow} - I_{\text{ir}\uparrow} + I_{\text{sen}} + I_{\text{lat}} + I_{\text{conv}} \tag{23}$$

where $I_s{\downarrow}(1-\alpha_{\text{top}})$ = downwelling shortwave radiation,
α = albedo or shortwave reflectance
$I_{\text{ir}}{\downarrow}$ = downwelling longwave radiation,
$I_{\text{ir}}{\uparrow}$ = upwelling longwave radiation,
I_{sen} = turbulent sensible heat flux,
I_{lat} = turbulent latent heat flux,
I_{conv} = heat convected by rain or falling snow.

Ground heat flux and soil temperature profiles are modeled, but soil moisture is kept constant. The steady-state bottom boundary condition is set by the initial soil temperature and moisture profile specified by the user.

SNTHERM was based initially on the mass and energy-balance snow model of Anderson (1976); it incorporates the mixture theory approach espoused by Morris and Godfrey (1979), Morris (1987) and Morland et al. (1990), and the technique for gravitational flow of water through the snowpack of Colbeck (1971, 1972, 1976, 1979). The water flow scheme is coupled to the equilibrium temperature in frozen strata using thermodynamically derived freezing curves for typical sand, silt, and clay soils. This conceptualization of water infiltration through the snow is one of an even, horizontal wetting front proceeding downward. In reality, finger flow occurs and tends to accelerate the arrival of melt to the bottom of the snowpack (Colbeck 1979, Marsh and Woo 1984), while capillary tension draws water along finer grained snow layers. Deformation of the snow cover over time takes into account settling due to metamorphism, and compaction due to overburden, melt, or sublimation. Vapor flux through snow is assumed driven by diffusion of saturated air and is computed by Fick's law. The residual water content of snow (irreducible water saturation) is assumed to be 4% of the snow pore volume.

The numerical solution is obtained by subdividing into snow layers, each represented by the governing equations for heat and mass balance. SNTHERM uses a control volume numerical procedure (Patankar 1980) for spatial discretization that allows for compaction of the snow cover. Use of the control-volume technique conserves the quantities over a finite control volume (ΔV) rather than at an infinitesimal point as with a finite-difference scheme. The rate of change of these quantities within a control volume ΔV must equal their net flow across the boundary surface plus their rate of internal production. As snow compacts over time, the one-dimensional grid is allowed to compress, so that volume elements continue to correspond with the original element of snow. The rate of flux is taken with respect to the deforming grid. A Crank-Nicholson central difference scheme is used to solve partial differential equations in the time domain. A new hydrologic version of SNTHERM limits the number of nodes in the bottom two-thirds of the snowpack while maintaining detail near the surface to gain efficiency for water resource applications.

The sum of the constituent bulk densities is the total density (ρ_t) written as

$$\rho t = \sum_{k} \theta_k \rho_k = \sum_{k} \gamma_k \tag{24}$$

where θ_k = individual volume fractions of k constituents,
k = ice (i), liquid water (l), water vapor (v), and air (a),
ρ_k = density of each i, l, v, and a constituent,

γ_k = bulk density of each i, l, v, and a constituent.

All four snow constituents are assumed to be in local thermal equilibrium within the snow medium. The basic set of equations developed in the model can simulate a full variety of snow types because this mixture theory is used consistently throughout the model.

Limitations

Where speed and simplicity are principal concerns, SNTHERM may not be suitable.

Strengths

SNTHERM is applicable to a full range of meteorological conditions such as snowfall, rainfall, freeze–thaw cycles, and transitions between bare and snow-covered ground. This algorithm provides much useful information about the snowpack condition that would be useful for runoff forecasting and has already been used in distributed format. Slope and aspect are model input parameters.

SNAP

The SNAP model (Albert and Krajeski 1998) uses a full surface energy balance to estimate melt water input to a one-layer snowpack. This model includes a new mathematical solution to the flow of water through the snow that is more physically based than current operational models, yet computationally efficient. The mathematical solution begins with the simplified form of Darcy's equation as set forth by Colbeck (1972) in which capillary flow in snow is considered negligible compared to gravity flow. Albert's method then diverges from earlier mathematical approaches (Colbeck 1972, Tucker and Colbeck 1977) in that it derives an analytical expression for water volume flux, and then evaluates the expression using a Newton's method approximation.

SNAP should provide more accurate prediction of the magnitude and timing of snowmelt than current operational models, none of which attempt to physically model the flow of water through the snowpack. Because SNAP solves an analytical expression for water volume flux through a bulk layer snowpack, it is expected to be more computationally efficient than the multilayered SNTHERM model.

Surface energy balance

The surface energy balance is equivalent to that of SNTHERM (Jordan 1991), as described above. The exchange of energy with the ground is considered insignificant.

The water equivalent of melt, equal to the available energy from the surface energy balance, is added to rainfall and routed through the snowpack.

Flow through a one-layer snowpack

The water volume flux equation is

$$\frac{\partial U}{\partial t} = -n\phi^{-1}(1-S_{wi})^{-1}\left[\frac{\rho_w k g}{\mu_w}\right]^{\frac{1}{n}} U^{1-\frac{1}{n}}\frac{\partial U}{\partial x} \quad (25)$$

where U = volume flux of water (cm s^{-1}),
t = time (s),
n = dimensionless effective saturation (S) exponent,
ϕ = dimensionless porosity of snow,
S_{wi} = irreducible water saturation of snow (% of total volume),
ρ_w = density of water (g cm^{-3}),
k = absolute permeability of snow (cm^2),
g = acceleration due to gravity (cm s^{-1}),
μ_w = viscosity of water (g cm^{-1} s^{-1}),
x = vertical spatial coordinate (cm).

Equation 25 assumes the effective saturation exponent (n), effective porosity (ϕ), irreducible water saturation (S_{wi}), and permeability (k) are constant over each time step, but may vary over the melt season. The variation of n and S_{wi} over time are not well understood, and in the present model version are held constant at default values of 3.3 and 3%, respectively. Melt volumes are assumed to travel as waves through the entire depth of a single-layer snowpack. The method allows for volume flux waves to absorb the residual mobile water from preceding waves and to determine when the combined meltwater flux wave will reach the bottom of the pack.

Grain growth and permeability (k)

Grain growth occurs over the melt season increasing permeability, and the rate of melt infiltration. Conceptually the snowpack is one bulk layer with a wet portion (in which the irreducible water saturation has been met) and a dry portion (which has either not yet been wetted, or has refrozen). The weighted averages of the two parts are taken as the average crystal volume within the pack (V_{av}), and used to compute grain diameter (d, cm):

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$$d = 2\sqrt[3]{\frac{3}{4\pi}V_{av}} \quad (26)$$

where V_{av} is the average crystal volume. Then, the absolute permeability of snow (k, cm^2) is

$$k = 0.077 d^2 \exp(-7.8\rho_s) \quad (27)$$

where d is snow grain diameter (cm), and ρ_s is density of water (g cm^{-3}).

Snow depth and effective porosity

The model will use either user-supplied snow depths or will predict snow depth from the rate of densification over time due to metamorphism and overburden (Albert and Krajeski 1999). Snow densification due to metamorphism is determined as a function of temperature, dry snow density, and fraction of the snowpack that is wet. Densification due to overburden is determined as a function of temperature and bulk density of the snow. Effective porosity changes are updated, accordingly.

Refreezing within the snowpack

SNAP uses an analytical solution of the Neumann equation (Carslaw and Jaegar 1959) to predict the depth of refreezing in the pack,

$$X = \sqrt{\frac{2k_1\left(T_f - T_s\right)t}{\rho l}} \quad (28)$$

where X = depth of the freezing front,
k_1 = thermal conductivity of 0.3 g cm^{-3} snow (0.0045 J s^{-1} cm^{-1} °C^{-1}),
T_f = temperature of fusiom (taken as 0°C),
T_s = surface temperature (taken as ambient air temperature),
t = time,
ρ = density of the snow medium,
l = latent heat of fusion (333.05 J g^{-1}).

Thermal conductivity of the snow medium is determined using a depth averaged value of saturation, and T_s is a time averaged value over the most recent period in which the snowpack is predicted by the model to be less than isothermal (<0°C).

Limitations

SNAP currently does not estimate radiation with cloud cover nor adjust for slope and aspect, though these additional subroutines from SNTHERM could be added to SNAP. The model, developed and tested at the lysimeter at Sleepers River Research Watershed near Danville, Vermont, would benefit from additional validations. The model is fairly new and has had little use to date.

Strengths

The model provides a quasi-analytical solution for routing meltwater and rainfall through a snowpack. This model is based more on physics than current operational models, yet takes less computation time than a complete numerical solution.

SYNOPSIS

Current operational models

The operational algorithms reviewed have a common basis in early snowmelt investigations of the North Pacific Division (NPD), U.S. Army Corps of Engineers and U.S. Weather Bureau (U.S. Army Corps of Engineers 1956, 1960), and the National Weather Service (Anderson 1968, 1973, 1976). These initial model developments were energy balance approaches simplified for operational use that defaulted to a temperature index method when only temperature and precipitation data were available. SSARR and HEC-1 are based directly on the NPD snow investigations in the Sierras. The SSARR generalized energy balance equations allow use of coefficients for forest cover, solar exposure, and wind speed, where HEC-1 fixes these coefficients at midrange values. Rain and non-rain periods in the meteorological record are used to distinguish heavily overcast conditions, since these have specific energy balance characteristics. A temperature index method is used in both SSARR and HEC-1 when only temperature and precipitation data are available. Snowmelt in HEC-1 is simply a surface energy balance with no consideration of internal snowpack processes.

In later versions of SSARR, NPD adopted some of the internal snowpack processes used in NWSRFS for "ripening" the snowpack. NWSRFS snow algorithms are based on Anderson's (1973, 1976) work in the Sierras and in Vermont, which drew from the work of Anderson and Crawford (1964) and the U.S. Army Corps of Engineers (1956). In contrast to SSARR and HEC-1, Anderson's NWSRFS operational algorithm computes snowmelt using a simplified energy budget for rainy periods rather than a temperature index method, but similarly defaults to a temperature

index method for non-rainy periods. The NWSRFS approach accounts for several internal processes within a one-layer snowpack including a lag for water flow through the pack. On the other hand, the NWSRFS energy balance equations do not directly allow for forest or solar angle variation, as the SSARR energy-balance equations do.

SRM and PRMS reflect the use of technological developments in the 1980s. SRM developers (Martinec 1960, 1989) took a pragmatic approach; they adopted a simple temperature-index method that also relies heavily on snow-covered area information. SRM is most suitable to basins where clear skies allow frequent satellite views of the snow-covered area. The PRMS algorithm (Leavesley et al. 1983, Leavesley 1989) was based on Obled and Rosse's (1977) two-layer snow model, the latter drawing from Anderson's work (1976). PRMS came later in the development chronology; this is reflected in its use of hydrologic response units (HRU), a concept that steps toward distributed snowmelt modeling by segmenting the landscape into land cover types that would melt similarly. The snow algorithm used in PRMS (Obled and Rosse 1977) reflects the observation that the rates of change of snowpack conditions in the surface layer of snow are more rapid than those in the bulk of the snowpack. PRMS considers more forest effects on snow interception and melt and is run within a modular software system at the USGS (Leavesley et al. 1992).

CRREL models compared to current operational models

SNTHERM and SNAP attempt to be fully generalized algorithms that require no calibration. For example, SNTHERM model validations in deep Sierra and relatively shallow New England snowpacks show good model performance without any regional adaptations or calibrations (Jordan and Melloh, in prep). SNAP, a one-layer model, has less detailed internal processes, making it more time efficient while offering a generalized approach to water flow through the pack. This no-calibration approach to flow routing through the pack, though it requires further validation of its generality, has particular merit in contrast to using a lag based on curve-fitting at a particular lysimeter site.

The three main differences between the operational and CRREL model approaches are the surface energy budgets, detail in the snowpack internal processes, and need for calibration. In all snowmelt models a surface energy budget drives the snowpack computations. When there are no internal snowpack processes considered, the surface energy balance is simply equated to snowmelt leaving the snowpack. The operational models vary from just surface energy budgets, to models that consider internal processes in one- or two-layer snowpacks (Table 1, Fig. 1). The generalized energy balance equations used in SSARR directly allow for forest cover, wind exposure, and solar exposure through coefficients in the surface energy balance. In many of the operational model approaches these environmental effects are imbedded in melt factors that must be judiciously selected or calibrated. In PRMS adjustments for forest canopy are made for net longwave and shortwave radiation. In SNTHERM, adjustment for solar exposure is integral to the surface energy balance, but forest canopy effects on winds and radiation are adjusted externally in the meteorological data stream. Similar solar aspect and canopy adjustments could be made available to SNAP.

Multilayered snowpack schemes were too numerically intensive to run on computers at the time these operational models were developed. SNTHERM is a full mass-energy balance model initially based on Anderson's mass and energy balance model (Anderson 1976), and expanded using mixture theory (Morland 1990) and Colbeck's routines for flow through snow (1972). SNTHERM models the condition of multiple snowpack layers and would be the model of choice when detailed information is desired. A new, streamlined version of SNTHERM that reduces the snowpack to a maximum of five layers may be more suitable to operational hydrology applica-

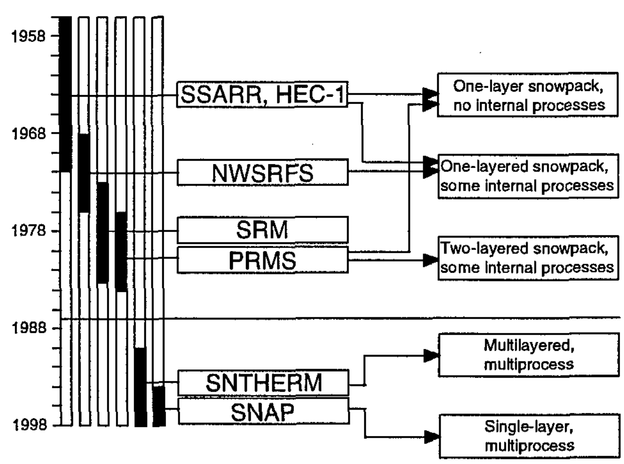

Figure 1. Chronology of operational snowmelt model development, including the two CRREL models.

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Table 1. Modeled processes.

	Snow interception by canopy	*Solar radiation through canopy*	*Snowpack temperature*	*Albedo*	*Cold content*	*Water retention*	*Rain on snow*	*Flow through snow*	*Multilayer snow*
HEC-1-T									
SSARR-T	x		x		x	x			
NWSRFS			x		x	x	x	x	
PRMS	x	x	x	x	x	x	x	x	2
SRM									
SRM-RAD									
HEC-1-E		50%		x					
SSARR-E	x	x		x	x	x			
SNTHERM			x	x	x	x	x	x	many
SNAP			x	x	x	x	x	x	

Notes:

NWSRFS and SSARR-E may run on 3- to 24-hr intervals.

PRMS runs on a 12-hr interval.

HEC-1-T and SSARR-T are temperature-based methods; HEC-1-E and SSARR-E are energy balance methods.

tions. SNAP is an energy balance approach using a one-layer snowpack, and offers a new mathematical approach to flow routing through the pack, aimed at saving computation time.

Data requirements

The reviewed snowmelt algorithms break into three groups based on input requirements (Table 2). The first group are temperature index methods that require only temperature and precipitation. These algorithms provide approximations of snowmelt that are best suited for forested basins (rather than open areas), without topographic variation (slope and aspect variability), but have predictive capability in all cases where temperature is a good predictor of snowmelt. The second group requires the additional use of satellite derived snow-covered area information, or radiation data. Net radiation is considered a better single predictor of snowmelt than temperature,

Table 2. Meteorological data requirements.

	Air temperature	*Precipitation*	*RH (%)*	*Dewpoint temperature*	*Wind speed 2 m*	*Wind speed 50 ft*	$I_s\downarrow$	$I_s\uparrow$	$I_{ir}\downarrow$	Q^*	*Snow cover area*
HEC-1-T	x	x									
SSARR-T	x	x									
NWSRFS	x	x									
PRMS	x	x					x*				
SRM	x	x									x**
SRM-RAD	x	x								x	x**
HEC-1-E	x	x		x		x	x				
SSARR-E	x	x		x		x	x				
SNTHERM	x	x	x		x		x†	x†	x†		
SNAP	x	x	x		x		x	x	x		

Notes:

* PRMS can substitute cloud cover information for incoming solar radiation.

† SNTHERM can substitute cloud cover, solar aspect, and slope for radiation observations.

** SRM requires snow-covered area maps derived from satellite data.

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but radiation data are often not readily available. The use of satellite data to determine snow-covered area can greatly augment simple snow models. In fact, in an operational system with (1) good satellite snow covered area data, (2) frequent model updates to match runoff conditions, and (3) experienced forecasters, the detail in snowmelt algorithms is a relatively less important aspect of the model system. In this context, the key to improved temperature-index modeling would be (1) fuller use of satellite remote sensing data, and (2) the integration of terrain data to better define melt patterns. This type of method would be more suitable for regions that are not frequently overcast, since this would interfere with the remote mapping of snow cover. The third group is models that require additional predictive data including, relative humidity, wind speed, and either radiation or cloud cover data. These full energy balance algorithms may be drivable with mesoscale meteorological model data forecasts, a data source that is likely to figure heavily in future forecasting approaches. Simplified energy balances based on temperature and precipitation could also be configured to drive the more detailed snowpack process models, SNTHERM and SNAP.

DISCUSSION

The more detailed snow models, SNTHERM and SNAP, are candidates for today's distributed and operational snowmelt models. Data handling is much less difficult today, and computers are much speedier than during the 60s and 70s when the existing operational models were developed. Though we have the electronic capabilities to take advantage of more detailed models and handle more meteorological data, the type and number of meteorological sensors accessible electronically may not have increased. Meteorology at high elevations in drainage basins has historically been underrepresented, and this is likely still the case. Current operational methods and models did, however, show significant shortcomings in simulating snowpack conditions during the upper Midwest (Red River at the north, and Missouri basins) flooding in 1997. Without the ability to incorporate full energy balance methods, or satellite snow-covered area maps, it may not be possible to forecast conditions beyond the range of "normal" meteorological conditions, as was suggested by Anderson (1976). This improved predictive capability of full surface energy budget approaches and more detailed internal process accounting should reduce the error in snowpack simulations and forecasting.

Current ability to forecast the weather constrains our ability to forecast snowmelt runoff. Future methods of estimating and forecasting snowmelt runoff for large river basins are likely to be driven by mesoscale meteorological forecast models. Tying snowmelt models with mesoscale meteorological model forecasts could be the key to improved short term snowmelt runoff forecasts within a decade. These meteorological forecasts are spatially continuous over a landscape at grid resolutions (10–100 km) compatible with distributed hydrologic modeling approaches and provide the parameters needed to drive an energy balance model of snowpack accounting.

In the interim, if there are not adequate data to run an energy balance model with more detailed snowpack accounting over an entire watershed, one could be run at a few key sites where data are available. Information on snowpack ripeness and onset of meltwater outflow provided by SNTHERM or SNAP could greatly aid forecasters. This, however, implies that good knowledge of what is going on at a handful of sites in a basin has strong implications for what is occurring on the basin as a whole. This leads to the idea of distributing point mass and energy models across drainage basins, large or small. The configuring and evaluation of distributed mass and energy balance models such as SNTHERM are the topics of current applied research at CRREL. The use of mesoscale meteorology models to drive snowmelt models is being investigated (Melloh et al., in prep), as well as optimal methods of segmenting basins into hydrologic response units for snowmelt (Melloh and Jordan, in prep., Davis pers. comm.*).

CONCLUSIONS

Data availability and time constraints will continue to drive the choice of surface energy balance models for a particular application; thus both temperature index and full energy balance methods are needed. Slope, aspect, and forest cover are extremely important to snowmelt; one need only look outside to see this demonstrated in real life. In the past only temperature and precipitation data were available to operational forecasters; today we also have digital terrain and forest cover data and our methods should take advantage of

* Personal communication, R. Davis, CRREL, 1998.

these. Methods to determine melt factors using digital terrain and forest cover data are needed where temperature index methods will be used in distributed snowmelt models.

The SNTHERM and SNAP models offer generality that is missing from the current operational models. First, use of a full energy balance to drive the snowmelt model reflects actual compared to the somewhat generalized conditions represented by temperature index methods. Secondly, adapting more physically based approaches to flow through a snowpack should permit wider applicability to a range of sites, and year-to-year variability. This is in contrast to earlier operational approaches of defining lag and retention equations by curve fitting based on one or two locations, and one or two study seasons. The SNAP snowpack routing technique is the most numerically efficient, and should be validated for generality by testing it at additional locations.

There is no single existing stand alone model that can be recommended for every application. In future watershed management systems, it would be best to make the surface energy balance and the snowpack internal process algorithms separate components. In this way, the choice of a surface energy balance technique could be made independent of the choice of the snowpack internal process technique. For example, improved snowpack routing through use of SNTHERM or SNAP does not absolutely require additional meteorological parameters; simplified surface energy balances can drive these snowpack models. Improvements are needed overall in the way forest canopies and other ground cover are accounted for in the otherwise robust surface energy balance approach used by SNTHERM and SNAP.

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One-dimensional snowpack algorithms in major operational snowmelt models used in the United States (HEC-1, SSARR, NWSRFS, SRM, and PRMS) are reviewed and contrasted with two U.S. Army Cold Regions Research and Engineering Laboratory (CRREL) algorithms (SNTHERM and SNAP) that are candidates for use in distributed operational models. In contrast to current operational models, the CRREL algorithms provide more detail in snowpack processes and require no calibration. The CRREL algorithms also include a full surface energy balance that requires more meteorological data than most operational models. Simpler surface energy balances could be used with the CRREL models. In future modeling systems, it would be preferable for the surface energy balance algorithms to be made independent of the internal snowpack process algorithms, so that available meteorological data can be used to drive a snowpack model of choice. Improvements are needed in the way that forest canopies and other groundcovers are accounted for in the surface energy balances of the CRREL models.

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