HYDROGEOLOGIC MODELING OF THE FLOW OF CATIONS AND ANIONS IN SELECT WATERSHEDS OF EASTERN CANADA WITH SPECIAL FOCUS ON SNOWPACK EFFECTS

by

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ABSTRACT

Streams in Eastern Canada appear to be subject to acid pulses, with some of these occurring during snowmelt seasons. A field sampling study was conducted to determine the general dynamics between snow accumulation and subsequent snowmelt and the snowmelt-induced acid pulses in stream water. This study revealed that snowpacks accumulate snow differently based on forest cover conditions (open, hardwood, softwood). Also, H^+ ions were found to leave the snowpack before most of the snow had melted.

Much work was done in terms of revising the forest hydrology model ForHyM, which simulates major water and energy flows through forest stands, including snowpack accumulation, snowmelt, soil infiltration, and stream discharge. The model now includes procedures to calculate snow density and amounts of ions released from the melting snow. Detailed regression analyses were done to empirically relate important hydrologic parameters (saturated hydraulic conductivity, porosity, field capacity and permanent wilting point) to soil conditions as affected by texture, organic matter content and depth of the soil. A user-friendly interface was built to facilitate calibration and model applications.

Data collected during field work in Fredericton, New Brunswick, and data from Kejimkujik National Park, Nova Scotia, were used to calibrate the ForHyM revisions. Specifically, model output was compared with corresponding field observations for snowpack depth, snowmelt chemistry, and stream discharge.

The model was used to simulate relations between H^+ pulses in snowmelt due to variation in atmospheric acid deposition and potential climate change. These simulations predict that decreased deposition for acidity will lead to proportional reductions in H^+ concentrations in snowmelt water. Also, warmer climate conditions up to 4°C will lead to smaller snowpack accumulations, due to less snowfall and more rain. This – in turn – should also lead to substantial reductions in the production of the snowmelt acid pulses.

Key words: hydrology, model, daily, snowpack, snowmelt, softwood, hardwood, open field, climate change, acid deposition, acid pulses, stream water.

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CHAPTER 1

INTRODUCTION

This thesis is about hydrological modeling, and snowmelt modeling in particular.

Specific objectives of the research include:

- To improve the Forest Hydrology Model ForHyM, in several regards: Firstly, to lay out all the basic assumptions into this model. Secondly, to perform a critical examination of the code and empirical equations used, and to improve on these to enhance model portability. Thirdly, to reexamine all the parameters that determine the retention and flows of water and heat in snow and soil.
- To obtain original data about snow properties and evolution of snowpack characteristics during the winter, under three contrasting forest environments (hardwoods, softwoods, open); these data include snow depth, snow density, timing of snowmelt, chemical composition of snowmelt and snowpack, and snow temperatures.
- To calibrate the model with the above field data and other data, and with year-round stream discharge data as well.

Based on these objectives, the thesis is constructed as follows:

- In Chapter 2 an overview of the issues faced during hydrologic modeling will be given, detailing the complexity of water and heat fluxes and their interactions in a natural environment.

- In Chapter 3, this is followed by a literature review concerning snow properties, mechanisms that occur in the snowpack and influence snowmelt, and ion release from the snow.
- Chapter 4 describes field work done to collect data about snowpack and snow chemistry.
- Chapter 5 presents general characteristics and other details about ForHyM.
- Chapter 6 details statistical work done to determine regressions between basic soil characteristics such as soil texture, depth and organic matter content with important hydrological parameters such as saturated hydraulic conductivity, porosity, field capacity, and permanent wilting point.
- Chapter 7 gives an explanation of all the improvements made to the model, the most important being the calculation of the parameters cited above, including all the required parameters that determine heat retention and heat flow, snowpack density, snowpack ion content, and ion fluxes.
- The calibration of the model is explained in Chapter 8, with data from field work done in Fredericton (UNB forest) and for Moosepit brook, near Kejimkujik National Park in Nova Scotia.
- Chapter 9 scenarios an application of model to assess effects of climate change and changing acid deposition on snowmelt hydrology and stream discharge, using Moosepit Brook near Kejimkujik National Park as a case study.
- Chapter 10 summarizes the work done with emphasis on what is new, and includes suggestions for further work.

CHAPTER 2

HYDROLOGICAL MODELING: OVERVIEW

2.1 INTRODUCTION

Any given landscape is under the influence of three major factors: precipitation (in the form of rain or snow), air temperature, and incoming energy from the sun. These factors are all important in terms of controlling vegetational growth and phenology, water storage (soil moisture, snow, ice), water fluxes (evaporation, transpiration, snowmelt, soil infiltration) and surface, snow and soil temperatures, for any particular landscape surface. This Chapter delineates the influence of these factors on general snow and soil conditions throughout the year, from year to year, and year after year. The particular purpose of this Chapter is to provide background for a detailed hydrological assessment of general soil and snow moisture and heat conditions on open fields and underneath closed forest canopies. Since this Thesis also deals with the modeling of chemical components within the snow, brief introductory references are provided for the accumulation and loss of chemical components within the snowpack.

2.2 THE WATER CYCLE

Water evaporates into the atmosphere from oceans, lakes, rivers, forests and soil surface. Water returns to the ground in the form of rain, snow, hail, fog, condensation, dew, etc. Some of the water is intercepted by forest and other vegetal canopies, and re-evaporates. Some of the water becomes surface runoff and/or percolates through soils and subsoil, and flows from there into bogs, marshes, brooks, streams, lakes, and rivers, back to the sea. Figure 1 illustrates the processes involved.

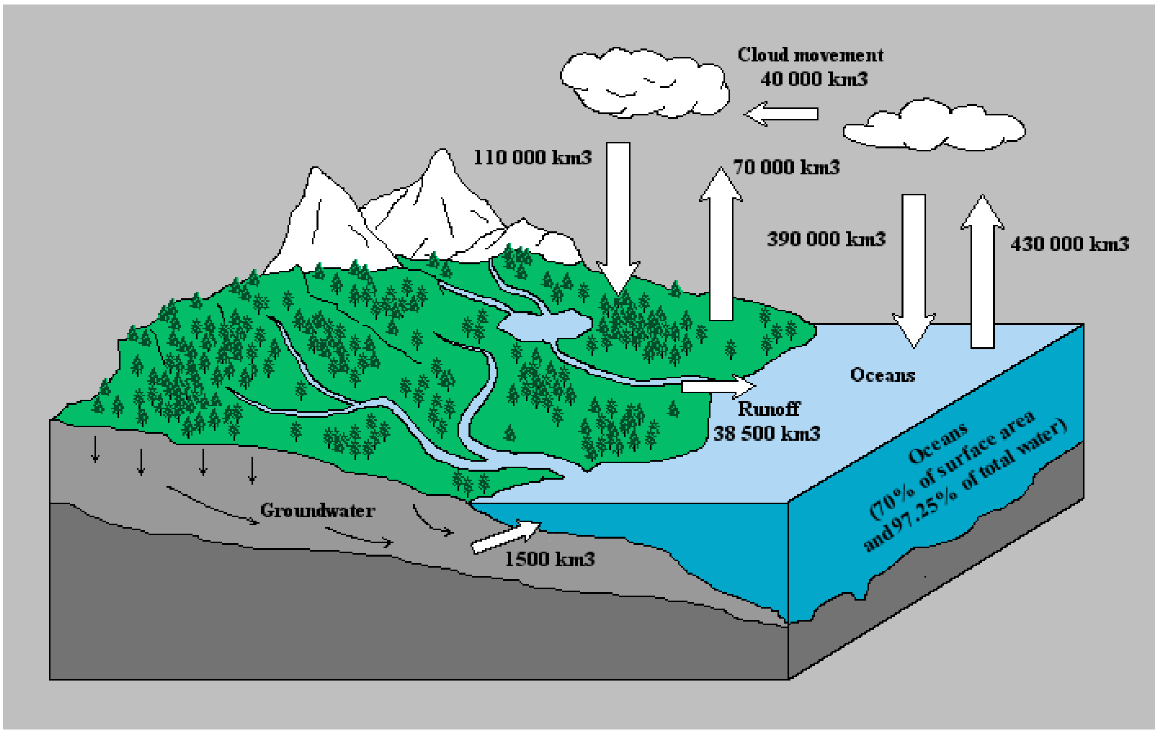

Fig. 1. The hydrologic cycle (after Brady and Weil 1999).

Hydrologic modeling attempts to simulate these processes that affect the water between the moment when water is deposited into a catchment, and the moment when the water flows out of the catchment. In this regard, hydrological modeling is guided by the laws of mass and energy balance, i.e., all water and energy inputs into a catchment must equal all water and energy outputs of that catchment, plus any water and energy quantities that are retained within the catchment. In addition, hydrological models attempt to quantify the timing and magnitude of the major water and heat flows and storage compartments within a catchment, and the timing and magnitude of all major water and energy flows into, through and from the catchments. In this context, the term **catchment**, or drainage basin, or watershed is defined as “the area that contributes all the water that passes through a given cross section of a stream. The horizontal projection of the area of a watershed is called the **drainage area** of the stream above the cross section ” (Dingman 2002).

The watershed concept provides ideal mass and energy units for landscape-level hydrological modeling. In theory, one can define the catchment area for any point along any stream. In practice, one often chooses the catchment area above the point where stream discharge measurements are taken regularly, or where the stream merges into another stream, or is crossed by a road. Figure 2 shows an example of a stream and its watershed terrain within the University of New Brunswick Forest, in Fredericton. This Figure was developed with the software *Arcview.*

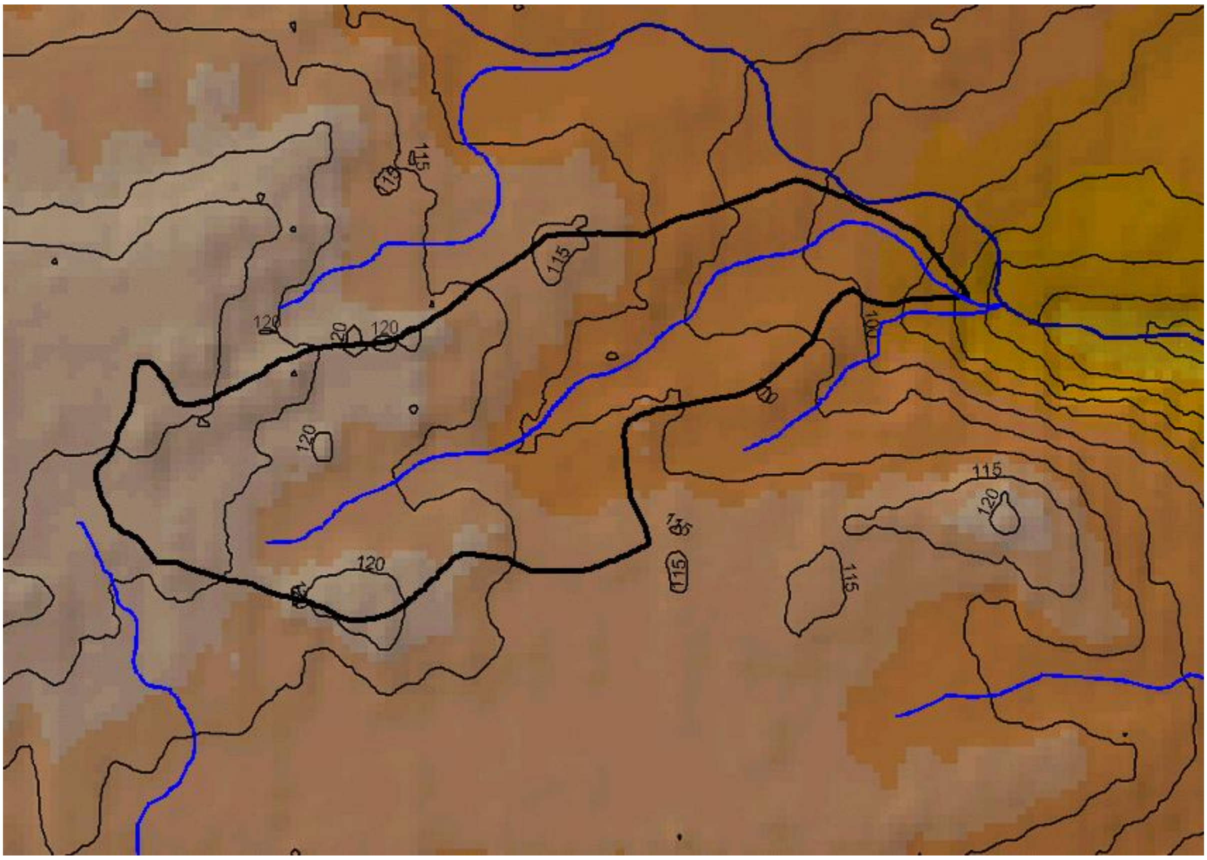

Fig. 2. Example of a watershed (thick line) in the UNB forest in Fredericton. Watershed area is 1 km^2. Also shown are the 5m contour lines.

Watershed sizes ranges from less than one hectare for a small feeder brook to thousands of square kilometers for major rivers. Between small to large watersheds, there is a continuum of gradual change in terms of water quality, water quantity, stream discharge, stream channel conditions an biotic communities. For example, for small streams, stream discharge peaks quite quickly after each precipitation event. For large streams and rivers, stream and river discharge is delayed and less peaked. Figure 3 shows the hydrographs from two nested watersheds in Nova Scotia, at Kejimkujik National Park: Moosepit Brook (17.7 km^2) and Mersey River (295 km^2).

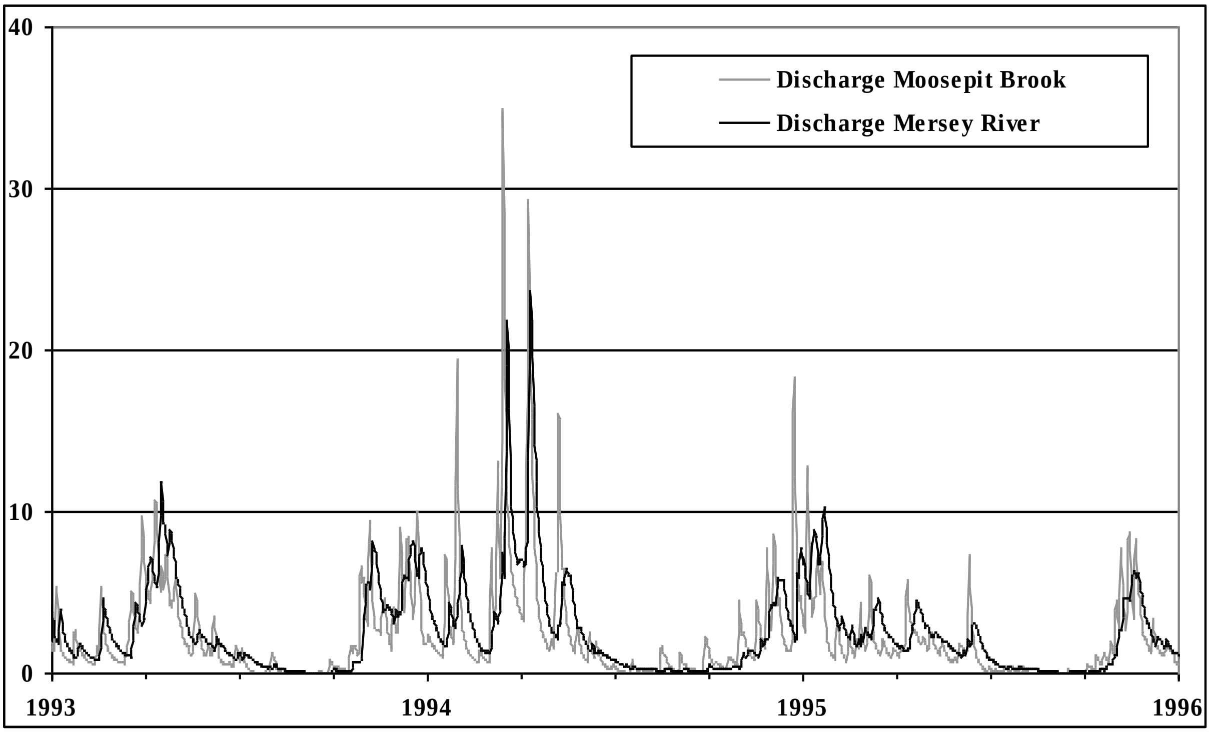

Fig. 3. Hydrographs (in mm day^{-1}) of Moosepit Brook and Mersey River, Kejimkujik National Park, Nova Scotia, Canada, from January 1st, 1993 to December 31st, 1995.

For scientific purposes, the preferred unit for watershed modeling is a watershed that is relatively uniform in terms of climate (low change of elevation from watershed divide to stream exit), vegetative cover (ideally represented by one forest cover type), substrate type, and substrate permeability.

Vegetation, especially forests, intercept some precipitation at canopy level. When the precipitation is rain, some of the intercepted rain evaporates back into the atmosphere. When it is snow, some of the snow may evaporate also, especially in arid regions. Otherwise, most of this snow falls to the ground. This process is encouraged by wind, rain, or melting due to sunshine and warm air temperatures.

Vegetation helps with water infiltration into the soil: some water flows from the leaves to the branches along the trunk and to the ground. This is called stemflow. In the forest, the vegetation builds the first layer of the soil through litter fall. This layer is called the forest floor and has a fairly low density, about 0.15 g cm^{-3}. At this density, the overall pore space within this layer amounts to 90%. This level of porosity enables high rates of water infiltration. The organic matter (OM) within the forest floor further assists by retaining water up to about 300% of OM mass. Gradual release of water from the forest floor allows gradual infiltration into the mineral soil below the forest floor. As a result, the forest floor acts as a sponge, and – in many areas – protects the underlying soil from surface erosion.

The forest vegetation further affects local soil moisture conditions by removing water from the soil through root uptake. That water is lost to the atmosphere due to transpiration from the leaves, especially during hot and windy days. The combination of water evaporation from soil and vegetative surfaces, and transpirational losses from leaves is called "evapotranspiration". During prolonged droughts, continued evapotranspiration can deplete soil moisture down to the permanent wilting point of the soil. The maximum amount of water that be taken up by the atmosphere at any point in time is called "potential evapotranspiration" (PET). This amount increases rapidly with increasing temperature and decreasing relative humidity (RH) of the air. When RH is 100%, PET equals zero. Specifically, "PET tells us how fast water vapor would be lost from a densely vegetated plant-soil system if soil water content were continuously maintained at an optimal level. The PET is largely determined by climatic variables such as temperature, relative humidity, cloud cover and wind speed" (Brady and Weil 1999).

The soil has a limited storage capacity of water, and it may take time for the water to infiltrate into some of the more compacted soils. If the soil is saturated with water, or if water cannot infiltrate into the soil fast enough, surface runoff occurs. Four parameters are necessary to determine soil moisture content and infiltration:

- Soil porosity: this is the amount of space in the soil that contains air and water. It is the maximum storage capacity of the soil.
- Field capacity: this is the water content that can be held against the force of gravity in the soil. It is used to calculate how much water remains in the soil 24 hours after soil saturation occurred.
- Permanent wilting point: below this point water content is not extractible by plants. When soil water content reaches this point, vegetation starts to wilt and die.
- Permeability (saturated hydraulic conductivity): this is a coefficient of proportionality between underground water flow rate and the hydraulic gradient that forces the water to flow through saturated soil (Darcy's Law). This coefficient is used to estimate how fast water can infiltrate into the soil, how fast water can percolate through the soil into deeper layers, and how fast water can flow sideways if deeper soils are blocked or saturated. With decreasing soil moisture content, soil permeability decreases rapidly, because water flow is increasingly confined to flow within thin water films on the surface of individual soil mineral grains.

In near-polar regions of the world, such as Canada, winters are long and seasonal snow stays on the ground for extended periods of time. As a result, streamflow is generally low in winter, but high during the spring snowmelt period. Stream flow decreases again in summer and early fall due to extensive evapotranspiration. Whether the soil is frozen or not also influences water flow through soils and watersheds, as shown in Figure 4.

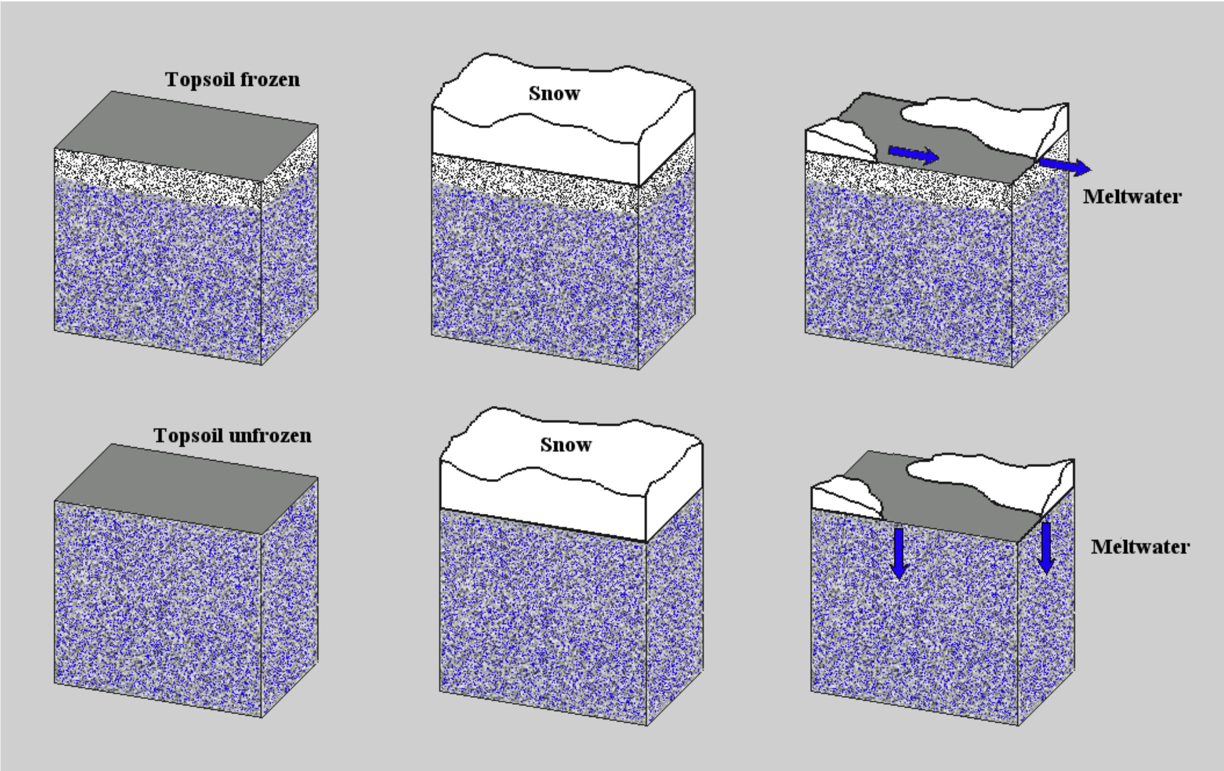

Fig. 4. Different paths followed by snowmelt water, based on specific soil conditions (after Brady and Weil 1999).

2.3 ENERGY PROCESSES

Vegetative surfaces and exposed soil surfaces receive energy and heat through various means:

- direct solar radiation
- diffuse (short-wave) and emitted (long-wave) radiation from the atmosphere
- heat transfer from the air due to conduction and convection

Figure 5 provides an example of the relative magnitude of these fluxes at the scale of the earth: out of 100% of solar radiation, 46% is quickly absorbed by the earth's surface (19% is direct solar radiation, and 27% is indirect, coming from the air and clouds), 32% is reflected or reemitted back to space by soil and atmosphere, and 22% is absorbed by the atmosphere itself, and some of this heat warms the soil by conduction and convection.

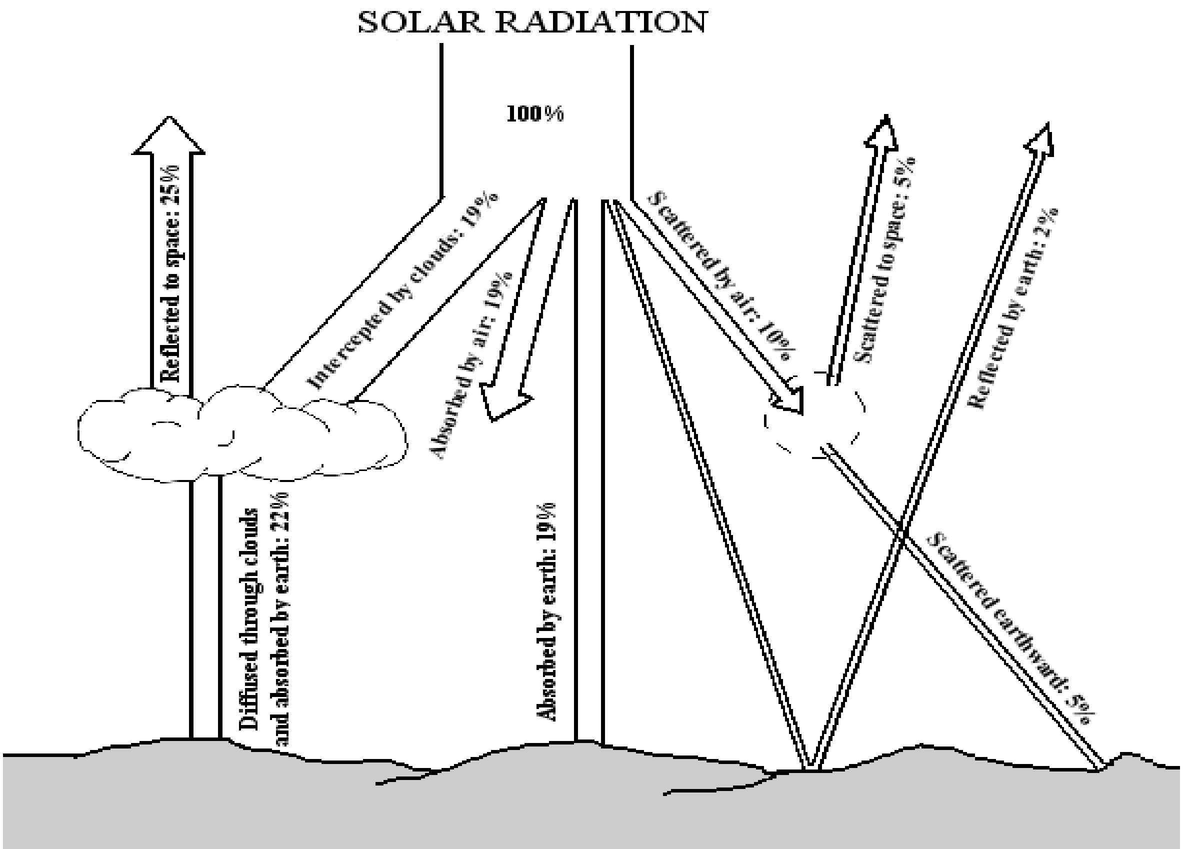

Fig. 5. Radiative energy inputs at the scale of the earth
(after Miller and Thompson 1975).

The heat absorbed by the earth is re-emitted to the sky through radiative processes (long-wave radiation) and non-radiative processes (conduction, convection, evaporation).

The changes in the air temperature and the radiation received by the earth's surface affect the soil temperature: energy exchanges happen continuously between the air and the soil. Figure 6 shows how temperature varies with depth during a year. In general, the temperature in the soil follows the surface temperature, but with a delay. The speed at which heat moves through the soil depends on the heat conductivity of that soil. The energy that is necessary to increase the soil temperature by one degree Celsius is called the heat capacity of the soil, and quantifies the energy accumulated in the soil. At a certain depth (e.g.,12 m), soil temperature is fairly constant year-round (Yin and Arp 1993).

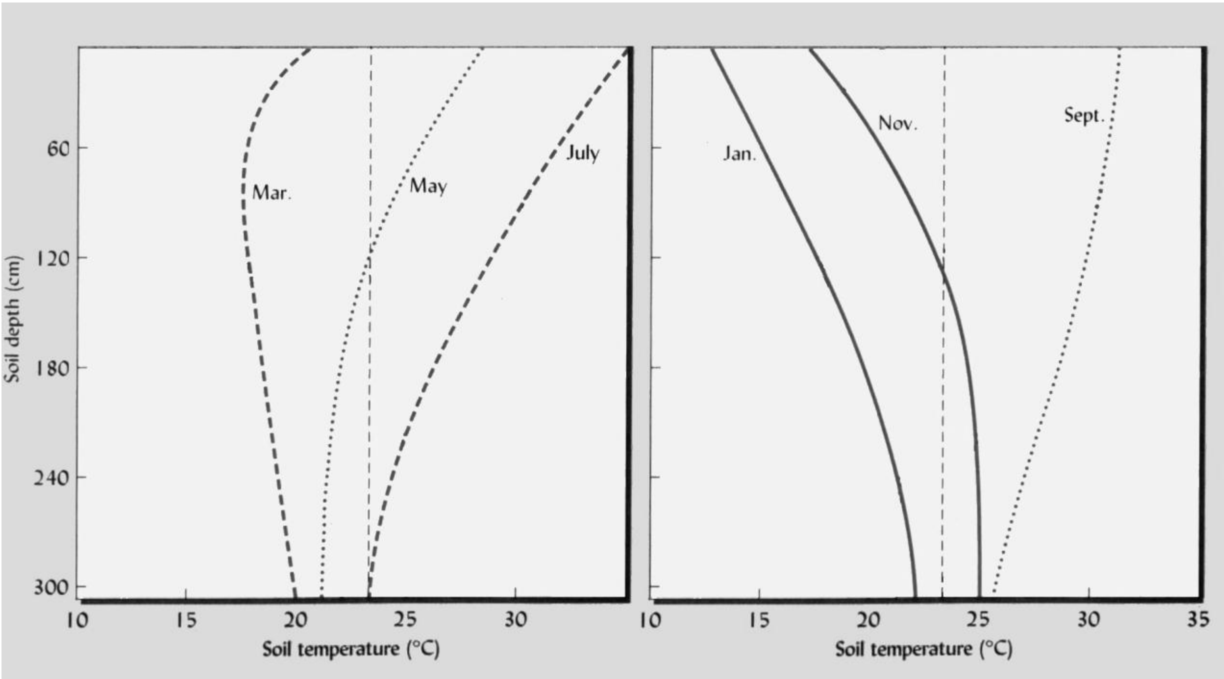

Fig. 6. Example of variations of soil temperature throughout the year (College Station, Texas) (after Brady and Weil 1999).

Trees intercept much of the incoming radiation, so that the air temperature is different inside the forest compared to outside. The interception depends on the leaf area of the trees. At the forest canopy level, this area changes with stand development. It also varies between summer and winter, especially for deciduous trees. What determines the development and growth of the leaves in the spring is the cumulative impact of cumulative degree-days above a certain threshold (4°C in average). Leaf fall at the end of the growing season is generally determined by the return of shortened day length and the resulting frost occurrences as these occur most predominantly in the early hours during cloud-free sky conditions.

When a snowpack is present on the ground, it acts as a thermal insulating layer for the underlying soil: having a snowpack on the ground limits the depth of soil freezing. In general, when the air temperature decreases below 0°C, and the soil reaches the temperature of 0°C, the water contained within the soil starts to freeze. Then, the soil temperature remains at 0°C until all of its water is frozen. Similarly, when a frozen soil starts to thaw, its temperature is equal to 0°C until all of the ice is totally melted. Only afterwards can the soil temperature become positive again. The energy that is required to melt ice is called the latent heat of fusion. Therefore, this latent heat of fusion also determines how much energy is released when water freezes. In principle, the temperature of a snowpack can only be at 0°C or less than 0°C. When the snow starts to melt, part of it stays at 0°C, and part of it may remain at less than 0°C.

The speed of the release of meltwater from the snowpack depends on the amount of heat energy flowing into the snowpack. In this, vegetation, aspect, shading, and surface reflectivity of the snow (snow albedo) all affect the rate of snowmelt because all of these determine the extent of radiative heat transfer, absorption and reflection at the snow pack surface during sunny days.

Heat capacity of the soil, and heat conduction into and through and from the soil depend on: soil texture, OM content, water content, ice content, and air content. This is because water, ice, air, soil minerals and OM all have varying heat capacities and heat conductivities.

2.4 CHEMICAL COMPONENTS WITHIN THE SNOWPACK

Polluting substances that are in the atmosphere arrive on vegetative surfaces, on soils and snowpacks through wet and dry deposition. When pollutants are part of the snow, these pollutants are not released until the snowpack starts to melt. However, when snowmelt starts, some of these pollutants may be released from the snowpack in a manner that is faster than the snowmelt process itself. Snow, compared to rain, may have a greater ability than rain to catch pollutants within the atmosphere. For example, Raynor and Hayes (1983) found that nitrite and nitrate in falling snow averages nearly five times the concentration found in rain. Daly (1995) found that there are two sources for nitrate in stream water during the snowmelt period: nitrate that accumulated within the snowpack, and nitrate that accumulated in the soil. In this Thesis, rate of loss of chemical components from the snowpack is examined in detail, by way of field observations, and by snowmelt modeling.

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CHAPTER 3

LITERATURE REVIEW: SNOWPACK AND SNOWMELT MODELS

3.1 INTRODUCTION

This Chapter reviews current knowledge about the characteristics of snowfall, snowpack dynamics and snowmelt, with special attention given to ion transfers through snowpacks, and related modeling requirements.

3.2 SNOW PROPERTIES

Snow distribution is irregular across the landscape, and is "a function of variation of climate, topography and land use within the river basin, which all can be related to spatial scales"(Bengtsson and Singh 2000). Spatial scales in snow hydrology vary from the size of each snow grain to regional scales at 1000 km or more. On large regional scales, general snow conditions are largely affected by climate.

Snow precipitation increases and temperature decreases with elevation, which means that precipitation in valleys may fall as rain, and may fall as snow higher up on hills and mountains. Stottlemeyer (2001) found that snow water equivalents in the Rocky Mountains gradually increases with elevation, and snow water equivalents are greater on northeastern aspects.

Some of the snow that falls onto forested areas lodges on the trees, and some of this snow evaporates, thereby resulting in less snow underneath forested areas than in open fields. Snow is dislodged from the canopy in various ways: mechanically by wind and subsequent branch and twig movement, and through partial melting due to rain, or solar radiation (Brandt 1986).

Snowmelt proceeds at a faster rate on areas exposed to solar radiation and wind than in shaded areas. Therefore, melt begins earlier and is faster in open areas compared to forested areas. Snowmelt is generally faster on south-facing slopes than on north-facing slopes.

Snowmelt water that reaches the bottom of the snowpack moves towards the exit of the watershed either as surface flow, or after infiltration and percolation through the soil, as groundwater. When the ground is frozen, the infiltration capacity of the soil is reduced, thereby increasing surface run-off during periods of intense snowmelt. Nyberg *et al.* (2001) noticed that soils containing more OM had shallower frost depth than mineral soil nearby, but also contained significantly more frozen water.

The runoff from small watersheds is quickly transferred to the streams, especially if there are no bogs, swamps, or lakes (Bengtsson and Singh 2000). As a result, response times in such streams are short, and fluctuations in stream flow are large, especially during the snowmelt season. In large rivers, there is usually a steady rise of the flow through the snowmelt period.

Bengtsson and Singh (2000) noted that processes that contribute to snowmelt may change during each day, being generally highest during the middle of the day and least during the night. The surface melt can vary from minute to minute, when clouds temporarily mask the sun. Considerable snow melting may occur during rain if rain temperature is positive and when the air is saturated, especially if part of the snowpack is already at 0°C. When rain falls on cold snow, all of the rain may freeze within the snowpack.

Before runoff can take place, the meltwater must percolate through the snowpack. Water cannot leave the snowpack until the liquid water content of the snowpack is above its water holding capacity.

In forested basins, snowmelt discharging from small watersheds may follow distinct diurnal patterns. In very small catchments, discharge peaks closely follow snowmelt peaks. In larger catchments, delays between snowmelt peaks and stream discharge peaks can amount to hours. In large rivers, diurnal variations due to snowmelt generally disappear.

Marsh and Woo (1984) studied water percolation into initially cold snow, and categorized the vertical percolation pattern into three distinctive zones, namely:

1. An upper zone of wet snow at 0°C.
2. An intermediate zone, where multiple wetting fronts (vertical " flow fingers") separate areas of wet snow at 0°C from dry colder snow.
3. A lower zone of dry snow with temperatures less than 0°C.

Marsh and Woo (1984) stated that the establishment of flow fingers in cold snow advances the timing of runoff from the bottom of the snow pack by a few days, as compared to a situation with no finger flow. Boggild (2000) observed that water release from cold snow needs to be fast in order to explain the rapid rise in the hydrograph associated with each spring flood event. He observed channel-like flow of melt water through the snowpack, as depicted in Figure 7.

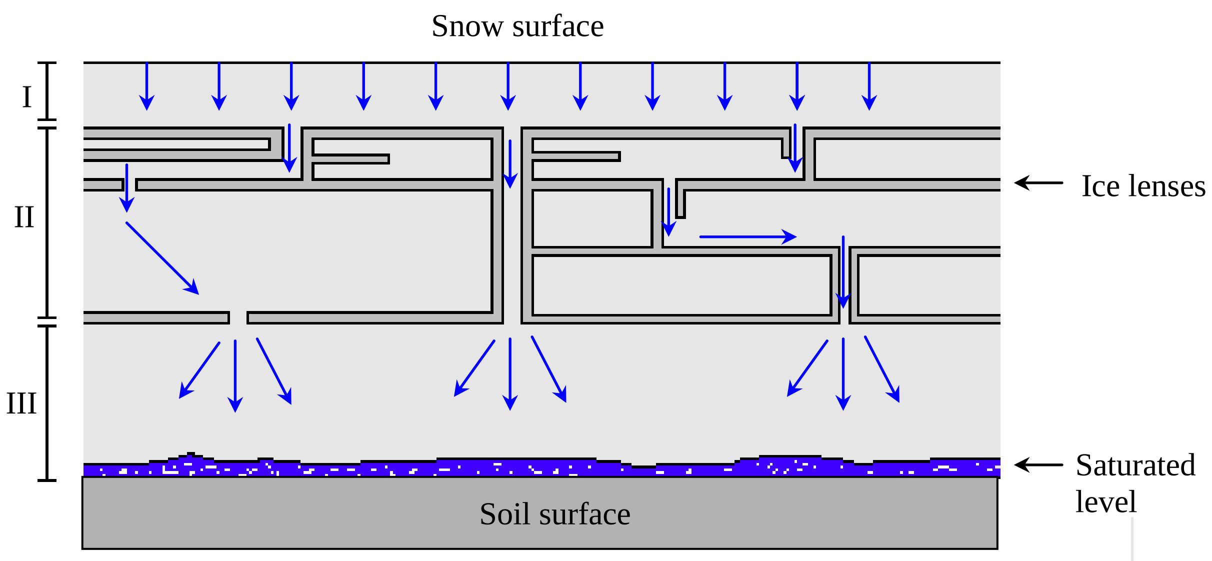

Fig. 7. Illustration of vertical flow pattern through icy snow packs (after Boggild 2000).

From the surface to about 10-15 cm down (Zone I), vertical flow is generally not concentrated into flow fingers. Below, in Zone II, water flows down along fingers (i.e., flow channels), and also along horizontal channels between individual ice lenses. Flow in the bottom zone (Zone III) is more dispersed. On the soil surface, lateral flow prevails because the underlying frozen soil is quite impermeable.

Boggild (2000) observed that when water is introduced into cold snow, two processes cause water retention, i.e., inter-grain retention of melt water and related re-freezing. As a result, snowpack density also increases.

3.3 ION PULSES THROUGH SNOWPACKS AND WATERSHEDS

Acid pulses in streams happen after major snowmelt events, or heavy rainfalls, and this can cause fish mortality. Recent work (Clair 1995) has shown that there are two distinct types of low pH events in the stream water of southwestern Nova Scotia. The first type occurs during late autumn when the stream base flow conditions are changed due to the onset of heavy rainfall. The second type occurs during winter and spring, due to snowmelt. Since there are many snowmelt periods, there are also many pH fluctuations in the stream water during this time. In both cases, Clair (1995) showed that pH values can decrease significantly within a few hours, and that these decreases are generally accompanied with sharp increases in discharge.

Cadle *et al.* (1984) noted that NO_3^- was displaced from the top of the snowpack during periods of rain and thaw, and Hendrickson *et al.* (1989) stated that concentrations of both NO_3^- and H^+ ions increased towards the base of the snowpack in response to rain and thaw.

Several researchers observed that when snow melts, the concentration of ions in the meltwater is not constant, but varies with time. Barry and Price (1987) noted that concentrations of NO_3^- and H^+ ions were highest in initial meltwater, and decreased more or less continuously until the end of the melt. Brimblecombe *et al.* (1986) observed that 80% of the chemical loading of the snowpack was released during the first 20-30% of each major melt period. Daly (1995) also found that snowmelt is responsible for major nitrate exporting from the snowpack.

Cragin *et al.* (1993) observed a similar phenomenon: while rinsing natural snow grains with de-ionized water, they noted that the initial 30% of meltwater contains 50-80 % of the total ions present in the snow. This phenomenon, now referred to as " fractionation" or " the ionic pulse", can contribute to episodes of "acid flush".

Cragin *et al.* (1993) found that ions like SO_4^{2-}, NO_3^- and Cl^- were not retained by snow if a solution containing these ions percolated through the snowpack. In another experiment, they found that ions concentrate on the surface of snow grains, while the interior of the snow grains becomes ion depleted. They offered the following explanation: “As the snow pack ages and vapor is transferred from warmer to colder crystals, the concentration of impurities increases for the shrinking grains until they have completely disappeared, leaving their impurities upon the surfaces of mature grains. At the same time, the larger grains are growing, rejecting chemical species within the newly developed layers of the ice lattice. This rejection is energetically favorable because impurities located on ice grain surfaces or at disordered grain boundaries cause less strain than if they are located within the ice matrix itself”. Figure 8 illustrates this point:

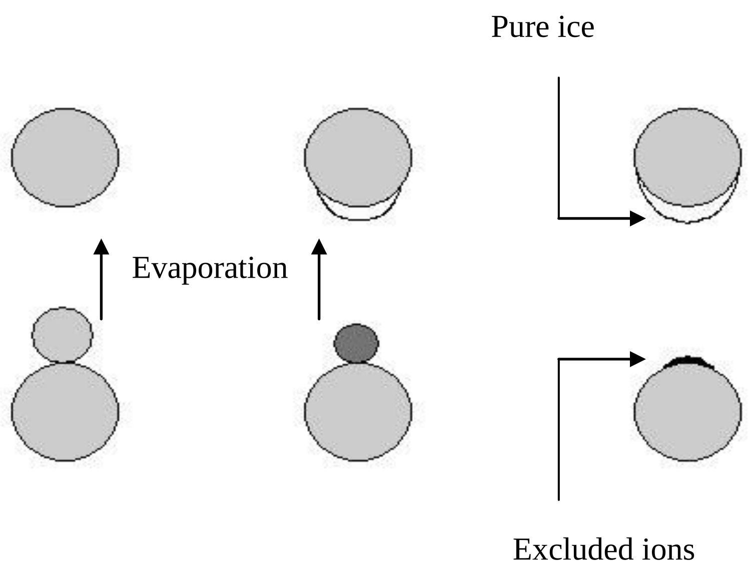

Fig. 8. Description of the phenomenon of ion exclusion within snowpacks (after Cragin *et al.* 1993).

In addition to the transformation of snow during the winter, the magnitude of the ionic pulse can also be influenced by diurnal melting and refreezing (Colbeck 1981; Suzuki 1991), rainfall on snow (Tranter e*t al.* 1992), and biological activity (Jones 1991).

Another phenomenon is preferential elution within the initial snowpack meltwater; i.e. not all ions leave the snowpack at the same rate. Cragin *et al.* (1993) observed that "less soluble chemical impurities, such as sulphate, are excluded more efficiently and therefore appear sooner and in higher concentrations in the eluate than more soluble species, such as chloride".

3.4 SNOWMELT MODELS

Bengtsson and Singh (2000) investigated how snowmelt and runoff can be modeled in relation to watershed size. A good simulation was obtained for large watersheds with a simple runoff and water storage calculations, and using the energy balance and the degree-day approach to simulate extent of snowmelt. In this, land cover and topography also needed to be considered. For small watersheds, which react quickly to variations in weather, snowmelt calculations need to be done on a finer scale, to capture hourly and diurnal variations. Storage on the surface, in the soil, in the groundwater and in small ponds and lakes must also be considered, because water storage within the basin attenuates the effects of fast precipitation and snowmelt events. For all of this, a good resolution of all spatial features within each watershed is required. For very large watersheds, it is more practical to divide a large basin into zones of different melt conditions than to know daily weather variations (Bengtsson and Singh 2000).

Clair *et al.* (2001) developed a statistical model to simulate frequency and intensity of acid peaks in streams in Nova Scotia, based on direct stream pH and stream discharge data and relationships, one for summer and fall, and one for winter and spring. The summer-fall regime generally has a poor correlation between stream discharge and ionic concentrations because of complex runoff-rainwater interactions. The winter-spring regime has better correlations. The data suggest that the larger the basin, the better is the pH/discharge relationship. Also, the smaller the basin, the greater are the day-to-day variations. With respect to extreme pH-discharge events, Clair *et al.* (2001) stated that: "The generation of simulated data by statistical approaches is not acceptable when the objective of the study is to study frequency of extreme events. Hydrological modeling may be more useful. The research showed that low pH episodes in Nova Scotia are not limited to a single spring melt period, but can occur at any time of the year in most basins."

Daly (1996) developed a model called SCATS (Small CAtchment Transport in Snowmelt) that he applied to the stream discharge ion pulse calculations for a small forested watershed in Vermont, with special attention given to nitrate ions. His study, however, suggested that nitrate peaks in stream discharge were likely not due to within-snow fractionation, but were more likely due to leaching processes in the soil. He also stated that "a relatively simple conceptual models using the *Stella* format may provide a useful tool for researchers in the interpretation of nitrate transport in snowmelt."

Boyer *et al.* (2000) used TOPMODEL for calculating snowmelt proportionally to temperature, and for simulating the flushing of dissolved organic carbon (DOC) from basins. The model accounted for heterogeneity in the landscape, and splits the watershed area into four elevation/aspect zones. In so doing, they were able to simulate that DOC peaks occur before stream discharge peaks. This is in agreement with other studies, showing that stream DOC levels increase rapidly just after spring melt, peak before maximum discharge occurs, then quickly decrease thereafter as melting continues. These authors added that "the results of the disaggregated TOPMODEL simulations for the four elevation/aspect zones show that the low elevation categories are capable of generating some of the early flows and the high elevation zones the later flows. Re-aggregating the stream hydrograph on the basis of the zonal simulation yields a reasonable reconstruction of the discharge hydrograph.” In this respect, simulation results improved considerably by introducing zonal weight factors to account for zonal snow accumulation differences among zones.

Melloh (1999) reviewed various snowmelt algorithms as currently found in the literature. He noted that the algorithms all have a surface energy balance module. Algorithms with least meteorological data input rely on air temperature to define the surface energy balance; precipitation amounts are sometimes used to account for the differences to the superficial energy contributions during cloudy weather versus clear weather. Detailed surface energy balance algorithms require air temperature, relative humidity, and wind speed, as well as cloud cover or radiation data. Simple algorithms tend to ignore processes internal to the snowpack. With the availability of fast computers, and with digital elevation maps and land cover data, it is now possible to describe the surface energy balance fairly accurately across a landscape. This ability has led to the development of “distributed snowmelt models”.

Below is a list of various snowmelt models.

HEC1 (US Army Corps of Engineers 1960)

This is a very simple model, but there are no simulations for snow ripening, pore water retention and flow of water through the pack. Temporal changes in the snowpack depth or snow water equivalent are also not calculated; cloudy conditions are inferred from meteorological records, when rainfall is reported.

SSARR, Streamflow Synthesis And Reservoir Regulation (US Army Corps of Engineers 1991)

In this model, canopy interception is considered, to calculate evaporation from snow and rain. Interception amounts are calculated monthly, to account for seasonal changes in vegetative cover. Its short-and long-wave surface energy balance equation deals with variable forest cover (leaf area index), and is adjusted for solar angle, and wind effects on condensation and convection. The albedo can be varied over the snowmelt season, but related changes must be provided as part of the input.

NWSRFS, National Weather Service River Forecast System (Anderson 1973, 1976; Peck 1976)

This model includes 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. Processes not considered include vapor exchange due to condensation and sublimation, snow interception due to forest canopy, and redistribution of snow due to the wind.

SRM, Snowmelt Runoff Model (Martinec 1975, 1989)

In this model, rain infiltration or refreezing within the snowpack are not simulated. Early in the season before the snowpack is ripe, rainfall is assumed to add to the existing snowpack water equivalent. The user must specify when the snowpack becomes ripe based on judgment.

PRMS, Precipitation Runoff Modeling System (Leavesley *et al.* 1983)

In this model, the bottom of the snowpack has a negative temperature, and the meltwater or rain is refrozen in the pack and decreases the cold content of the snowpack. When the snowpack becomes isothermal, meltwater is first used to satisfy the water storage capacity of the snowpack; the remainder leaves the bottom of the snowpack. Heat conduction from the soil to the snowpack is assumed negligible.

SNTHERM (Jordan 1991)

This is a one-dimensional mass and energy balance model that simulates the following processes in a multi-layered snowpack: snow accumulation, compaction, grain growth, melt condensation melt, advection, pore water retention and water flow through the pack. Required model input includes air temperature, dew point temperature, precipitation, wind speed, and solar radiation or cloud cover. Slope and aspect are additional input parameters. Each layer within the snowpack has its own heat and mass balance. The model uses a numerical procedure for volume control (Patankar 1980) to allow for snow compaction.

SNAP (Albert and Krajeski 1998)

This is a one-layer snowpack model. Grain growth occurs over the melt season, which increases the overall snowpack permeability.

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CHAPTER 4

MEASUREMENT OF SNOWPACK PROPERTIES

4.1 INTRODUCTION

For the purpose of snowpack and snowmelt modeling, and given the general lack of simultaneous information about snow, snowpack depth, density, ion content, and snowmelt within the same snowpack, it was decided to obtain such information through a special case study. This case study was done within the UNB Forest, in the fall and winter of 2001/2002. Preliminary work in the preceding winter determined the final sampling strategy for obtaining the needed snowpack data. Three contrasting sites were chosen in close proximity to sample from extreme surface conditions above the snowpack: one mature softwood site, with a dense spruce/fir canopy above the ground (mainly red spruce and balsam fir), one mature tolerant hardwood site (mainly sugar maple, beech, yellow and white birch, aspen), and one open, clear-cut site. The terrain at each location had a pronounced mound and pit surface topography. This topography was caused over time by occasional wind-induced uprooting of tall trees: each time such a tree is uprooted, it leaves a gouge in the soil, and a mound next to the gouge. As a result, the entire area had a distinctive mound and pit appearance, with each pit forming a mini-catchment between less than 5 to about 10 m^2. At each of the three sites, several such mini-catchments were selected for snowpack characterization and for collecting snowmelt. This Chapter provides details and results generated from this field study.

4.2 STUDY AREA

Figure 9 shows the locations of the sampling locations within the UNB forest, using the local digital elevation map for background. Also shown on this map is the local network of highways (white), forest roads (black) and streams (blue). At each location, four mini-catchments were selected for snowdepth and snowmelt sampling. Two additional sites near the mini-catchments were used for profiling the snowpack at each of the three locations for snow density and ion content profiling, from the top to the bottom of the snowpack.

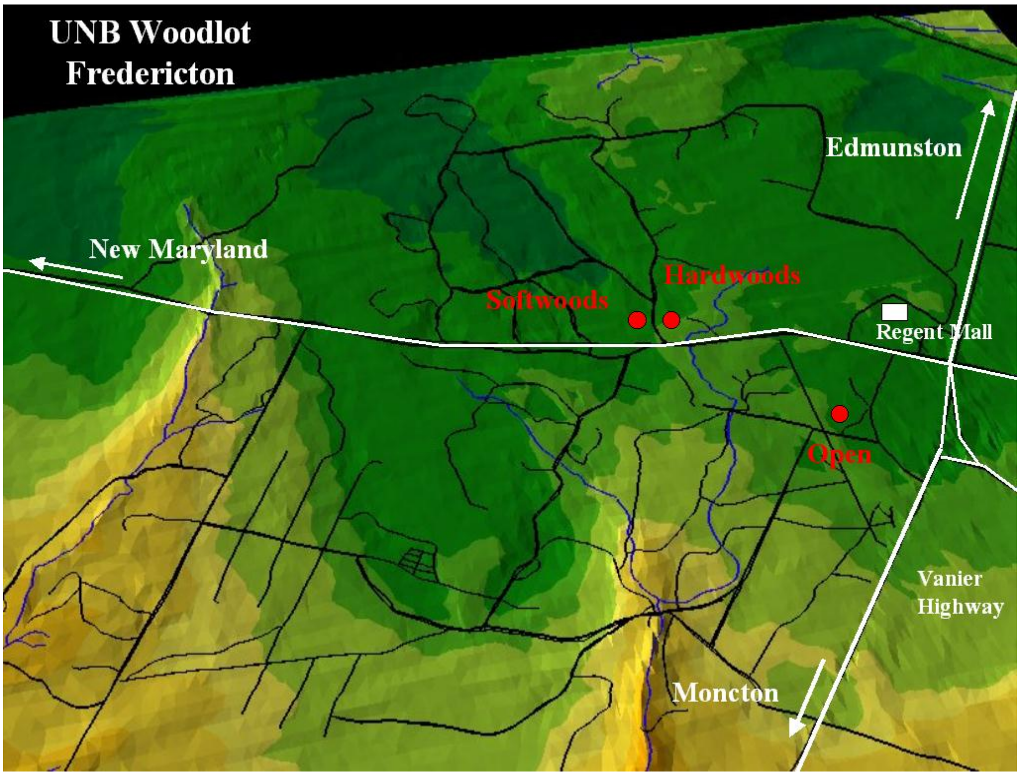


Fig. 9. Location of the snow sampling sites, UNB woodlot, Fredericton.

4.3 SITE PREPARATION AND SAMPLING METHODS

Preliminary measurements in 2000/2001 revealed that the mini-catchments needed to be selected according to a number of criteria:

- Mini-catchments needed to be well drained, with fairly regular slopes.
- Mini-catchments needed to be close to one another to ensure similar snowpack features.
- Mini-catchments needed to be within walking distance from a main forest road, to facilitate sampling of all desired snowpack parameters within a few hours after any snowmelt event.

Altogether, 12 sites were chosen among the natural mini-catchments for collecting snowmelt. Plastic sheets (polyethylene drop sheets) were draped across each mini-catchment to catch the snowmelt (Figure 10). The snowmelt that would accumulate on top of these sheets was then directed to flow towards the exit point of each mini-catchment. At this point, a hole was dug for bucket placement and snowmelt collection. The buckets were half-buried in the ground before the soil was frozen. A plastic polyethylene bag was placed into each bucket to catch the snowmelt, and to remove the snowmelt from the bucket. A hole was made in the side of each bucket, to admit the end of a small trough (half of a small diameter plastic plumbing pipe) to drain the plastic sheet within each mini-catchment into the designated bags (Figures 11 to 13). Snowmelt samples were retrieved from each mini-catchment after each snowmelt event. At that time, collector bags were replaced, to continue with the snowmelt sampling until the end of the snowpack season. Events that led to measurable snowmelt accumulations within the snowmelt sampling bags occurred when rain fell on the snow, and when the air temperature went considerably above 0°C for an extended period of time (several days). Amount of snowmelt was determined, and snowmelt samples were also retained for ion analysis.

Fig. 10. General setup for snowmelt collection.

Fig. 11. Detailed view of the snowmelt collection system.

Fig. 12. Snowmelt-collecting site covered with snow.

Fig. 13. Illustration of water dripping from the bottom of the snowpack during a snowmelt event.

Graduated metallic bars were placed at each of the 12 sites before the first snowfall, so that each mini-catchment could be easily located when covered with snow. Also, the bars were used to ascertain depth of snowpack, in cm.

Snowpack profiling involved sampling of snow temperature, snow density, pH and snow ion contents from top to the bottom. This was done weekly. For this purpose, additional plastic sheets were laid on the ground at each of the three sites (hardwood, softwood, open area) prior to the snow season. These sheets were needed to ensure that the snowpack would not be contaminated from the soil. Preliminary observations showed that the bottom of the snowpack often contained many plant fragments (leaves, twigs, seeds) and other debris from the forest floor below. These fragments and debris, in turn, would influence the pH, the dissolved organic carbon content (DOC) and the ion content. Physical conditions at the bottom also differed, consisting mainly of coarse-grained ice rather than snow. Snowpack profiling of temperature, density, pH and ion contents was done weekly, as follows: twice near each mini-catchment, with one profile done above the extra plastic sheets, and one profile done above areas with no plastic sheets underneath, to ascertain the extra ion loading of the snow pack from the soil below.

Snow temperature was determined with an electronic thermometer, equipped with a long measurement tip, to measure temperatures up to 12.5 cm away from an exposed snowpack surface. Temperatures were measured by exposing a vertical snow profile, and by taking temperature readings every 10 to 15 cm from top to bottom. Snow density was also determined on the same snowpack profile, by gently sliding a cylindrical jar into the exposed face horizontally, until the bottom of the jar coincided with the exposed surface. The jar was then withdrawn, and was capped to keep the snow within the jar. This process was repeated with new jars every 10 to 15 cm from top to bottom. Snow density was determined by dividing weight of snow within the jar with the inside volume of the jar.

Additional snow samples were taken from the same spots and gathered into pre-labeled polyethylene freezer bags, to obtain amounts sufficient for detailed ion analyses and pH measurements. This sampling was also done once a week, at two locations: above the extra plastic sheets, and above areas with no plastic sheets underneath, to ascertain extent of ion loading of the snowpack from the soil below.
All of these snow samples were analyzed for pH, and all major anions (Cl^-, NO_3^-, SO_4^{2-}, $H_2PO_4^-$) and cations (Ca^{2+}, Mg^{2+}, K^+, Na^+). The snowmelt that was retrieved from each mini-catchment was analyzed in the same way.

Anions were determined by way of high performance liquid chromatography, using ion exchange analysis columns (Dionex system). Cations were determined by way of Atomic Absorption Spectrophotometry (AAS). The pH of the snow samples was determined potentiometrically, with electrode calibrations done for pH 4 and pH 7 buffer solutions.

Each of the samples that were retrieved from the field was filtered in case there were visible signs of debris. Also, each snow sample needed to melt completely, to ensure that the results would be consistent. Insufficient melting led to underestimates of ion content, and overestimates of pH. For the pH measurements, test solutions were equilibrated to room temperature.

Systematic snowpack and snowmelt sampling started on January 23rd 2002. When rain was forecast, snowdepth measurements were taken before and after the rain, to determine the effect of snow melting on snow depth and on snow density. These measurements were also done after days of high air temperatures.

At first, the accumulated snowpack profile of the snowpack had no visible layers. First rain (a few millimeters) of each precipitation event was usually trapped by the snow, and no runoff occurred. Snow depth, however, diminished, the density of the snowpack increased, and the rain caused the formation of a layer of coarse-grained ice crystals in the top part of the snowpack. Other rainfalls that occurred later infiltrated deeper in the snowpack, with some of them causing runoff. When this happened, the snow was wet along the whole profile, and had an even temperature of 0°C.

Runoff sampling problems occurred in some of the mini-catchments when water started to freeze on top of the plastic, forming an ice dome at the bottom of the snowpack. This prevented runoff from entering the buckets. For some of the sites the solution was to break this ice because the ice layer was not too thick. Two of the twelve sites had so much ice covering at the exit so that these locations did not yield runoff until the ice started to melt at the end of the snowpack season.

Another problem occurred when substantial amounts of snowmelt started to flow in the spring, all at once. At these times, water overflowed from the plastic bags, and filled the hole outside the buckets. As a result, the buckets were pushed out of the soil. This was fixed by placing water-filled buckets on top of each buried bucket.

4.4 RESULTS

Average snowdepth measurements are shown in Figure 14. Note that the hardwood site accumulated more snow than the other two sites, possibly for several reasons:

- Often, more snow is trapped in depressions than on ridges (the hardwood site was located on a side of a local depression).
- Hardwood canopies present rougher surfaces than softwood surfaces or open areas. This would induce local turbulence which would increase overall snow catch below the hardwood canopy, especially in forests that are patchy, i.e., vary from softwood cover to hardwood cover, as was the case in this situation.
- The hardwood site was on a shaded side of a valley, rather than on a ridge. Partial shading would keep the air temperatures immediately above the snowpack cooler than at more sunny locations. This would lower the local rates of snow melting and snow re-crystallization.
- At night, cool temperatures would continue because of direct heat losses through the highly transparent hardwood canopy.

There would be less snowpack accumulations underneath colder softwood canopies because of snow interception by the canopy. Also, snow and rain dropping from the softwood canopy would impact the snowpack below, making it denser than otherwise.

At one point in the winter, snow nearly totally disappeared in the open area, and only a few patches of snow were left. This was very likely due to high exposure to solar radiation, and also due to faster heat transfer from the atmosphere into the snowpack because of higher wind speeds near the ground. On sunny days, measurements showed that the snowpack in the open area often warmed up to 0°C, often yielding water dripping from the snowpack into the buckets. During rain, water dripping would occur even faster.

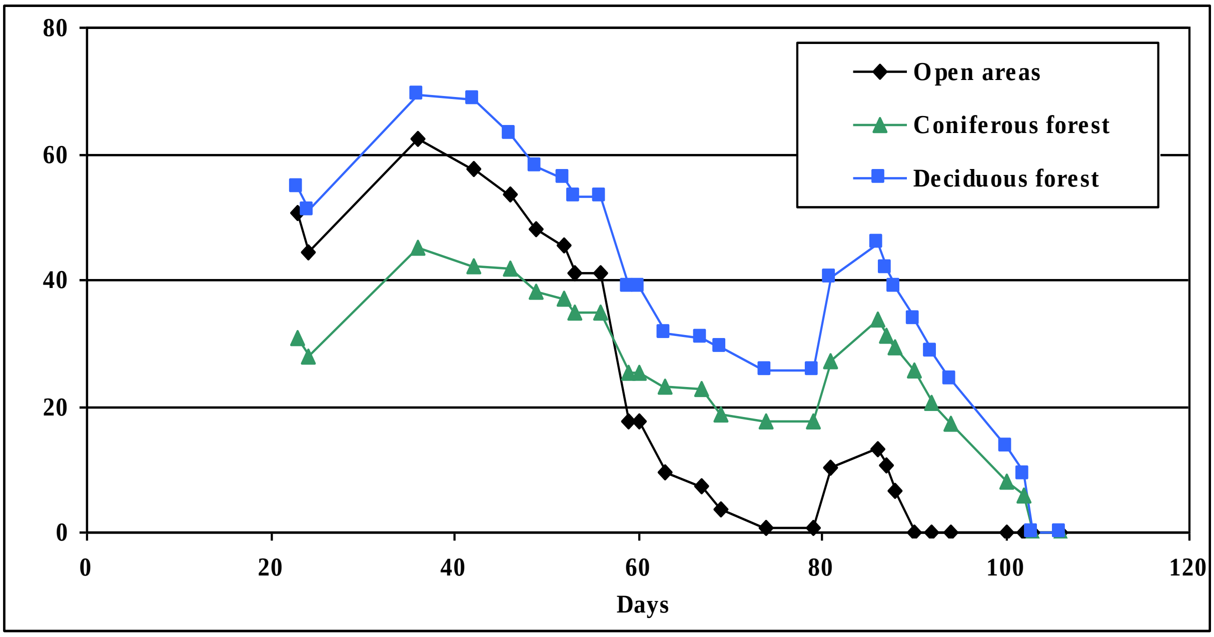


Fig. 14. Average measured snowdepth (in cm) in and outside of forest for winter 2001-2002.

Figure 15 illustrates the evolution of snowpack density during the winter. The low values recorded for days 39 and 81 correspond to freshly fallen snow at the top of the snowpack. The measured density of freshly fallen snow was determined to be 0.15 g cm^{-3}, on average. With exposure to rain and internal snow melting, the snowpack became denser. The high values registered for day 43 are due to the fact that the snowpack was totally wet: it had rained the night before, and the water that had entered the snowpack was still wet. It appears that – upon ice formation – snowpack densities would decrease again. This would partly be due the normal expansion of water during freezing, but could also be due to the process of crystallization, with the snow crystals forming an open assembly rather than a closed packed arrangement.

In general, the type of forest cover did not seem to affect snowpack density. However, measurement of snow pack densities underneath the softwood cover was biased after day 67: there was much ice crusting in this snowpack. In this case, jars were used to retrieve snow samples only from where the jar could be made to move into the snowpack without shattering. Snow crusting occurred because of extensive water dripping from the softwood canopy. These drips would - for the most part - freeze, as they would start to flow through the cold snowpack. Therefore, actual density of the snowpack in the softwoods became fairly high, and especially so towards the end of the winter.

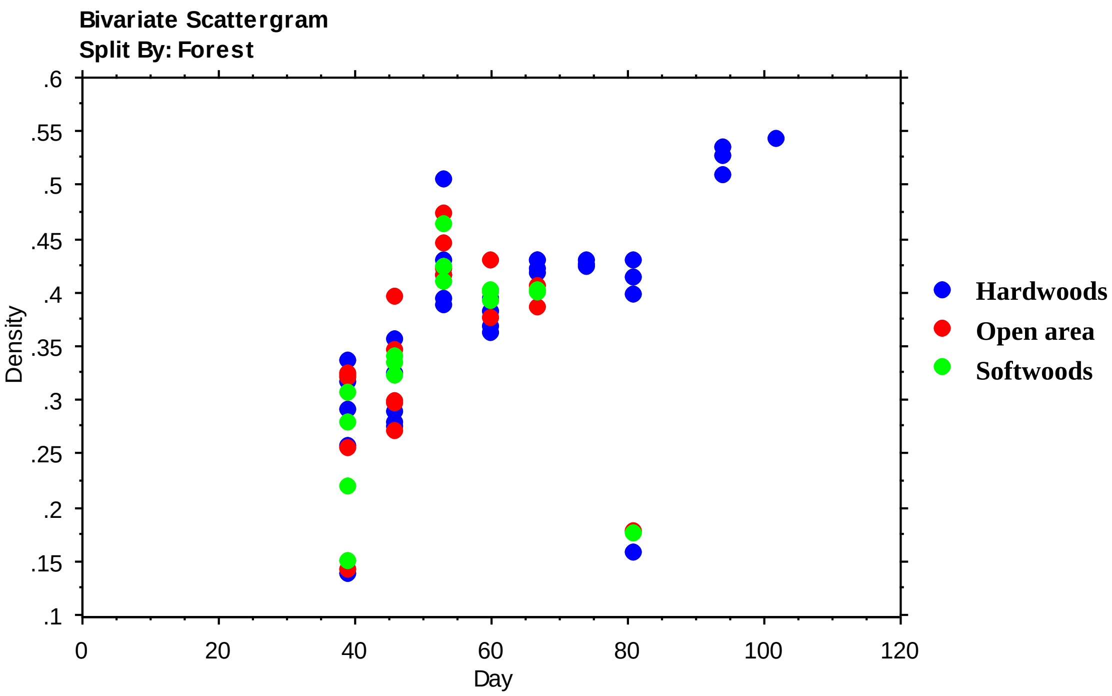


Fig. 15. Density profiles of snowpacks along the winter.

With respect to snowpack and snowmelt pH, we note (Figure 16) that the pH was often lower in the snowmelt than in the snowpack. This means that H^+ were preferentially released from the snowpack, leaving less acidic snow conditions behind. This change occurred gradually between days 39 and 80, when no significant snowfall and only minor rainfall events occurred. Apparently, a little amount of water seeping through the snow would assist in this general elution of snow acidity. A small snowfall on day 81 decreased the average pH slightly, but – afterwards – the pH continued rising again. The snowmelt pH followed the same trend: snowmelt registered increasing values from days 39 to 80, a drop after day 80, and then pH values increased again after that.

Concerning the other ions (Figures 17 to 19), it is difficult to see clear trends. Nitrate seemed to leave the snowpack early. The same trends also may have occurred for SO_4^{2-} and Ca^{2+}. However, the snow within the snowpack was often contaminated with plant and soil debris, as indicated by the high fluctuations of the ion contents within the snowpack. Contaminations occurred in various ways. In the open sites when snow had disappeared on day 80, heavy rain splashed some soil on the plastics so that - when snow fell later and melted - the resulting water was not clean. On the forest sites, but leaves and branches had fallen into the snowpack. Plant debris would deteriorate in time, and would release various ions, including nitrate, sulphate and calcium.

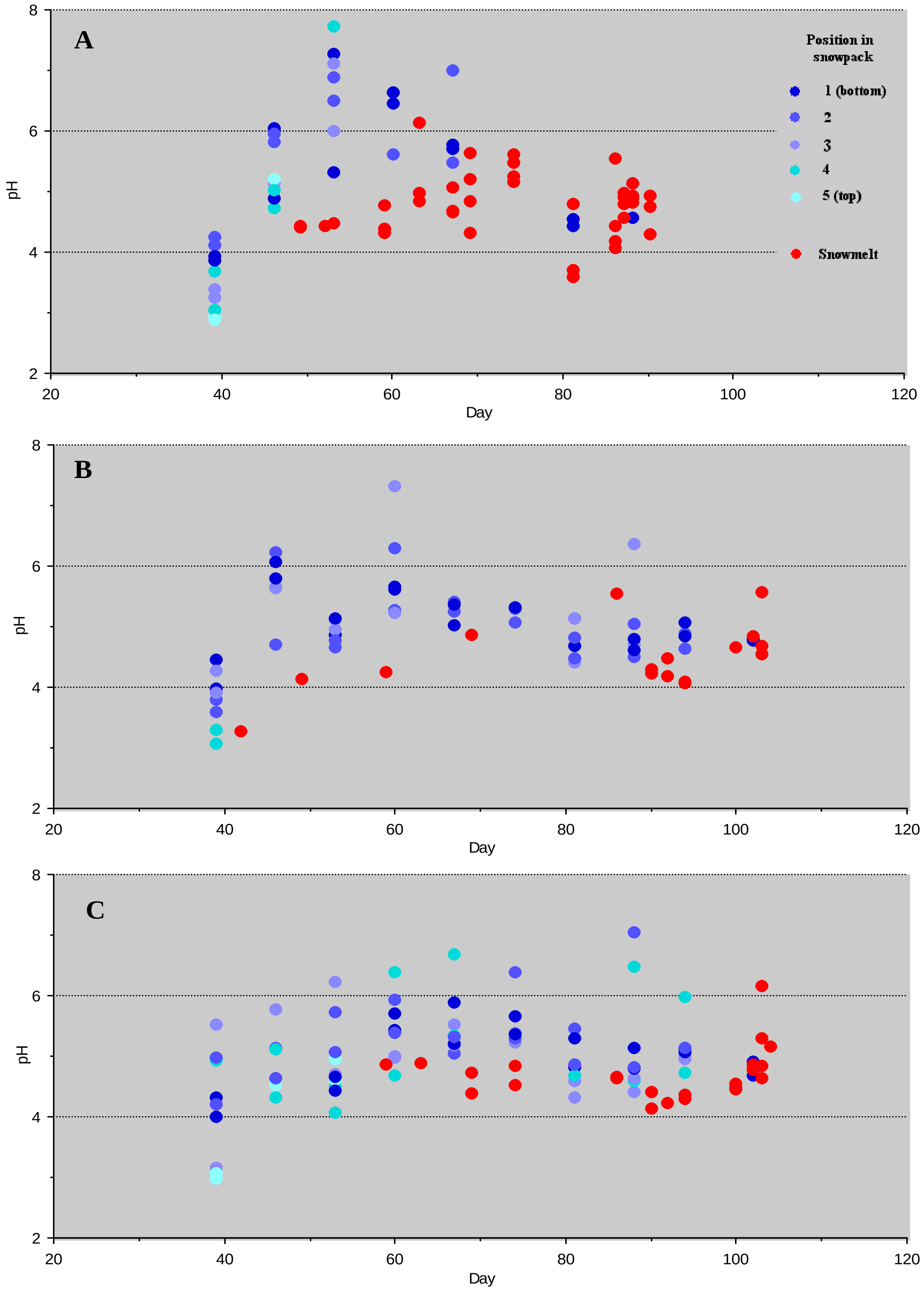


Fig. 16. pH of snowpack and snowmelt measured during winter 2001-2002 in open areas (A), coniferous forest (B) and deciduous forest (C).

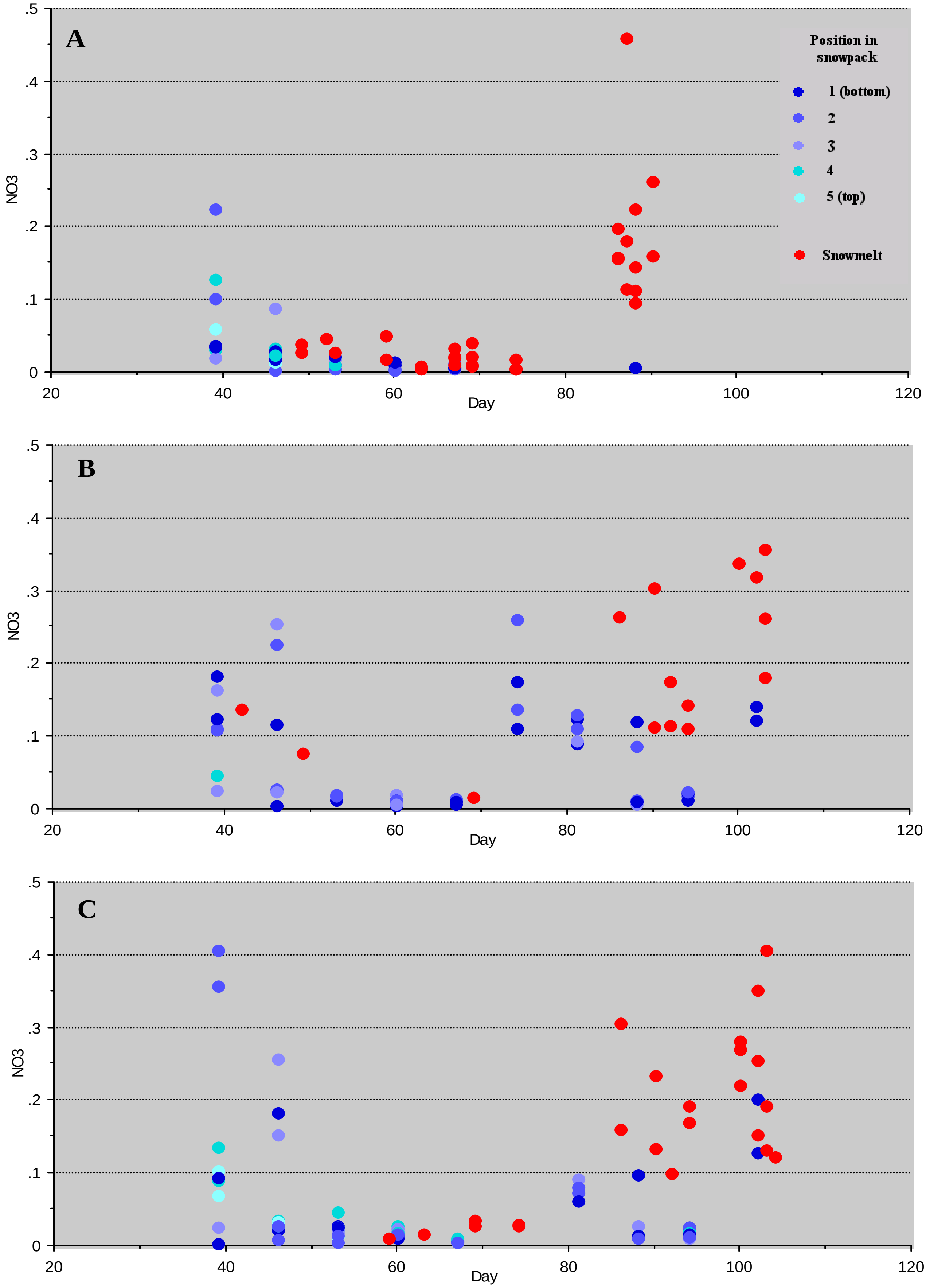


Fig. 17. Nitrate concentrations in snowpack and snowmelt (in ppm) measured during winter 2001-2002 in open areas (A), coniferous forest (B) and deciduous forest (C).

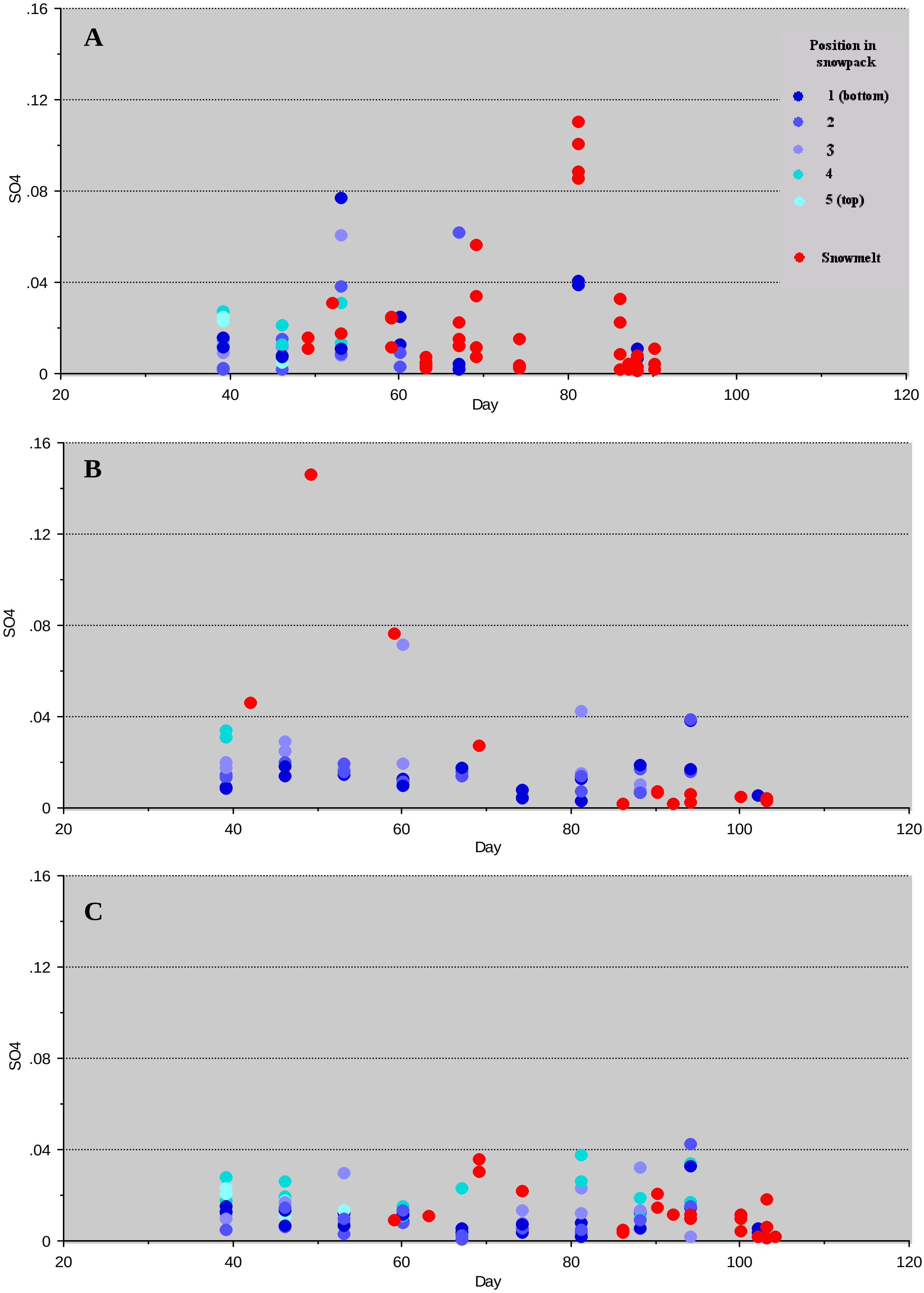


Fig. 18. Sulphate concentrations in snowpack and snowmelt (in ppm) measured during winter 2001-2002 in open areas (A), coniferous forest (B) and deciduous forest (C).

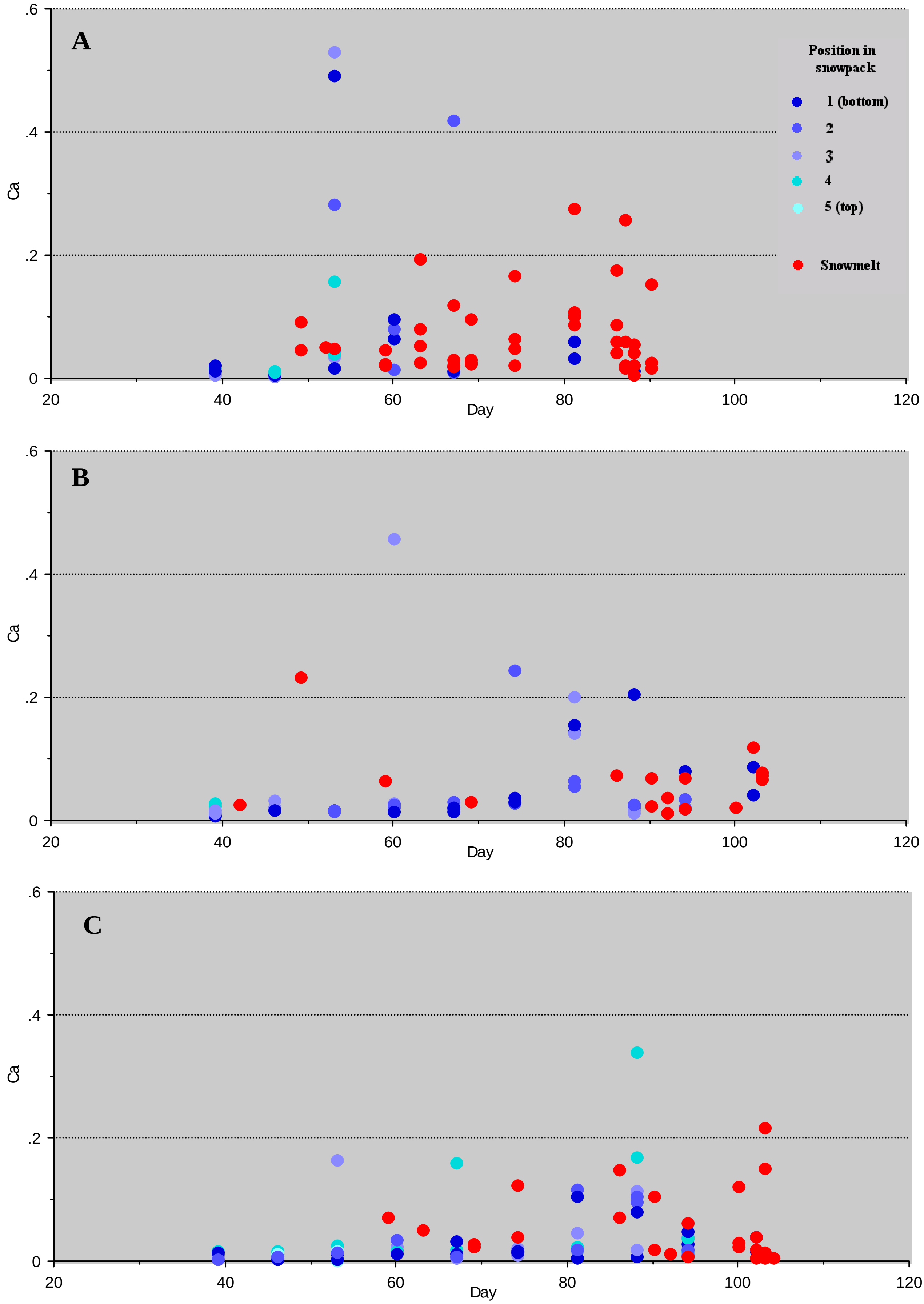


Fig. 19. Calcium concentrations in snowpack and snowmelt (in ppm) measured during winter 2001-2002 in open areas (A), coniferous forest (B) and deciduous forest (C).

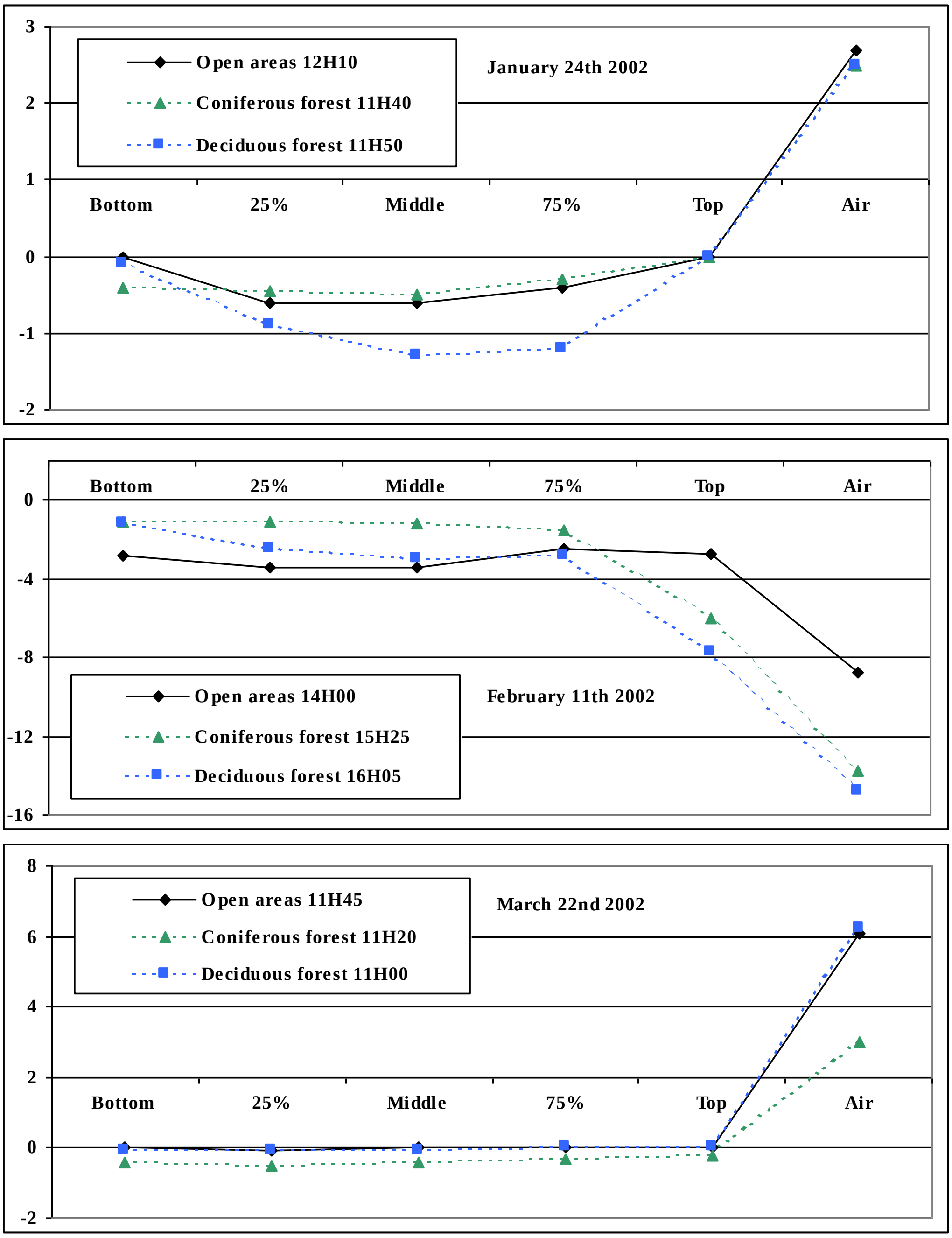


Fig. 20. Snowpack temperatures at various moments in the winter (in °C).

Figure 20 shows snowpack temperature for 3 days, to represent the snowpack temperatures for January, February and March. These profiles reveal several trends.

- The snowpack insulated very well against the fluctuating air temperature. The bottom of the snow pack had temperatures that remained between –4 and 0°C for the entire winter, even though the air temperature varied between +10 and –25°C.
- The snowpack was generally coolest underneath the hardwood stand, but responded more quickly to positive air temperatures than the snowpack underneath the softwood canopy. The open snowpack corresponded to the fluctuating air temperature the fastest.
- The snowpack temperature in the open quickly approached 0°C when air temperatures were positive. This was not the case for the snowpack temperatures underneath the softwood and hardwood canopies.
- When the air temperature stayed above 0°C for a few days, the whole snowpack reached 0°C.
- Whenever the air temperature remained steady for a while (like a day), the temperature gradient within the snowpack became linear.

CHAPTER 5

THE FOREST HYDROLOGY MODEL (FORHYM)

5.1 INTRODUCTION

This Chapter introduces ForHyM as it is used to model all major water and heat flows through forest stands and small forest catchments (Arp and Yin 1992; Yin and Arp 1993). Since this model incurred several changes since its original conception, it was important to critically review this model, and to rebuild this model for the purpose of rebuilding, simplifying and improving many of the semi-empirical formulations that formed the foundations of the model in principle. Some of the improvements that resulted from this rebuilding are described in Chapter 7.

5.2 GENERAL DESCRIPTION

In general, ForHyM simulates water and heat retention and water and heat fluxes through any forest stand, year-round, on a daily basis. Required input refers to daily mean air temperature and daily precipitation (rain or snow). Input also refers to assigning water and heat flow parameters such as soil permeability, specific heat capacity and heat conductivity as these parameters vary with soil moisture, soil texture, OM and coarse fragment content. Similarly, such values need to be assigned to the snowpack as it also varies in time with respect to density. Additional parameters refer to water retention, namely water retention at the point of soil and snow saturation, at field capacity, and at the permanent wilting point. In terms of forest condition, it is important to specify depth of soil layers (forest floor, A, B, C), and type of forest cover (% softwood, and % canopy transparency). Average slope and aspect also need to be specified.

The ForHyM model is realized with the *Stella* modeling software. This software enables the programming of all the processes involved by way of three object types:

- Basic variables (shown as circles): these can take on constant values, or can be defined in terms of other variables and/or constants variables.
- “Flow” variables (shown as circles attached to broad arrows): these can be used to calculate the flow of water or heat; invariably, these flows are associated with stock variables; they are programmed in the same way as the basic circles.
- “Stock” variables (shown as boxes): these variables can be used to accumulate quantities of water or energy over time in any compartment, e.g., any soil layer, the forest canopy, the snowpack; stock variables strictly follow the principle of energy and matter conservation; stock variables need to be initialized.

The values taken by each of these three variable types can be tabulated or plotted on a graph. Tabulation can be done for each time step, or can be done for set time intervals. Calculations proceed by checking values within all circles and boxes, and by upgrading these according to the programmer specifications, one step at a time. *Stella* provides a choice for the calculation: the Euler method, or the Runge-Kutta method. For calculations that are forced by stepwise input, as is the case for daily weather, it is best to use the Euler method, even though this method is less efficient numerically than the Runge Kutta method. The reason for this choice is that Runge Kutta calculations over-smoothes the expected results.

The programming interface between the computer and the user is visual and hierarchical rather than linear: all variables and their connections to one another are screen-displayed. Figure 21 provides an example.

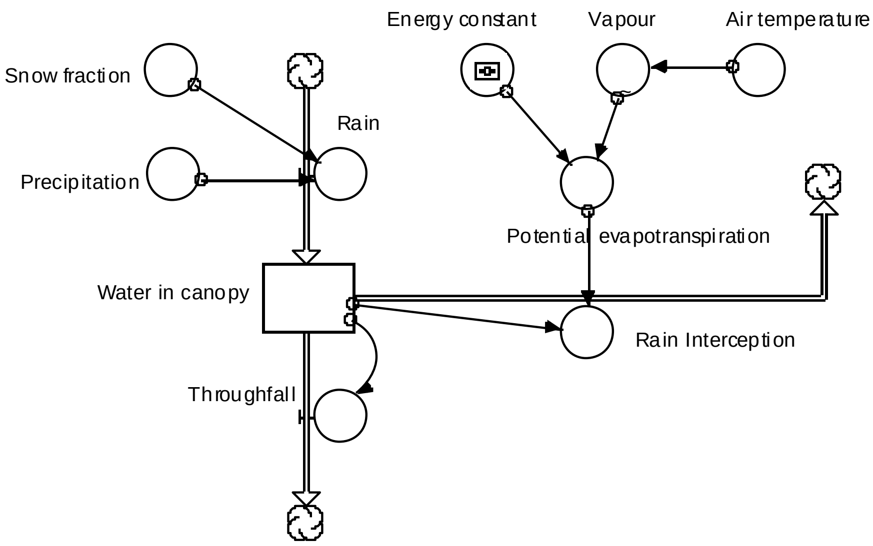


Fig. 21. Sample of *Stella* program display. Shown are basic variables (circles), flow variables (attached to broad arrows) and stock variables (boxes). Thin arrows shows how variables are connected with one another. Cloud like symbols mean output into or input from the “environment”. Circles containing parameters that are used for model calibration are flagged with a small box inside.

Data that are required to run the model (e.g., weather data) are supplied to the model through the same screen display. For this purpose, each circle allows the user to enter a constant, an equation, a graph, or a table. The table function allows the user to copy data into any circle, but only one variable per circle. For cross-referencing purposes, each graph or table within any circle requires the user to declare the range and steps of the x-axis for the copied y-data. Data that are supplied to each *Stella* circle must be evenly spaced, with no missing data.

Prior to running the model, the user must choose time range and the time step “dT”: if dT is one day and daily data are used, then the model will be a daily model. For daily data, dT should be one day or less. A smaller dT gives more precision, but the length of the simulation is inversely proportional to dT. For an annual model, one can also choose dT = 1, but all inputs must either be consistent with annual rates, or the appropriate conversions to that effect must be put in place within the program.

The ForHyM model is basically a one-dimensional model to simulate vertical fluxes. However, it can also be used to simulate water that exit soils and watersheds horizontally. The user should specify all inputs in terms of per hectare quantities. In this way, total water discharge rates can be calculated either in terms of discharge per hectare, or discharge per watershed area. To do the latter, the user would multiply the per hectare discharge values with the hectares of the catchment area. The model recognizes the following forest stand compartments: canopy, snowpack, and forest floor (FF), soil (layers A and B) and subsoil layers. Each of these compartments requires proper and site-specific initialization input. These inputs refer to thickness and texture (sand, silt and clay percentages, coarse fragment percentage and OM content) for the different layers; seasonal maximum for the leaf area of the forest canopy, latitude, altitude, aspect, average slope of the location, percentage of coniferous trees and deciduous trees, and depth of the roots of the trees.

The model calculates

- Daily solar radiation.
- Energy that goes into or out of the surface of the soil or the snowpack (in the winter).
- Temperature at various depths in the soil.
- Snow density, snowdepth and snowmelt amounts.
- Formation and disappearance of ice in the soil.
- Water content of each layer and fluxes of water between layers, runoff, and stream discharge.
- Ion fluxes from snowmelt.
- All essential heat and water flow parameters, based on specifications of soil texture, OM content, and soil depth, and model internal calculations for moisture and ice.

5.3 SPECIFYING SOIL, SNOW, AND CANOPY PARAMETERS

This module calculates, for each layer, and for each time step, the following variables:

- Porosity (by calculating bulk density Db).
- Saturated hydraulic conductivity (Ksat).
- Field capacity (FC).
- Permanent wilting point (PWP).

Each of these variables depends on several soil properties, as follows

Db = f (sand content, OM content, depth)

Pore space = 1-Db/Dp, with Dp the particle density of the soil.

Ksat = f (Db, sand content)

FC = f (sand content, clay content, OM content, Db)

PWP = f (sand content, clay content, OM content, FC)

For more details, see Chapter 6. Figure 22 shows the module in *Stella*.

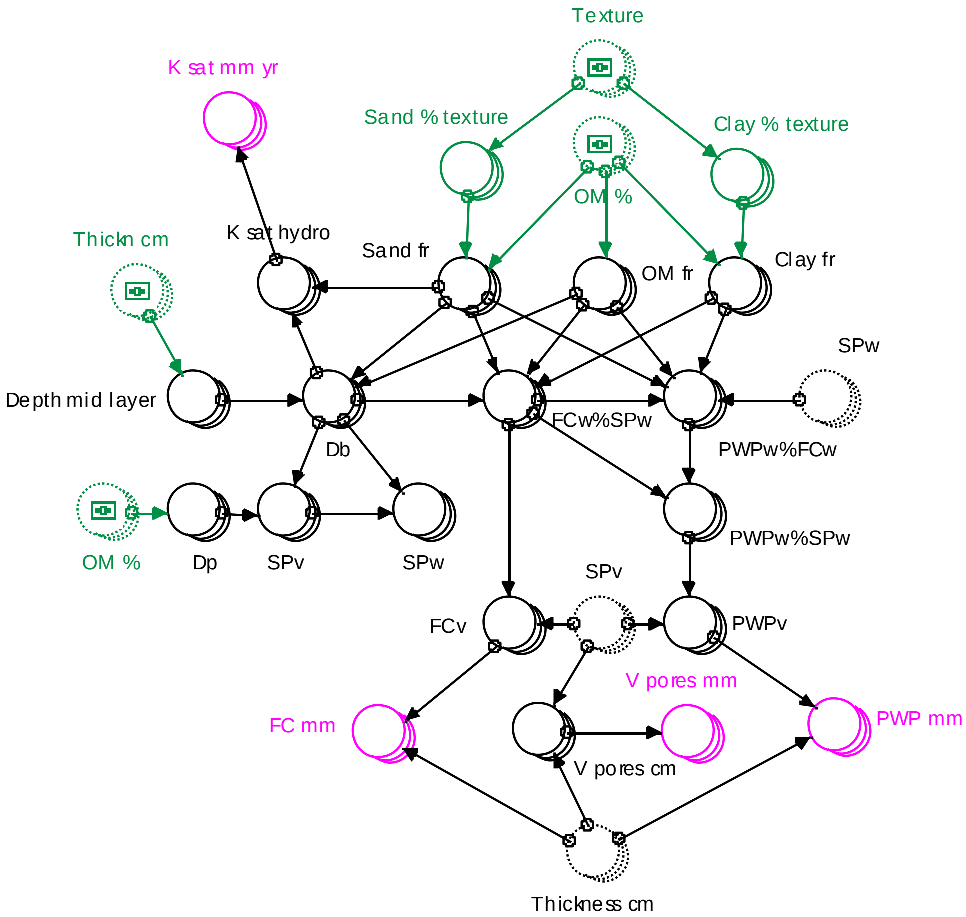


Fig. 22. Soil parameters calculations in the ForHyM model.

The **Leaf area index (LAI)** is the one-sided surface of all leaves in the forest canopy, in m^2/m^2. In ForHyM, this index determines the interception of precipitation and the amount of evapotranspiration by the trees. It is a weighted average of the leaf area of coniferous trees and deciduous trees within the forest canopy. The model uses this index based on maximum leaf area per softwood/hardwood species, anticipated canopy transparency, and anticipated changes with season. For hardwoods, cumulative degree-days in spring are used for time of leaf development; day length in the fall is used to determine time of leaf abscission. Figure 23 illustrates this.

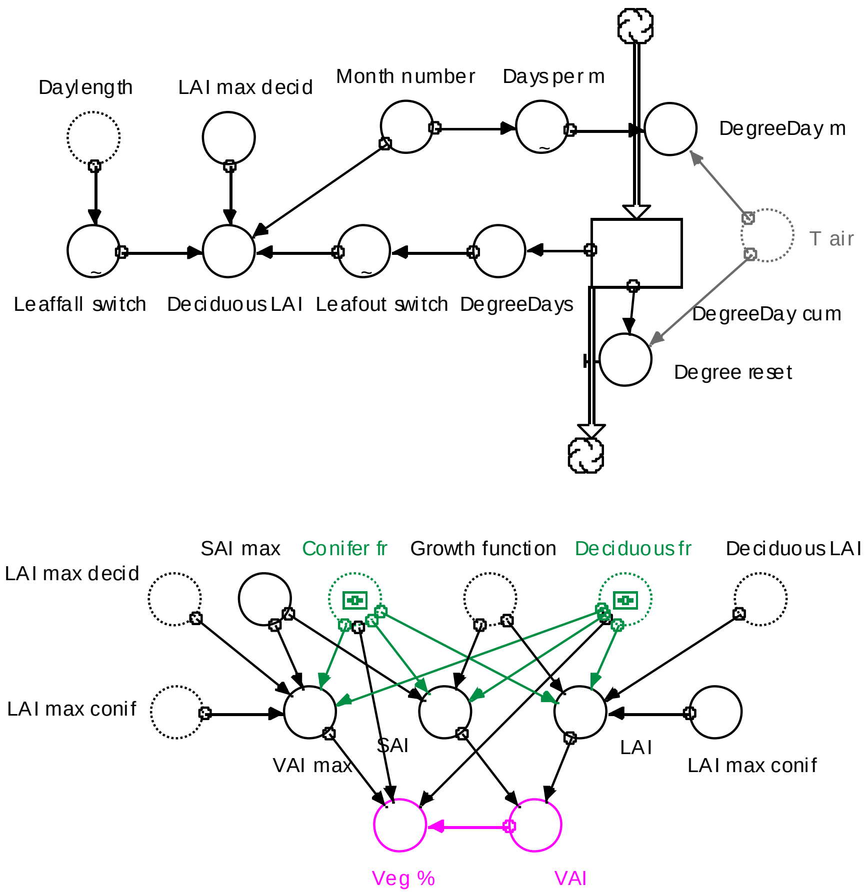


Fig. 23. Vegetation cover calculations in the ForHyM model.

5.4 THE HYDROLOGY MODULE

The hydrology module within ForHyM uses daily precipitation data as input for snow and rain. Interception of precipitation by the trees occurs whenever it rains or snows. The maximum interception capacity is proportional to the leaf area index. What is not intercepted falls to the ground: snow accumulates in the snowpack and rain enters the snow or the soil. If it rains and there is a snowpack, some of the rain is retained in the snowpack. A part or all of this rain may be retained within the snow as ice, or liquid water depending on the snowpack temperature. The maximum amount of liquid water that can be stored in the snowpack is equal to the field capacity of the snowpack. Some of the accumulated snow evaporates back in the atmosphere. Snowmelt water goes to the surface of the soil if it cannot be retained in the remaining snow (Figure 24).

Liquid water that reaches the soil surface either infiltrates the soil or becomes surface runoff. The process of calculating flow through the soil is recursive, because what can infiltrate in one soil layer depends in part of how much water leaves that layer and the ones below, and whether the underlying soil is already saturated. Therefore, the deepest underground flow is calculated first, then the water content of the subsoil layer, then the flows that would go into the subsoil. If there is more water coming in than going out, then part of the water is used to fill empty pore space. The rest is required to flow laterally and becomes “interflow”, or remains in the layer above.

Percolation and interflow rate calculations are based on Darcy's Law. For interflow, average slope at the watershed scale determines the flow gradient. For saturated soils, flow rates are determined using estimated values for the saturated hydraulic conductivity. When the soil water content is between saturation and field capacity, a similar equation is used, but with a soil-moisture dependent adjustment for the hydraulic conductivity. Below field capacity, there is no flow.

Some water is removed from the two A and B soil layers, and from the forest floor through evapotranspiration, based on leaf area index and atmospheric temperature: water is removed from the soil by the roots, and evaporates at the level of the leaves.

The moisture content of each layer (snowpack, forest floor, soil, subsoil, canopy) is determined by adding all flow inputs (positive) and outputs (negative) on top of the moisture content in each layer at the preceding time step (antecedent condition). In the soil, water content never decreases below the permanent wilting point, and soil percolation and runoff flows cease from any soil layer once the moisture content of that layer is drawn to or below the field capacity of that layer.

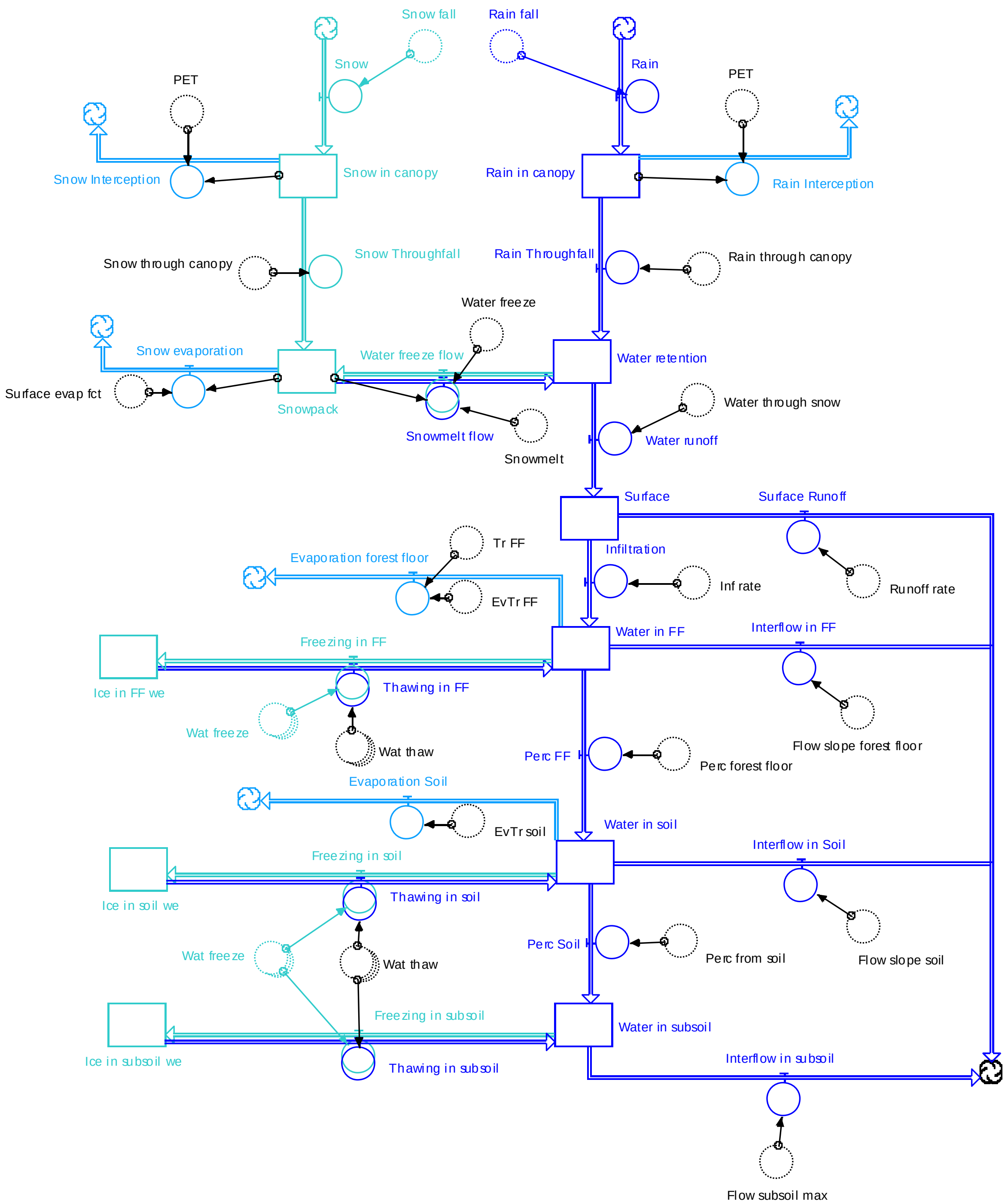


Fig. 24. Hydrology calculations in the ForHyM model.

5.5 THE ENERGY / HEAT FLOW / TEMPERATURE MODULE

The energy, heat flow and temperature calculations are based on the following considerations, at each time step:

- The amount of incoming and outgoing solar radiation, and long wave radiation.
- The albedo of the soil or the snow to calculate the amount of reflected solar radiation
- The temperature at the surface of the soil or snowpack, based on the energy balance of all incoming and outgoing energy flows for this surface (Yin and Arp 1993); this balance includes the albedo factor, solar short-wave and long-wave radiation, sensible heat exchange between the surface and the atmosphere, and heat conduction from or into the ground or snow below.
- The temperature at a depth of more than 12m in the subsoil is assumed to be constant year-round, and is calculated from the mean yearly air temperature and air temperature range, and the mean annual snow depth (Yin and Arp 1993).
- The temperature at surface and the temperature at 12 m depth define the boundary value conditions for the algorithm that is used to calculate the temperatures between these boundaries, at any depth.
- The amount of energy available for snowmelt is the sum of the solar radiation, heat transfer from or to the atmosphere, and heat conduction between the snow and the underlying ground and the soil, and the heat of the rain.
- The amount of energy needed to melt the snowpack is used in part to freeze some or all of the incoming liquid water within the snowpack The energy brought into the snowpack on account of rain is included in the calculations as well, to adjust the snowpack temperature according to the overall energy balance of all incoming and outgoing energy flows.

- The amount of energy that is released by ice formation, and the amount of energy needed to melt or thaw the ice or snow is also considered for each layer (snowpack, forest floor, each soil layer). This is done by tracking all the energy requirements that are involved with the liquid-ice phase transformation in each layer. The model, in addition, tracks moisture content (ice, liquid water) and pore space in each layer, because the amounts of air, water, and ice, together with mineral soil solids and amount of OM in the soil all influence the parameters that control the heat and water flow through the snow and through the soil. Specifically, the model calculates the heat capacities and heat conductivities of each soil layer, and these are used to determine the heat flows between the various layers. Basically, the heat capacity of a given soil is the arithmetically weighted average of the heat capacities of its five main components: minerals, OM, water, ice and air. The weighting factors are the volumetric fractions of each constituent. The conductivities are calculated as a geometrically weighted average of the same five basic components of the soil, with the same weighting factors. For the snow, there are three components: air liquid water, and ice.
- The algorithm used to calculate temperatures in the soil is based on the one-dimensional heat-conduction equation:

$$\rho c \frac{dT}{dt} = \frac{d}{dx}\left(k\left(\frac{dT}{dx}\right)\right)$$

With ρ as volumetric mass density, c as heat capacity and k as thermal conductivity.

- The soil is divided in layers (forest floor, A, B, C, and twelve one-meter-thick layers). When a snowpack is present, another layer is added on top of the soil. The physical and thermal properties of each layer are assumed to be homogenous inside each layer, and temperatures are calculated at the middle of each layer. The finite-element algorithm that is used to solve the above differential equation is the “fully implicit discretization method” (Patankar 1980). The recursive formula for the temperature calculations is given by:

$$a_i\, T_i = b_i\, T_{i+1} + c_i\, T_{i-1} + d_i$$

for layer i = 1 to n, with T_0 and T_{n+1} as known boundary temperatures, and a_i, b_i, c_i and d_i as recursive coefficients that can all be derived from the differential equation above, for each layer and time step. This finite-difference equation is then solved by way of the Tri Diagonal Matrix Algorithm, such that:

$$T_i = p_i\, T_{i+1} + q_i$$

$$p_i = f(p_{i-1})$$

$$q_i = f(p_{i-1}, q_{i-1})$$

The variables are calculated in the following order: p_i, i = 1 to n, q_i, i = 1 to n, T_i, i = n to 1. Figure 25 illustrates this. This algorithm generates a stable solution for each time step.

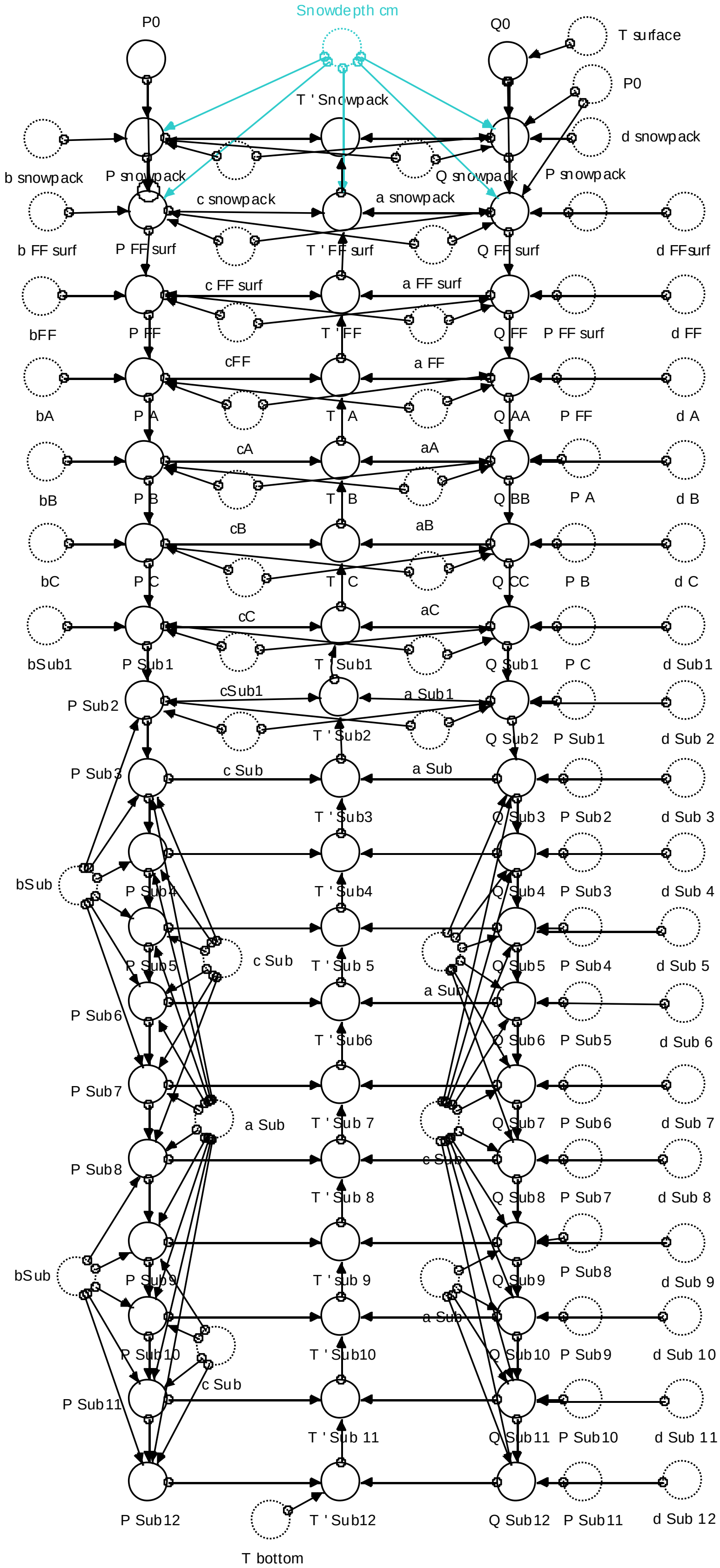


Fig. 25. Temperature calculations in the ForHyM model.

5.6 INTERDEPENDENCIES BETWEEN THE HYDROLOGY AND HEAT FLOW CALCULATIONS

When the temperature of a moist layer becomes negative, that layer's temperature is reset to 0°C until all of the heat that is lost beyond the 0°C point is converted to ice. Once all the moisture is frozen, temperatures are allowed to drop further. The reverse is also true: once cold ice reaches 0° C, ice begins to melt in proportion with the incoming heat.

There is constant flow of information between the hydrology and heat modules of the model: at each dT, the heat module is set to calculate how much energy is available to melt ice or freeze water in each layer; in the hydrology module, the energy released to change water into ice is used to calculate the amount of water that is diverted from percolation and runoff at that time. In turn, the amount of ice and water in each layer is used to re-calculate heat capacities and conductivities in each layer.

5.7 REFERENCES

Arp, P.A., and Yin, X. 1992. Predicting water fluxes through forests from monthly precipitation and mean monthly air temperature records. Canadian Journal of Forest Research. 22: 864-877.

Patankar, S.V. 1980. Numerical heat transfer and fluid flow. Hemisphere publishing corporation, McGraw-Hill book company. pp. 1-59.

Yin, X., and Arp, P.A. 1993. Predicting forest soil temperatures from monthly air temperature and precipitation records. Canadian Journal of Forest Research. 23: 2521-2536.

CHAPTER 6

SOIL PARAMETERS PREDICTION FROM SIMPLE MEASURES OF SOIL TEXTURE, ORGANIC MATTER AND DEPTH

6.1 INTRODUCTION

This Chapter lays the foundations for calculating soil hydraulic permeability, soil density and pore space, field capacity and permanent wilting point from soil specification about texture, OM content and soil depth. For example, high levels of soil OM are usually associated with large pore space, and therefore large soil permeabilities. Reductions in OM, in contrast, lead to denser soil. Sandy soils, as opposed to loamy soils, tend to have low field capacities for water retention, and the differences between water content at field capacity and permanent wilting point are also quite low in sandy soils. In most hydrological models, interdependencies among the various soil properties that determine water flow are generally ignored. Nevertheless, it is important to systematically evaluate these interdependencies, to ensure that the resulting model calculations are as realistic and as portable as possible, particularly within the context of environmental change, with change in soil OM being one of the most important variables to track over time. Building these equations is essential for expanding existing hydrological models (or developing new models) to simulate soil water flow and soil water retention across soil substrates of varying substrate conditions. Typically, the specification of soil parameters such as permeability, pore space, field capacity and permanent wilting point are essential for the initialization of such models. The specific objective of this Chapter, therefore, is to develop equations that reliably capture the general interdependency between the variables cited above.

6.2 OUTLINE OF METHODS

Data about soil hydraulic permeability, soil density and pore space, field capacity, permanent wilting point, soil texture, OM content and depth of soil were obtained from select soil survey reports that provide this information. A sample of survey reports that contained all the required data and metadata was located among the soil survey reports that were done for New Brunswick and Nova Scotia (Table 2). In these reports, most of the surveyed soils represent pristine forest conditions under a temperate to boreal maritime climate, and in a glaciated landscape.

Multiple linear and non-linear regression analyses were used to establish semi-empirical relationships between the soil hydrological properties (dependent variables) and the soil physical properties (independent variables). Non-linear regression analyses were used to ensure that the final equations conformed mathematically with the entire feasibility range of any of the contemplated soil properties, ranging from open space (as in an organic soil that is very fluffy) to a solid (as a soil that has been consolidated into a rock-like structure, with no pore space). Also, within a dry soil, if the soil OM represents, e.g., 20 % of the soil weight, then the soil mineral portion of the soil weight can only amount to 80%. For soils that would be moist, moisture would be allocated to stay within the available pore space: shrinking and swelling is not considered. This would be reasonable in the soils of Atlantic Canada, where the contents with soil minerals that could swell or shrink (such as montmorillonite) is generally very low or absent.

6.3 DEFINITIONS

Soil is a porous material, and - in the natural environment - the pores of the soil often contain some water. Figure 26 shows a schematic description of this.

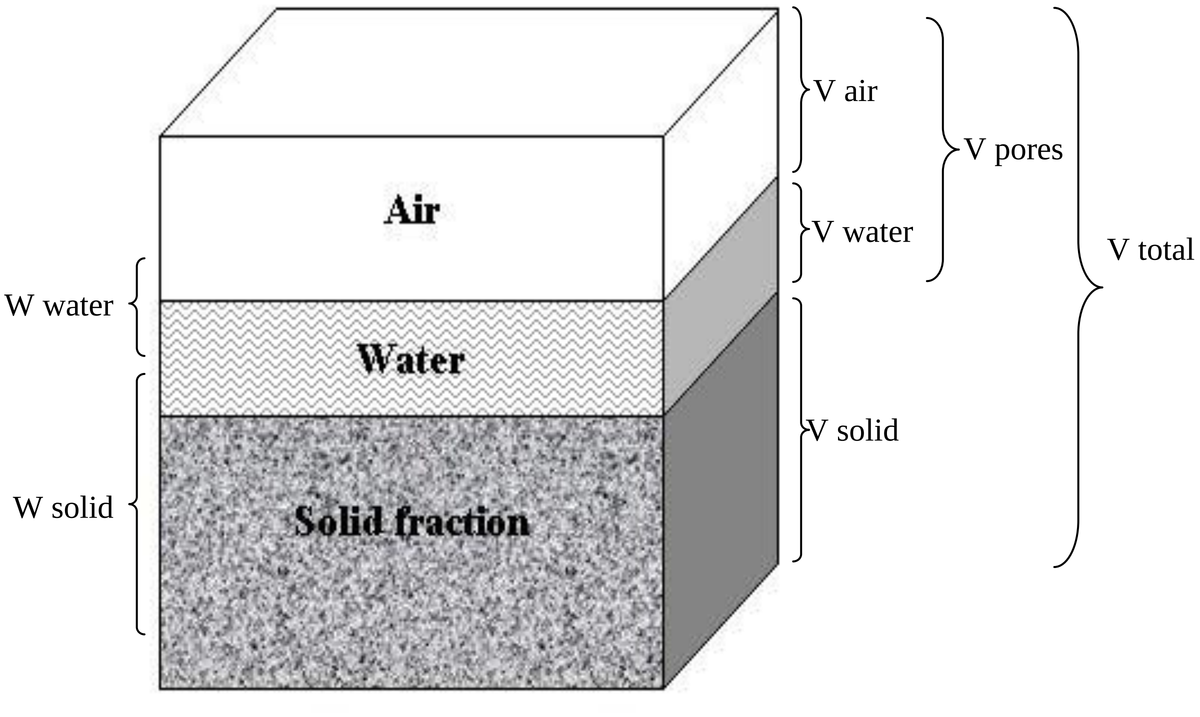


Fig. 26. Soil description.

W solid: weight of the dry soil
W water: weight of water in the soil at a given moment
V solid: volume of the solid fraction of the soil
V water: volume of water present in the soil
V air: volume of air present in the soil

Two commonly used soil parameters are particle density **(Dp)**, which is the density of the solid fraction of the soil, and bulk density **(Db),** which is the overall density of the dry soil.

$$Dp = \frac{W_{solid}}{V_{solid}}$$

$$Db = \frac{W_{solid}}{V_{total}}$$

$$Db = \frac{M_{solid}}{V_{total}}$$

Bulk density, particle density and pore space are related to one another by the following equation (See appendix 1 for more details):

$$\frac{V_{pores}}{V_{total}} = 1 - \frac{Db}{Dp}$$

Three other useful soil parameters are:

- **Saturation point by weight (SPw):** the soil is said to at saturation point when all the pores are saturated with water. Saturation point by weight is defined as the ratio of the weight of water present in the soil at saturation over the weight of the dry soil (for a given volume of soil).
- **Field capacity by weight (FCw):** after water has drained from soil due to gravity, the larger pores are drained, but the smaller pores remain filled with water; at this point, the soil is said to be at field capacity. Field capacity by weight is the ratio of the weight of water present in the soil at field capacity over the weight of the dry soil.
- **Permanent wilting point by weight (PWPw):** plants remove water from the soil until they begin to wilt, but even at the wilting point there are still substantial amounts of water in some soils (especially clayey soils). However this water is held too tightly to permit absorption by plant roots. Permanent wilting point by weight is the ratio of the weight of water present in the soil at permanent wilting point over the weight of the dry soil.

Based on determinations with soil pressure plates, it is possible to relate FC or PWP to specific soil moisture contents, and to specific pressure values that closely simulate conditions of field capacity and permanent wilting point in soils, as shown in Table 1:

Table 1. Definitions of saturation point, field capacity and permanent wilting point

Status of the soil:	**Saturation Point**	**Field Capacity**	**Permanent Wilting Point**
Definition	When all pores of the soil are filled with water	When only small pores of the soil contain water, after water has drained from soil due to gravity	When plants have removed all the water they could from the soil
By volume	$SP_v = \frac{V_{water}^{SatPt}}{V_{total}}$	$FC_v = \frac{V_{water}^{FC}}{V_{total}}$	$PWP_v = \frac{V_{water}^{PWPt}}{V_{total}}$
By weight	$SP_w = \frac{W_{water}^{SatPt}}{W_{solid}}$	$FC_w = \frac{W_{water}^{FC}}{W_{solid}}$	$PWP_w = \frac{W_{water}^{PWPt}}{W_{solid}}$
Pressure applied on soil for measurement	**0 bar**	**0.33 bar**	**15 bar**

Another important soil property essential for soil hydrological modeling is the soil saturated hydraulic conductivity or Ksat, which can be used to calculate the quantity of water that flows per unit of time (Q) through a column of saturated soil under a certain pressure gradient, according to Darcy's law (Brady and Weil 1999):

$$Q = \frac{K_{sat} * \Delta P * A}{L}$$

Where

K_{sat} is the saturated hydraulic conductivity, a property of the particular soil
ΔP is the hydrostatic pressure difference from the top to the bottom of the column
A is the cross-sectional area of the column through which the water flows
L is the length of the column

6.4 REPORTS, METADATA, AND DATA

The soil surveys reports that were used to compile texture (sand, silt and clay content), OM content, coarse fragment, soil saturation point, field capacity, permanent wilting point and saturated hydraulic conductivity are listed in Table 2. The leading numbers in this table are used extensively in the following analysis of the metadata in these references.

Table 2. Data sources for statistical regressions

Number	Area	Province	Year	Authors
1	Northern Victoria County	NB	1976	Langmaid *et al.*
2	Chipman-Minto-Harcourt	NB	1992	Rees e*t al.*
3	Madawaska County	NB	1993	Mellerowicz *et al.*
4	Carleton County	NB	1989	Fahmy and Rees
5	Colchester County	NS	1991	Webb *et al.*

- Units: The units of the water contents data varied among these references. In some references, the water amounts at saturation, field capacity and permanent wilting point were by “ %volume” (3,5), in others these amounts were by “ % weight” (1,2), and in one, this important qualifier was not mentioned (4).

- Methods of sampling: in some references, it was clearly stated that some of the soil samples had been disturbed as part of the sampling protocol for determining FC, PWP, saturated permeability (1), or were left undisturbed for the same purpose (2). Some others did not describe what procedure was followed between field sampling and final analysis (3,4,5). However, knowing if a soil was disturbed or not is essential to accurately know whether the quoted value for the soil parameter refers, e.g., to bulk density under field conditions, or bulk density under laboratory conditions. For this reason, some of the data were either used or not, depending on whether decisions regarding the true nature of the final values could be ascertained, based on the data values themselves and related consistencies or inconsistencies with other variables for the same soil. The final choices of which data were used for which purposes are listed in Table 3:

Table 3. Data chosen for each soil characteristic

	Data source used	Total number of data
Permeability	2,3,4,5	112
Bulk density	1(*), 2	237
Field capacity	1(*), 2	227
Permanent wilting point	1(*), 2	209

*: For (1), it was decided to use only some of the data, because some of the soil samples had been disturbed before water contents measurements were done. A comparison was done between Db from the data and Db calculated from

SPw = (1-Db/Dp)/Db; SPw and Dp were also among the data, but SPw was not measured in the field, so the sample may have been disturbed. Only those data for which there was less than 10% difference between the recorded and calculated field bulk densities were used.

To facilitate the overall modeling purpose, and to ensure that PWP estimates would always be below FC estimates, it was decided to express field capacity **as a percentage of total pore space**, and permanent wilting point **as a percentage of field capacity**. In this way, the theoretical limits for FCw/SPw and PWPw/FCw are clear: each of these two ratios has to vary between 0 and 1 under any condition. The order that was followed to determine the interdependencies of variables was the following:

1. Determine bulk density, as a function of variables such as sand, silt, clay, OM content, and soil depth. In turn, pore space can be calculated from:

$$\frac{V_{pores}}{V_{total}} = SPv = SPw * Db = 1 - \frac{Db}{Dp}$$

2. Determine hydraulic conductivity and FCw/SPw, as functions of variables such as sand, silt, clay, OM, depth, and bulk density.
3. Determine PWPw/FCw, as a function of variables such as sand, silt, clay, OM, depth, bulk density, and FC.

6.5 MULTIPLE LINEAR AND NON-LINEAR REGRESSIONS

Multiple linear regressions were used to determine which of the above variables would affect the dependent variables the most. However, the resulting multiple regression equations of the form:

$$\text{Soil parameter} = a + b * \text{Variable1} + c * \text{Variable2} + d * \text{Variable3} \ldots$$

were not used even though some of the resulting equations conformed very well with the data. The reason for this is the non-conformance of such equations with reality under all circumstances. For example, the linear equation obtained for bulk density was:

$$Db = 1.06 + 0.009 * depth - 1.304 * OM \qquad R^2 = 0.733$$

With depth in cm, and OM content between 0 and 1 (1 being 100%). Note, however, that when depth = 0 and OM =1 for example then Db = - 0.135 g cm^{-3}. Instead, Db should always be greater than 0 (pure air) and also smaller than 2.6 (rock).

The search for a good non-linear model that would conform with the above expectations yielded the following equation:

$$Db = \frac{1.27 + 1.33 * (1 - \exp^{(-0.0066 * depth)})}{1 + 7.18 * OM} \qquad R^2 = 0.826$$

As shown by the R^2 value: this equation not only keeps Db between 0 and 2.6 g cm^{-3}, but also provides a better statistical fit, while using the same number of statistical parameters (3).

6.6 RESULTS

Table 4 shows which were the most relevant independent variables (p-value < 0.001) for each dependent soil variable after performing linear multiple regression analysis, and how these affect each of the listed parameters (up or down).

Table 4. Comparison of the variables determining soil properties

Variable \ Soil parameter	**Db**	**FCw/SPw**	**PWPw/FCw**	**Ksat**
Sand content		↓		↑
Silt content				
Clay content			↑	
OM content	↓		↑	
Depth	↑			
Db		↑		↓
FCw			↓	

These associations can be understood as follows:

- The bulk density of the soil increases with depth because of increased soil compaction with increased soil depth. Bulk density, as to be expected, diminishes with increased OM content.
- The field capacity decreases with sand content, because sand provides more space for large pores than for small pores. Subsequently, FC should increase with bulk density, because a higher density means a systematic loss of large pores in favor of small pores, and thus more effective water retention.
- The permanent wilting point increases with clay and OM content because clay and OM have large surface areas and small pores, which exert considerable forces of adhesion on the remaining water within the soil.

- Saturated hydraulic conductivity increases with sand content because sand has large pores; with increased soil bulk density, permeability decreases, as to be expected.

The final non-linear regressions are given below by listing the final equations first, and then listing the values of the various coefficients that are embedded in these equations (Table 5). Often, variables that loaded highly on the linear multiple regression equations also appeared in the non-linear equations, but this was not always the case. Also the non-linear regressions generally increased the goodness of fit (as indicated by R^2) and there improved the quality of the predictions that can be generated from these equations as well.

Table 6 shows the correlations between the various coefficient estimates. An attempt was made to formulate the final non-linear regression models by making sure that the remaining correlations among the final set of regression coefficients would be as low as possible. In some cases, success was easily achieved, in other cases, significant correlations among the coefficients persisted. Here, further simplifications can in theory be made by establishing linear relationships among the highly correlated coefficients within a particular equation. The results obtained from doing so, however, were not as attractive as the final choices that are listed here.

Shown in Figure 27 are scatter plots of data versus best-fitted values, for equations (1) to (4). Figures 28 to 31 show the variations of the dependent soil parameters by varying the independent soil variables to illustrate that equations (1) to (4) do conform with the general feasibility range of the each of the four dependent variables of this Chapter.

$$\mathrm{Db} = \frac{\mathrm{a} + (2.6 - \mathrm{a} - \mathrm{b} * \mathrm{sand}) * (1 - \exp^{(-\mathrm{c} * \mathrm{depth})})}{1 + \mathrm{d} * \mathrm{OM}} \quad \textbf{(1)}$$

$$\log(\mathrm{K}) = \mathrm{a} + \mathrm{b} * \log(2.6 - \mathrm{Db}) + \mathrm{c} * \mathrm{sand} \quad \textbf{(2)}$$

$$\frac{\mathrm{FCw}}{\mathrm{SPw}} = \left(1 - \exp^{(-\mathrm{a} * (1 - \mathrm{sand}) - \mathrm{b} * \mathrm{clay})}\right) * \left(\mathrm{OM} + \frac{1 - \mathrm{OM}}{1 + \mathrm{c} * \exp^{(-\mathrm{d} * \mathrm{Db})}}\right) \quad \textbf{(3)}$$

$$\frac{\mathrm{PWPw}}{\mathrm{FCw}} = \frac{1}{1 + \dfrac{\mathrm{FCw}^{\mathrm{d}}}{\left(1 - \exp^{(-\mathrm{a} * \mathrm{clay} - \mathrm{b} * \mathrm{sand} - \mathrm{c} * \mathrm{OM})}\right)^{\mathrm{e}}}} \quad \textbf{(4)}$$

Table 5. Regression coefficients for Equations (1) to (4)

	a	**b**	**c**	**d**	**e**
Db	**1.23** +/- 0.05	**0.7** +/- 0.1	**0.011** +/- 0.001	**7.2** **+/-** 0.9	
Log (K)	**-0.9** +/- 0.1	**7.2** **+/-** 0.4	**2.0** +/- 0.2		
FC/SP	**0.3** +/- 0.1	**8.8** +/- 0.9	**8** +/- 2	**2.45** **+/-** 0.2	
PWP/FC	**1.1** +/- 0.2	**0.08** +/- 0.05	**4.1** +/- 2	**1.9** **+/-** 0.2	**2.6** **+/-** 0.4

Table 6. Correlation coefficients between regression coefficients for Equations (1) to (4)

Db	**a**	**b**	**c**	**d**
Aa	1			
Bb	-0.0014	1		
Cc	-0.43	0.75	1	
Dd	0.76	0.10	-0.096	1

Log(K)	**a**	**b**	**c**		
Aa	1				
Bb	-0.0062	1			
Cc	-0.90	-0.14	1		

FC	**a**	**b**	**c**	**d**
Aa	1			
Bb	-0.64	1		
Cc	0.24	-0.40	1	
Dd	0.30	-0.63	0.93	1

PWP	**a**	**b**	**c**	**d**	**e**
Aa	1				
Bb	0.96	1			
Cc	0.82	0.85	1		
Dd	-0.58	-0.54	-0.36	1	
Ee	0.86	0.88	0.84	-0.12	1

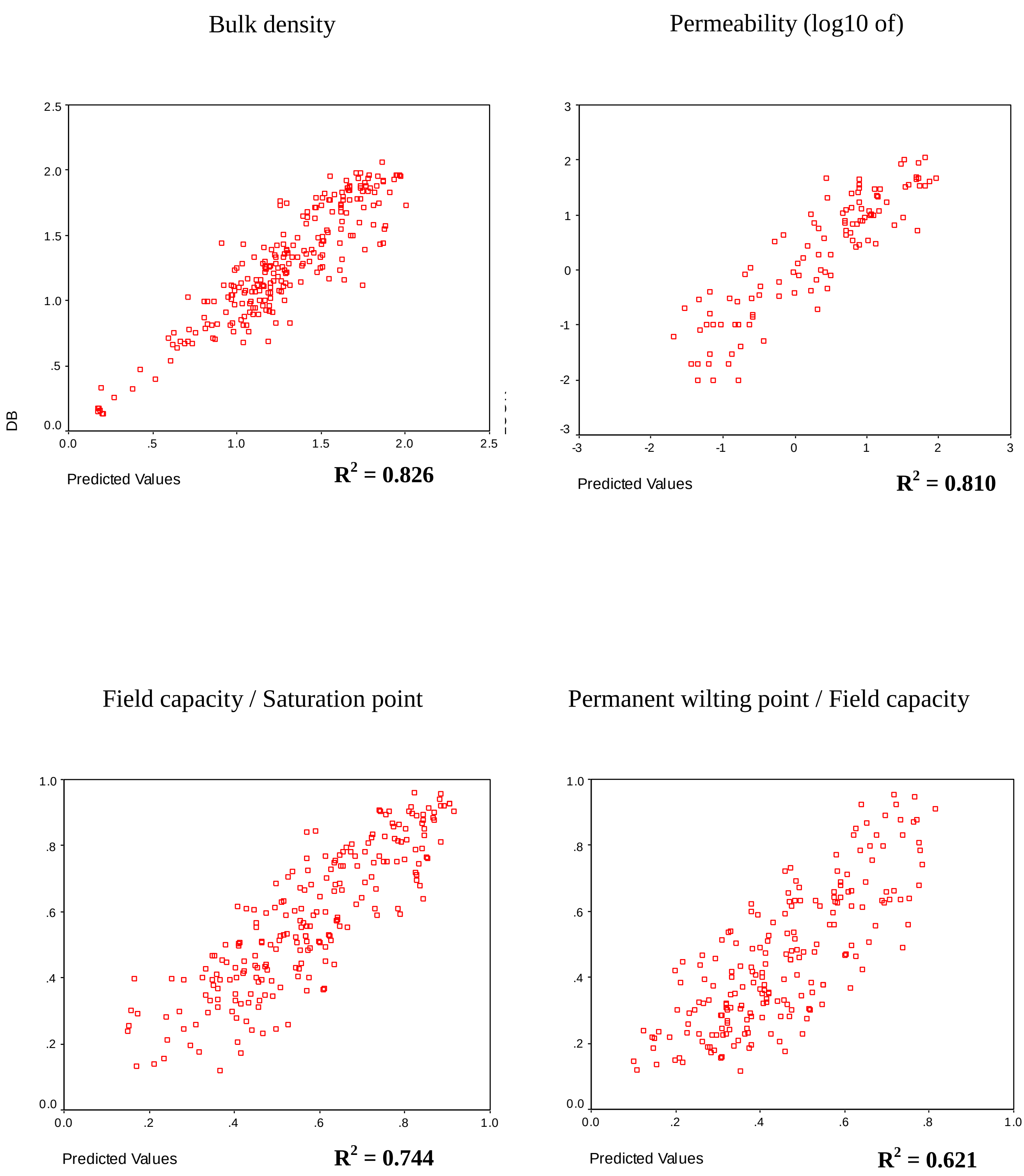


Fig. 27. Predicted values versus data for soil characteristics, for equations (1) to (4).

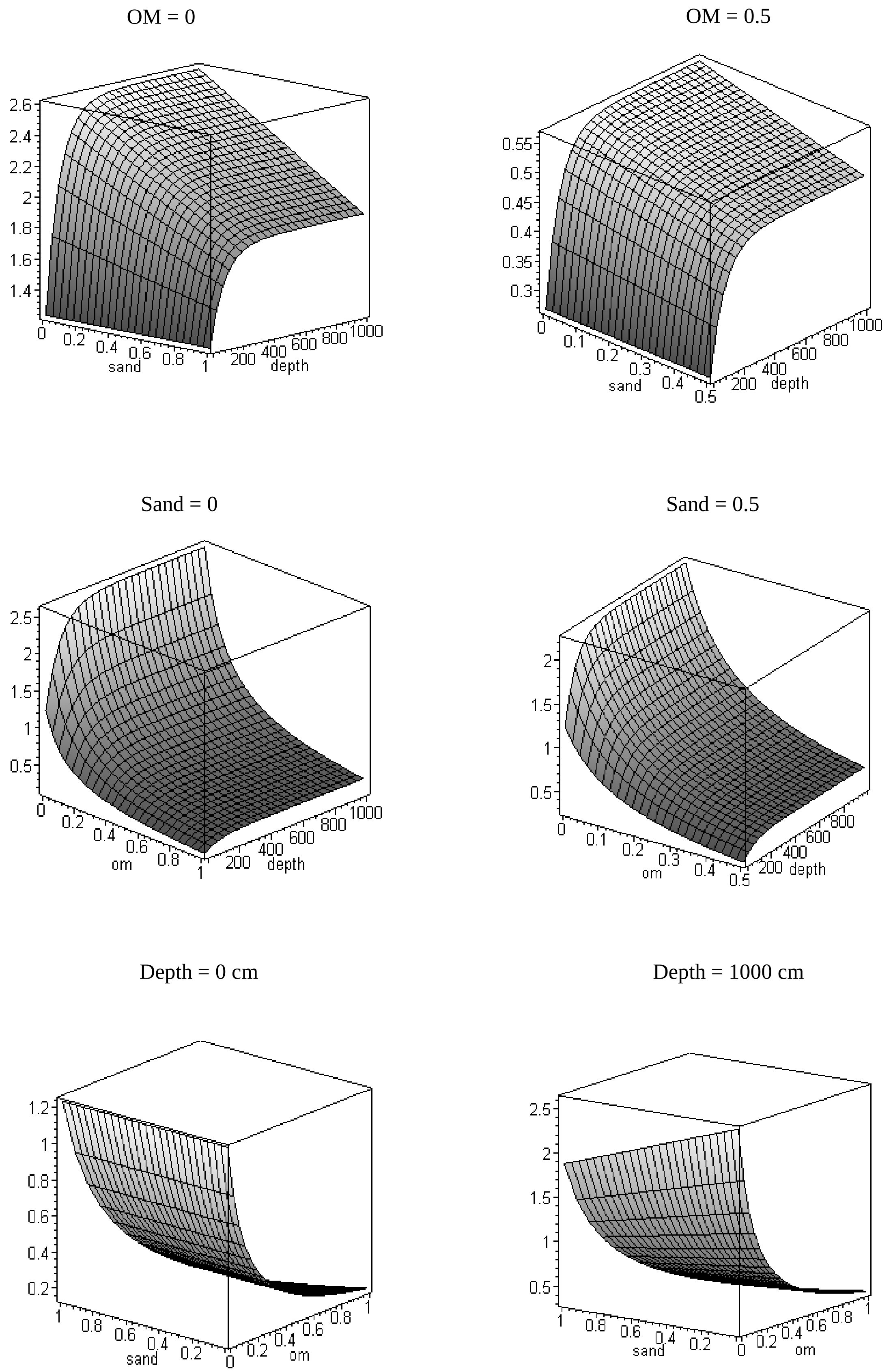
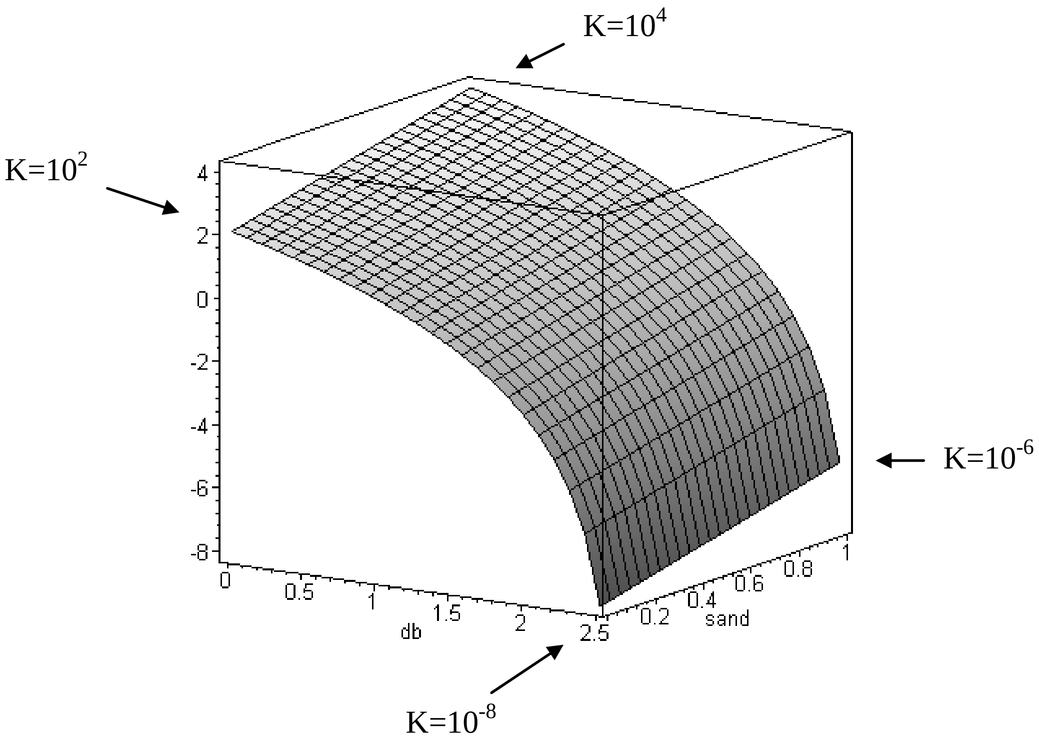
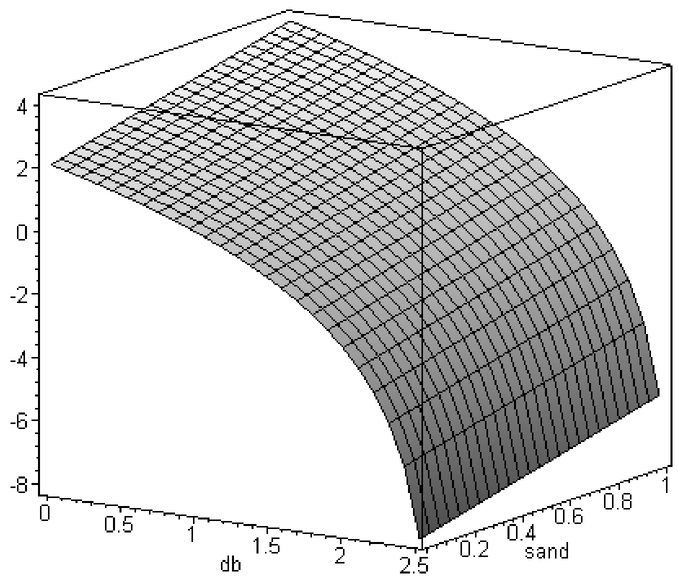


Fig. 28. Bulk density variations (in g cm^{-3}).


K=10^4
K=10^2
K=10^-6
K=10^-8
4
2
0
-2
-4
-6
-8
0
0.5
1
1.5
2
2.5
db
0.2
0.4
0.6
0.8
1
sand


4
2
0
-2
-4
-6
-8
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1
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2
2.5
db
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0.6
0.8
1
sand

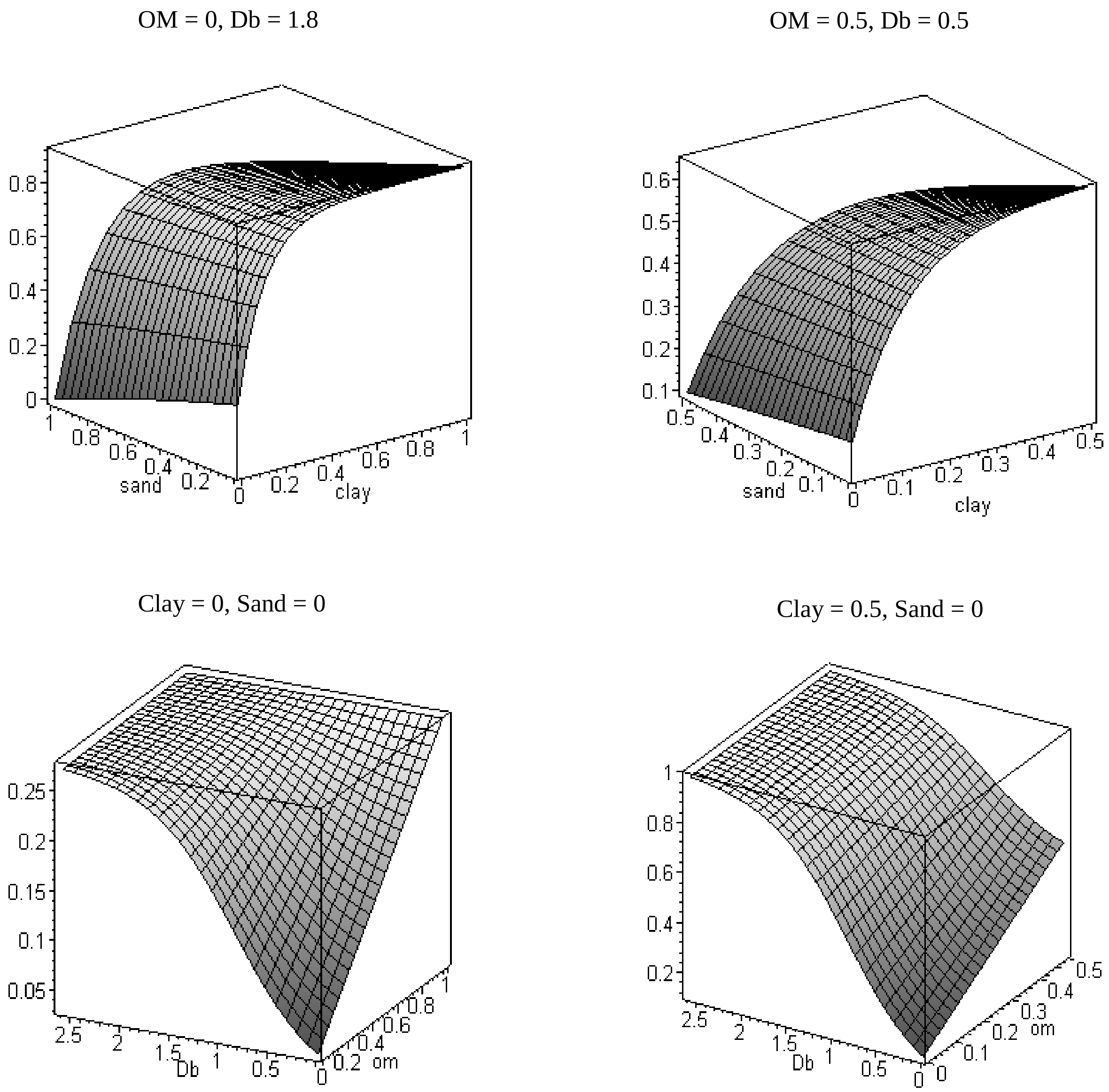


Fig. 30. Field capacity variations (FCw/SPw).

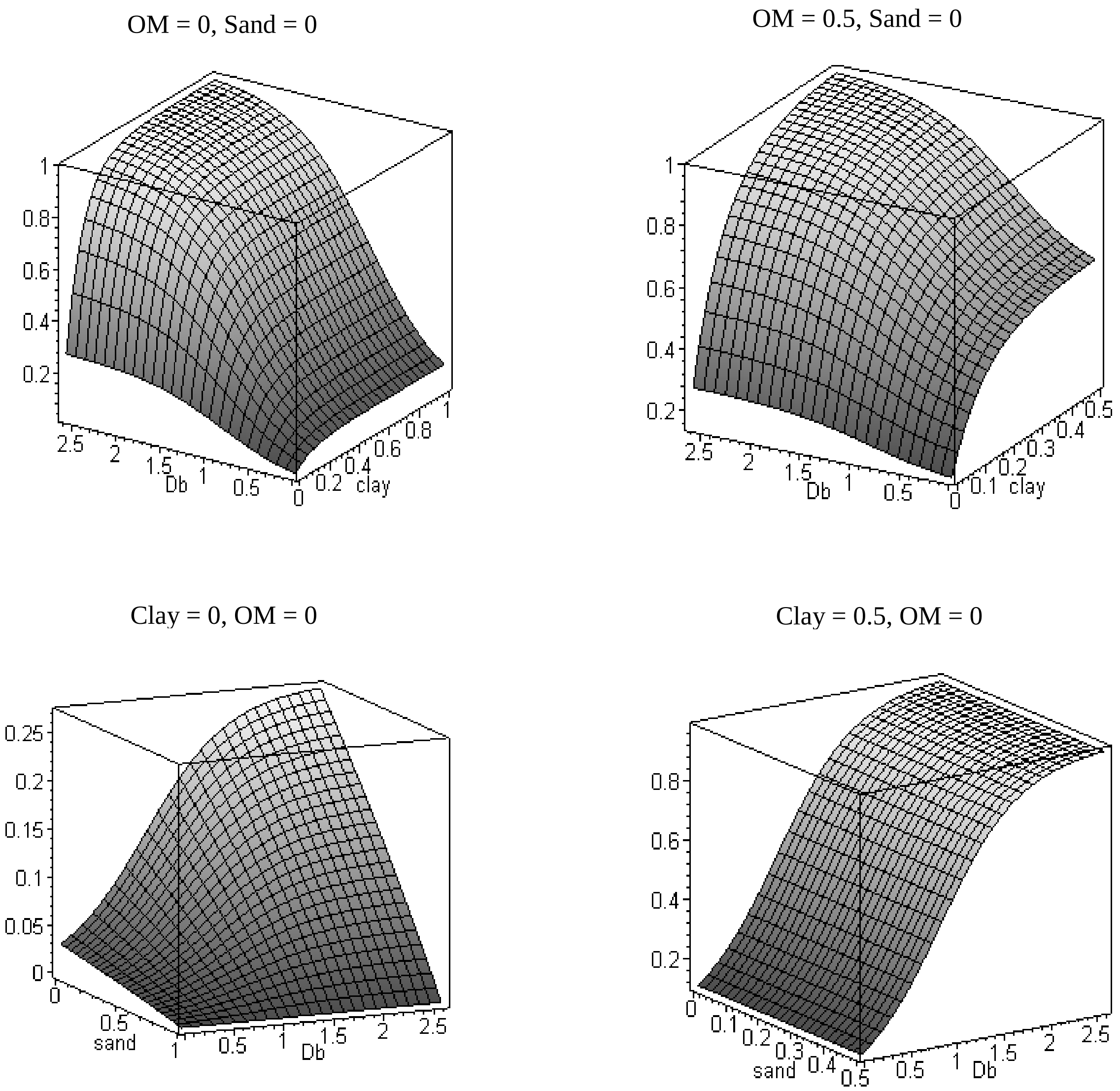
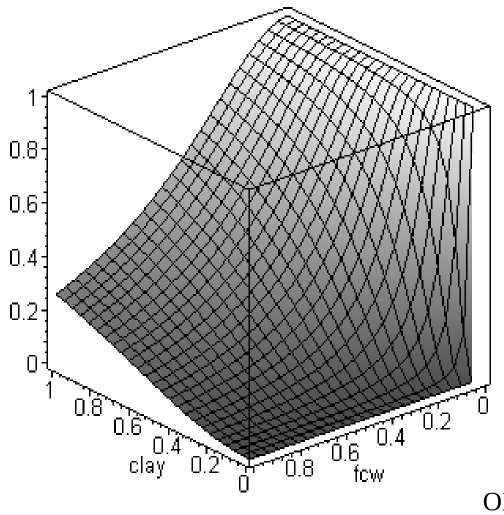
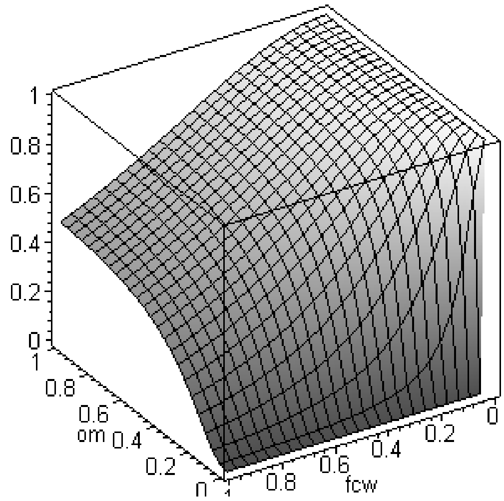


Fig. 30. Field capacity variations (FCw/SPw) (continued).

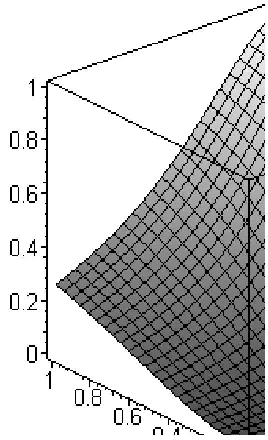
OM = 0, Sand = 0

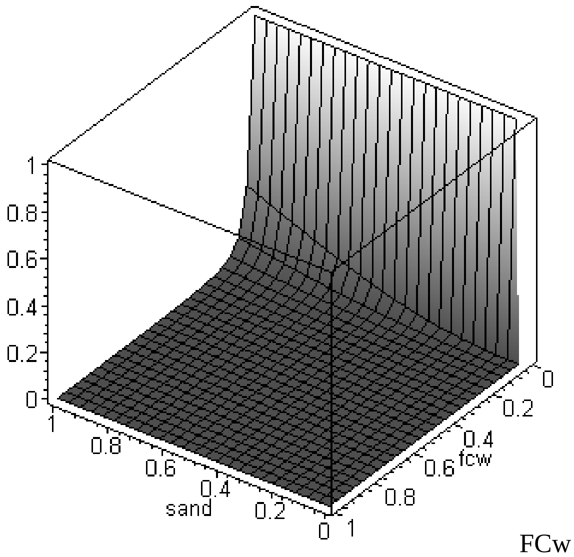


Clay = 0, Sand = 0

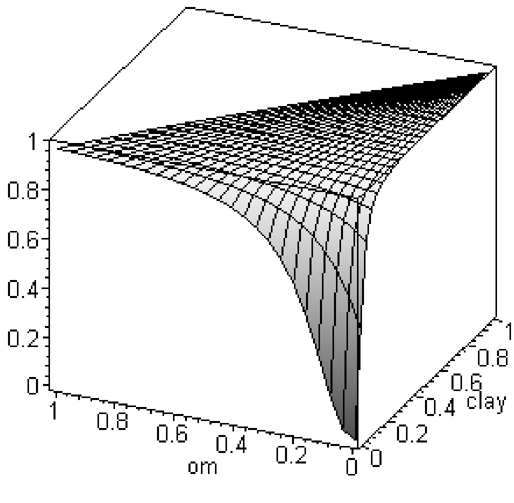


OM = 0, Clay = 0





FCw = 0.1, Sand = 0



FCw = 2, Sand = 0

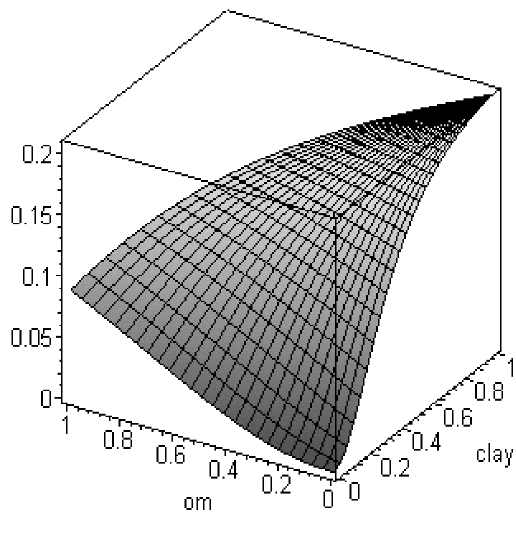


Fig. 31. Permanent wilting point variations (PWPw/FCw).

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CHAPTER 7

MODEL IMPROVEMENTS

7.1 INTRODUCTION

The existing ForHyM model was already giving good predictions about soil temperatures and water fluxes in forested watersheds, and the results were published (Arp and Yin 1992; Yin and Arp 1993). These earlier simulations were done using mean monthly air temperature and precipitations records. In the present work, daily temperature and precipitation data are used to give more accurate and realistic prediction about snowpack depth, snowmelt and stream discharge in relation to daily variations in weather, and in relation to variations in forest cover type. This Chapter focuses on various model improvements, as follows:

- Soil parameters such as porosity, saturated hydraulic conductivity, field capacity, and permanent wilting point needed to be re-programmed, as explained in Chapter 6.

- Heat capacities and heat conductivities of soil layers needed to be re-programmed based on the analysis in Chapter 6 and other considerations.

- Calculations concerning water flow and ice formation in soil and snow needed to be reformulated, to improve the overall placement of the calculations within the wider climate, landscape or watershed context, i.e., the model needs to respond – explicitly – to changes in watershed aspect and slope.

- The revised version includes the simulation of snowpack density, the original version simply used empirical assignments for snowpack densities, as these would vary across climatic regions.
- Ion fluxes as these would emerge from the melting snowpack are also calculated.

7.2 SOIL PARAMETERS

As detailed in Chapter 6, bulk density, soil porosity, field capacity and permanent wilting point can now be expressed as functions of soil texture, OM content and depth. These equations are applied to each layer of soil in the model. The user of the model specifies thickness (in cm), texture (sand, silt and clay content), and OM of each layer. Usually, in soil data books, the texture is analyzed after OM has been removed, so that:

$$sand_{texture} + clay_{texture} + silt_{texture} = 1$$

for the mineral part of the fine earth fraction of the soil (particles ≤ 2mm)

In the new ForHyM model, these values are recalculated so that:

$$(sand + clay + silt) + OM = 1$$

Doing this is necessary in order to comply with the equations of Chapter 6. For example the actual sand percentage to be used is:

$$sand = sand_{texture} * (1- OM)$$

Then, bulk density needs to be calculated as follows:

$$Db = \frac{1.23 + (1.37 - 0.72 * sand) * (1 - exp^{(-0.011 * depth)})}{1 + 7.21 * OM} \qquad (6)$$

The saturated hydraulic conductivity (in cm per hour) is then given by:

$$\text{Log10 (Ksat)} = -0.91 + 7.22*\text{Log10 (2.6-Db)} + 1.99*\text{sand}$$

The soil particle density (D_p) is obtained from:

$$\frac{1}{Dp} = \frac{OM}{D_{OM}} + \frac{1-OM}{D_{min}}$$

with D_{OM} = 1.1 and D_{min} = 2.6

From this, the soil pore space is calculated:

$$SPv = \frac{V_{pores}}{V_{total}} = 1 - \frac{Db}{Dp}$$

Then, the field capacity is obtained as a fraction of the volume of total soil, and of soil pore space (or soil saturation point):

$$FCv = \frac{FCv}{SPv} * SPv = \frac{FCw}{SPw} * SPv$$

$$\frac{FCw}{SPw} = \left(1 - \exp^{(-0.32*(1-\text{sand})-8.82*\text{clay})}\right) * \left(OM + \frac{1-OM}{1+8.03*\exp^{(-2.45*Db)}}\right)$$

The volumetric permanent wilting point (PWPv) is given by:

$$PWPv = \frac{PWPv}{SPv} * SPv = \frac{PWPw}{SPw} * SPv = \frac{PWPw}{FCw} * \frac{FCw}{SPw} * SPv$$

$$\frac{PWPw}{FCw} = \frac{1}{1 + \frac{FCw^{1.93}}{\left(1 - \exp^{(-1.13*\text{clay} - 0.077*\text{sand} - 4.06*OM)}\right)^{2.63}}}$$

For more details about these calculations see Appendix 1.

7.3 ESTIMATING THERMAL CONDUCTIVITIES

"Thermal conductivity is defined as the amount of heat transferred though a unit area in unit time under a unit temperature gradient "(Hillel 1980). In the model, it was decided to adopt the conductivity representation for frozen and unfrozen soils from Andersland and Ladanyi (1994):

$$K_{therm} = (K_{saturation}-K_{dry})*Ke + K_{dry}$$

With $K_{saturation}$ and K_{dry} as the saturated and dry thermal conductivities of the soil, respectively. Here, Ke is the Kersten number, and this number is a function of the degree of water saturation of the soil, i.e, Sr, with:

$$Sr = \frac{V_{water} + V_{ice}}{V_{pores}}$$

In general, for frozen soils:

$$Ke = Sr$$

And for unfrozen soils with Sr >0.1:

$$Ke = 1 + \log(Sr)$$

Also:

$$K_{dry}\ (W\ m^{-1}K^{-1}) = \frac{0.137 * Db + 64.7}{2700 - 0.947 * Db}$$

with Db as soil bulk density, in kg m^{-3}

For saturated unfrozen soils:

$$K_{saturation} = K_{solid}^{1-n} * K_{water}^{n}$$

For saturated frozen soils:

$$K_{saturation} = K_{solid}^{1-n} * K_{ice}^{n-wu} K_{water}^{wu}$$

where n is the soil porosity, wu is the volumetric fraction of unfrozen water, K_{water} = 0.57 W $m^{-1}K^{-1}$, K_{ice} = 2.2 W $m^{-1}K^{-1}$; K_{solid} is the thermal conductivity of the solid fraction of the soil.

K_{solid} is determined from the geometric mean:

$$K_{solid} = K_{quartz}^{q} * K_{0}^{1-q}$$

with q the quartz fraction of the total solid content, K_{quartz} = 7.7 W $m^{-1}K^{-1}$, K_0 = 2 W $m^{-1}K^{-1}$

For the ForHyM calculations, OM impact on K also needs to be considered. Therefore:

$$K_0 = K_{OM}^{OM} * K_{mineral}^{1-OM}$$

with K_{OM} = 0.25 W $m^{-1}K^{-1}$ (from Hillel 1980), and $K_{mineral}$ = 2 W $m^{-1}K^{-1}$ as before.

Data from Scheffer and Schachtschabel (1976) were used to determine quartz content in mineral soils. It was found that the mean quartz content of sand, silt and clay amounts to 65%, 47% and 20%, respectively, in general.

For unfrozen soils with water saturation smaller than 10%, the Kersten number is not defined. However, for Sr = 0, the soil is dry, so $K_{therm} = K_{dry}$. At this limit, the limiting value for Ke is 0. When Sr is slightly greater than 0, then Ke must also be close to 0. As Sr approaches 1, Ke needs to conform with 1 + log (Sr). Combining these considerations leads to the following generalized expression for Ke:

$$Ke = \left[\left(\frac{\exp^{13.6*Sr}}{2.3+\exp^{13.6*Sr}}\right)^3 - \left(\frac{1-Sr}{2.3+1}\right)^3\right] * \sqrt{Sr} \qquad (5)$$

Figure 32 illustrates how Ke varies from Sr equals 0 to 1.

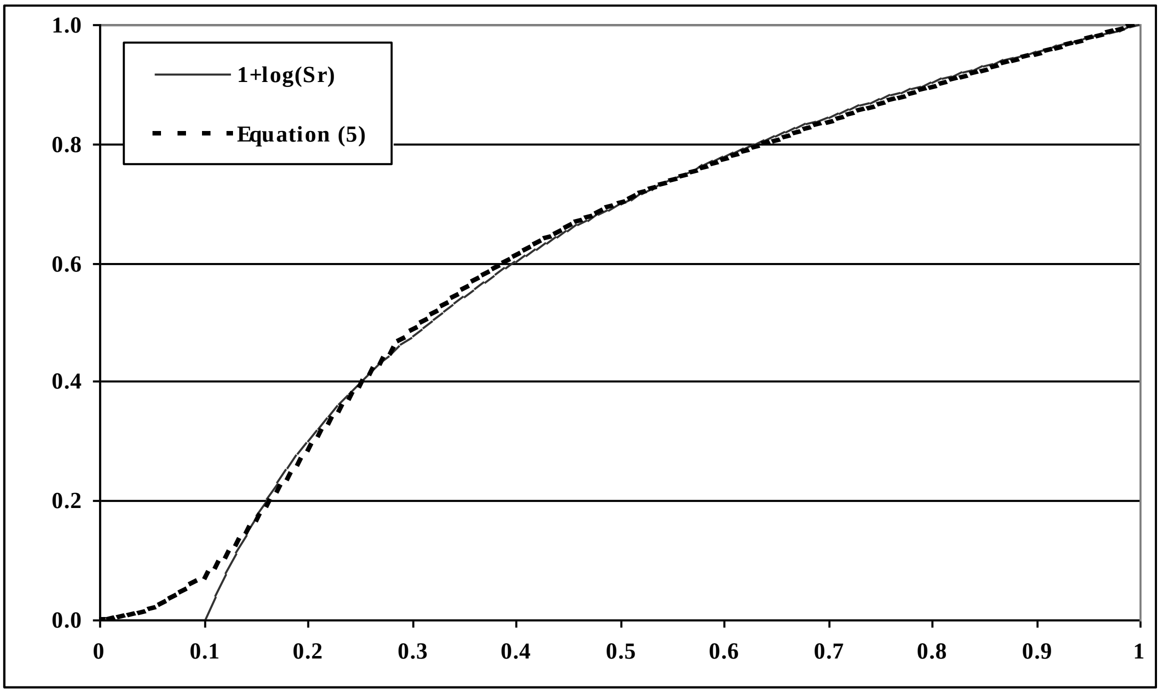

Fig. 32. Kersten number for unfrozen soils.

7.4 HEAT CAPACITIES

The heat capacity of a substance is the quantity of heat that is necessary to increase the temperature of this substance by one degree °C. A soil is made up of five substances: minerals, OM, water, ice and air. Its overall volumetric heat capacity is – therefore – the weighted average of the heat capacities (HC) of each substance. In this, the weighing must be done by the volumetric fractions (Vfr) of the components. The resulting formula is given by:

$$HC_{vol}^{soil} = Vfr_{mineral} * HC_{vol}^{mineral} + Vfr_{OM} * HC_{vol}^{OM} + Vfr_{water} * HC_{vol}^{water} + Vfr_{ice} * HC_{vol}^{ice} + Vfr_{air} * HC_{vol}^{air}$$

Table 7 shows the values of specific heat capacities by substance type.

Table 7. Volumetric heat capacities of soil components, after Hillel (1980)

	Volumetric heat capacity	
	cal cm^{-3} °C^{-1}	W cm^{-3} °C^{-1}
Minerals	0.48	$2*10^6$
OM	0.6	$2.5*10^6$
Water	1	$4.2*10^6$
Ice	0.45	$1.9*10^6$
Air	0.003	$1.25*10^4$

7.5 VOLUMETRIC FRACTIONS

The volumetric fraction of OM in the soil is linked to the mass fraction of OM (OM) as follows:

$$Vfr_{OM} = OM * \frac{Db}{D_{OM}}$$

This is because

$$Vfr_{OM} = \frac{V_{OM}}{V_{soil}} = \frac{W_{OM}}{D_{OM}} * \frac{Db}{W_{soil}}$$

and

$$OM = \frac{W_{OM}}{W_{soil}}$$

Similarly, the volumetric fraction of minerals in the soil is given by

$$Vfr_{min} = (sand + silt + clay) * \frac{Db}{D_{min}} = (1 - OM) * \frac{Db}{D_{min}}$$

and the volumetric fractions of water, ice and air are given by:

$$Vfr_{wat} = \frac{V_{water}}{V_{soil}}$$

$$Vfr_{ice} = \frac{V_{ice}}{V_{soil}}$$

$$Vfr_{air} = SPv - Vfr_{water} - Vfr_{ice} = \frac{V_{air}}{V_{soil}}$$

The sum of the volumetric fractions of water, ice and air is always equal to the volumetric pore space SPv. Moreover,

$$Vfr_{OM} + Vfr_{min} = \frac{Db}{Dp}$$

because

$$\frac{1}{Dp} = \frac{OM}{D_{OM}} + \frac{1-OM}{D_{min}}$$

and

$$SPv = 1 - \frac{Db}{Dp}$$

The sum of the five volumetric fractions is always equal to one. One important condition to respect is Db < Dp. Note that Db depends on three variables (OM, sand and depth) and Dp depends only on OM. For example, Equation 7 gives the maximum values of Db (based on equation 6 with infinite depth and sand = 0) as a function of OM only. Hence,

$$Db_{max} = \frac{2.6}{1+7.21*OM} \qquad (7)$$

The comparison between Db_{max} and Dp is illustrated in Figure 33. Db is always smaller than Dp.

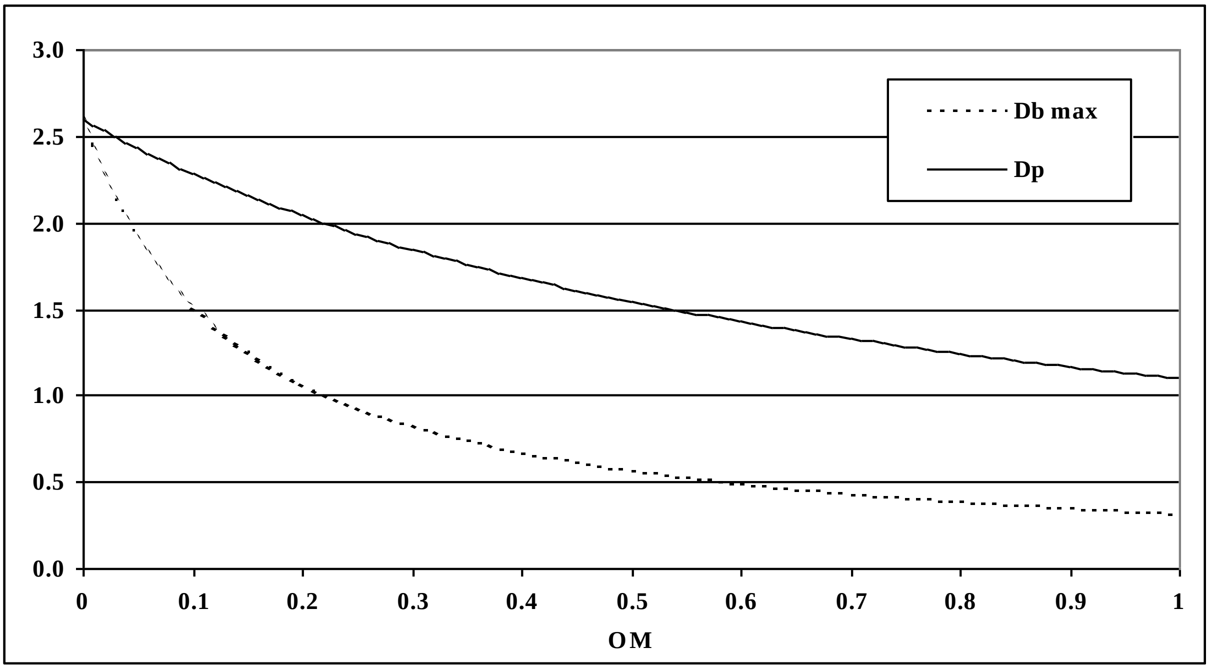

Fig. 33. Comparative values of bulk density and particle density (in g cm^{-3}) with increasing soil OM content.

7.6 ICE FORMATION

A major interaction between energy flux and water flux occurs when the soil reaches a temperature of 0°C. At this point, water undergoes a phase change. The model uses the latent heat of fusion (LHF) to quantify this conversion. This heat is equal to 79.67 cal g^{-1} (Lange 1956). The way the model performs ice calculations is as follows:

- Based on the soil temperatures at time **t,** called **Ti(t),** and the new boundary temperature at the surface, new temporary temperature values are assigned for of any layer **i,** called **Ti'**. These temporary temperatures are calculated byway of the implicit finite difference algorithm by Patankar (1980) , see Chapter 5.

- For any layer, if - for example - **Ti (t)** is positive (no ice in the layer) and **Ti'** is also positive, then no ice is made, and **Ti (t+dt)** is set equal to **Ti'**. If **Ti**' is negative, some ice is made until:

  either all the liquid water in the layer becomes frozen; then

$$Ti\ (t+dt) = (Ti' + \Delta T) < 0$$

  where:

$$\Delta T = \frac{\text{Energy released by freezing}}{\text{Heat capacity of the layer}} \geq 0$$

  or enough energy is released by the process of freezing to bring the temperature of the layer back to 0°C. Then **Ti (t+dt) = 0°C** and the amount of water that freezes is given by:

$$w = \frac{-T'i * \text{Heat capacity of the layer}}{\text{LHF}}$$

Situations when ice is formed occur when:

$Ti \geq 0$ and $Ti' < 0$

$Ti < 0$ and $Ti' < 0$ and if some water just flowed in the layer.

The ice present in the soil starts to melt when:

$Ti \leq 0$ and $Ti' > 0$

The equations governing melting of ice are similar to the equations for freezing, except that ΔT is negative. The temperature of a soil layer cannot become positive until all the ice has melted. In each case:

$$\Delta T = \frac{\text{Energy consumed by melting}}{\text{Heat capacity of the layer}}$$

The snowpack is treated in the same way as the soil layers. It has its own field capacity and can retain liquid water if it rains or if some snow melts. If - later - the air temperature becomes negative again, the liquid water in the snowpack is set to freeze again. The equations are the same as the ones cited above.

7.7 HYDROLOGY

Foliage intercepts incoming rain and snow. The model calculates the vegetation area index (VAI), which is expressed in m^2/m^2. It is the cumulative one-sided surface of all the leaves and branches that are above one square meter of soil. The values used in the model are 6.5 m^2/m^2 for a mature hardwood forest in summer, and 1 m^2/m^2 in the winter, with seasonal change induced by cumulative air temperature in the spring, and decreasing day length in the fall. For the mature conifer forest, VAI = 9 m^2/m^2 all year.

In the model, the canopy is given a specific field capacity; this capacity is assumed to be proportional to VAI. In general, canopy intercepted rain is set to evaporate, at a rate that depends on air temperature. The warmer the air is, the faster evaporation occurs. Intercepted snow is set to evaporate also, but more slowly because of cold conditions. Some of the intercepted snow is set to return to the ground to mimic several processes: wind, sun-induced melting within the canopy, and/or rain.

The snow on the ground is set to two ways how to retain water: by freezing if the snowpack temperature is negative, and on account of the field capacity of the snowpack. Details about the freezing process are given above. The default volumetric field capacity of the snowpack was chosen to be equal to the volumetric field capacity of uncompacted sand, which is around 7%. It is assumed that the grainy snow crystals within the snowpack exert a similar effect on water retention than sand particles.

The maximum amount of water that can be evaporated from the soil and from the trees by transpiration is called potential evapotranspiration (PET); PET depends on atmospheric conditions such as air temperature, relative humidity, wind speed, and day length. Actual evapotranspiration, (AET), in turn depends on leaf area, and soil moisture content. For a watershed partially covered with forest, a vegetation percentage (Veg %) is defined, to assess overall forest canopy coverage per watershed. If there is soil moisture available for evapotranspiration by trees, i.e. soil moisture > permanent wilting point, then:

$$\text{Evapotranspiration} = \left(\frac{\text{Water content} - \text{PWP}}{\text{Pore space} - \text{PWP}}\right)^{\text{Kevap}} * \text{PET} * \text{Veg\%}$$

where K_{evap} is a calibration parameter.

If water content is below the permanent wilting point, then there is no evapotranspiration through the vegetation.

Some evaporation may also occur from the first layer of the soil where there is no vegetation cover. This water is calculated from:

$$\text{Evaporation} = \left(\frac{\text{Water content}}{\text{Pore space}}\right)^{\text{Kevap}} * \text{PET} * (1 - \text{Veg\%})$$

Under saturated soil conditions, flow rates into and through the soils are calculated with Darcy's law:

$$Q = \frac{K_{sat} * \Delta P * A}{L}$$

where:

K_{sat} is the saturated hydraulic conductivity of the soil layer
ΔP is the hydrostatic pressure difference from the top to the bottom of the soil layer
A is the cross-sectional area of the column through which the water flows
L is the thickness of the layer.

But most of the time, the soil layers are not saturated. For unsaturated conditions, water can flow only from one soil layer to the next, or from one location in the same layer to another if the moisture content is above field capacity. In this case, the permeability of the soil is adjusted by:

$$Perm = K_{sat} * \frac{(Water_{layer} - FC_{layer})}{(Vpores_{layer} - FC_{layer})}$$

There are two kinds of flows in the model: **Vertical flows between layers**, called infiltration flows. These are defined by:

$$Q = \frac{Perm * \Delta P * A_v}{L}$$

where:

A_v is the cross-section for vertical flow, its surface is equal to one unit area (see Figure 34),
L is the thickness of the layer considered, and
ΔP is the height of the water column in the layer considered.
This height is equal to 0 when the water content is equal to field capacity, and is equal to the thickness of the layer when the water content is at saturation, i.e.,

$$\Delta P = Thickness_{layer} \frac{(Water_{layer} - FC_{layer})}{(Vpores_{layer} - FC_{layer})}$$

Lateral intra-layers flows occur along slopes. These flows contribute directly to stream discharge, at the watershed scale, and depend on the slope as well as the water content of the layer that conducts this water to the stream. These flows follow the same general equation namely,

$$Q=\frac{Perm*\Delta P*A_l}{L}$$

where $\frac{\Delta P}{L}$ is the mean slope of the landscape at the watershed scale, and:

$$A_l=Thickness_{layer}\frac{(Water_{layer}-FC_{layer})}{(SPv_{layer}-FC_{layer})}*Width$$

with Width = 1 the unit width of the soil column (see Figure 34)

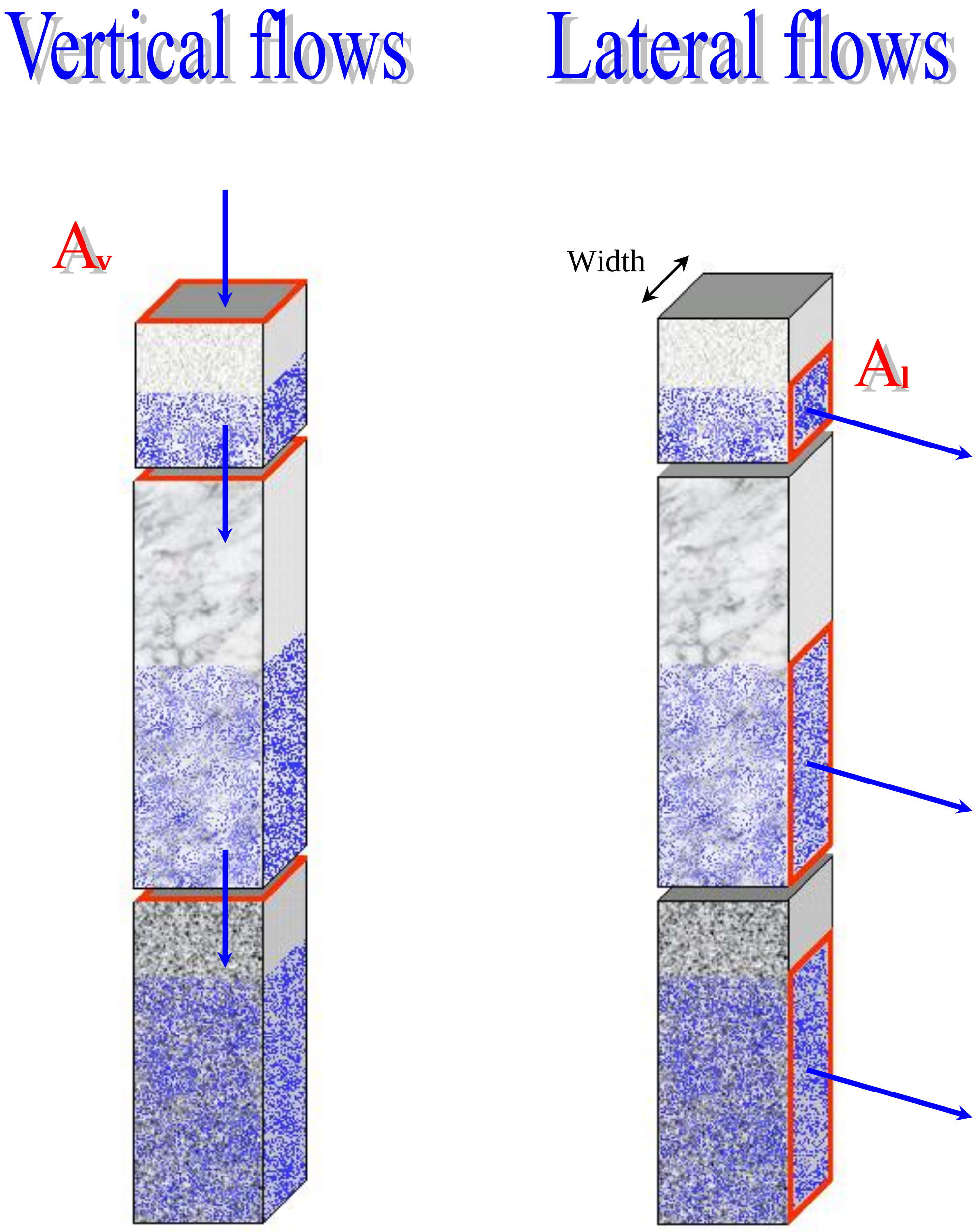

Fig. 34. Vertical and lateral flows in the model.

7.8 SNOW DENSITY

The mean density of the snowpack changes frequently during the winter: it decreases when a new snowfall occurs (freshly fallen snow has a density close to 0.15 g cm^{-3}), and increases whenever it rains or if some of the snowpack melts. The model uses the following equation to calculate snow density at each time step:

$$\text{Density} = \frac{\text{Snowpack water equivalent}}{\text{Volume of snowpack}}$$

The density is updated at each time step, by taking into account the following events whenever they occur: (i) fresh incoming snow, (ii) snow melting, (iii) snow evaporating, and (iv) liquid water entering the snow, subject to freezing. All water fluxes are expressed in terms of water equivalents (WE).

Distinction is made between **dry** density of the snowpack (the equivalent of the bulk density for a soil) and **wet** density of the snowpack. Wet density takes into account the amount of liquid water retained in the snow for the calculation of the snowpack water equivalents. The equation used is the following:

$$\text{Density}_{\text{dry}}(t+dt) = \frac{(SD_{WE} - SM - SE_{WE}) + FS_{WE} + WF}{\dfrac{(SD_{WE} - SM - SE_{WE})}{\text{Density}_{\text{dry}}(t)} + \dfrac{FS_{WE}}{D_{FS}}}$$

with

SD_{WE} as the water equivalent of the dry snowpack
SM as the amount of snowmelt water
SE_{WE} as the amount of snow lost by evaporation
FS_{WE} as the amount of incoming fresh snow
WF as the amount of liquid water getting frozen in the snowpack
D_{FS} as the density of fresh snow

7.9 ION CONCENTRATIONS IN SNOWMELT

Concentrations of ions present in snow and rain need to be known to model ion fluxes that result from the accumulation of rain and snow within the snowpack. All ions deposited with snow and rain are originally stored within the snowpack. A part of the ions are then allowed to leave the snowpack whenever there is more rain and/or snowmelt water than the snow can retain. The ions quantities that could potentially leave the snowpack are given by:

$$Q_{ions\,leaving} = k_{release} * \frac{Water_{through\,snow}}{Water_{through\,snow} + Water_{in\ snow}} * Q_{ions\ in\ snowpack}$$

where $K_{release}$ is an adjustable parameter to capture the rate of release that is specific for each ion type (minimum 0, maximum 1). The quantities of ions leaving are estimated in equivalents per hectare per unit time.

The concentration of ions in snowmelt water is given by:

$$C_{ions\ in\ snowmelt} = \frac{Q_{ions\,leaving}}{Water_{through\,snow}}$$

Concerning the release of acid cations (mainly H^+), H^+ concentrations are obtained from the known pH values of the snowmelt solution. Conversely, estimated H^+ fluxes, when divided by the corresponding amount of snowmelt, can be converted into model calculated snowmelt pH values.

7.10 REFERENCES

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CHAPTER 8

MODEL CALIBRATION

8.1 INTRODUCTION

This Chapter describes how the revised ForHyM model was calibrated with the following field observations: (i) daily stream discharge and snow pack water equivalents at Kejimkujik National Park in Nova Scotia, and (ii) detailed snowpack data for snow accumulations for three snowpack conditions: in the open, and underneath hardwood and softwood forest cover in the UNB forest, in Fredericton, New Brunswick (Chapter 4). The variables that were used for model calibration are: snow density, snow depth, snow temperature, amount of snowmelt, amount and concentrations of ions (H^+, Ca^{2+}, NO_3^-, SO_4^{2-}) in snowmelt and in stream discharge.

Model calibration is a crucial point of the modeling process. It consists of comparing model outputs with actual data, and includes the adjustment of model parameters in order to get the best possible simulation. This process requires having good quality data for model input (rain, snow, air temperatures) and for model calibration (snow depth, stream discharge, stream and snow chemical composition). It is good to calibrate the model on as many sites as possible and for as many variables as possible. However, one is limited by the quantity and quality of data that are available.

8.2 STUDY AREAS

The forest of the University of New Brunswick is situated on a bedrock plateau south of Fredericton, at an altitude of 116m and latitude of 45°56'N, and is composed of various forest cover types. Up-to-date weather data for model input (rain, snow, air temperature) are available from the weather station located 12 km east at Fredericton airport, from the city (Figure 35). A -1°C correction was applied to the air temperature record because the forest is about 100m higher in altitude than the airport. Snowpack data as described in Chapter 4 were used to calibrate the model for snow depth, snow density, and snowmelt chemical composition for the winter of 2001-2002, for each snowpack conditions: coniferous forest, deciduous forest, and open area.

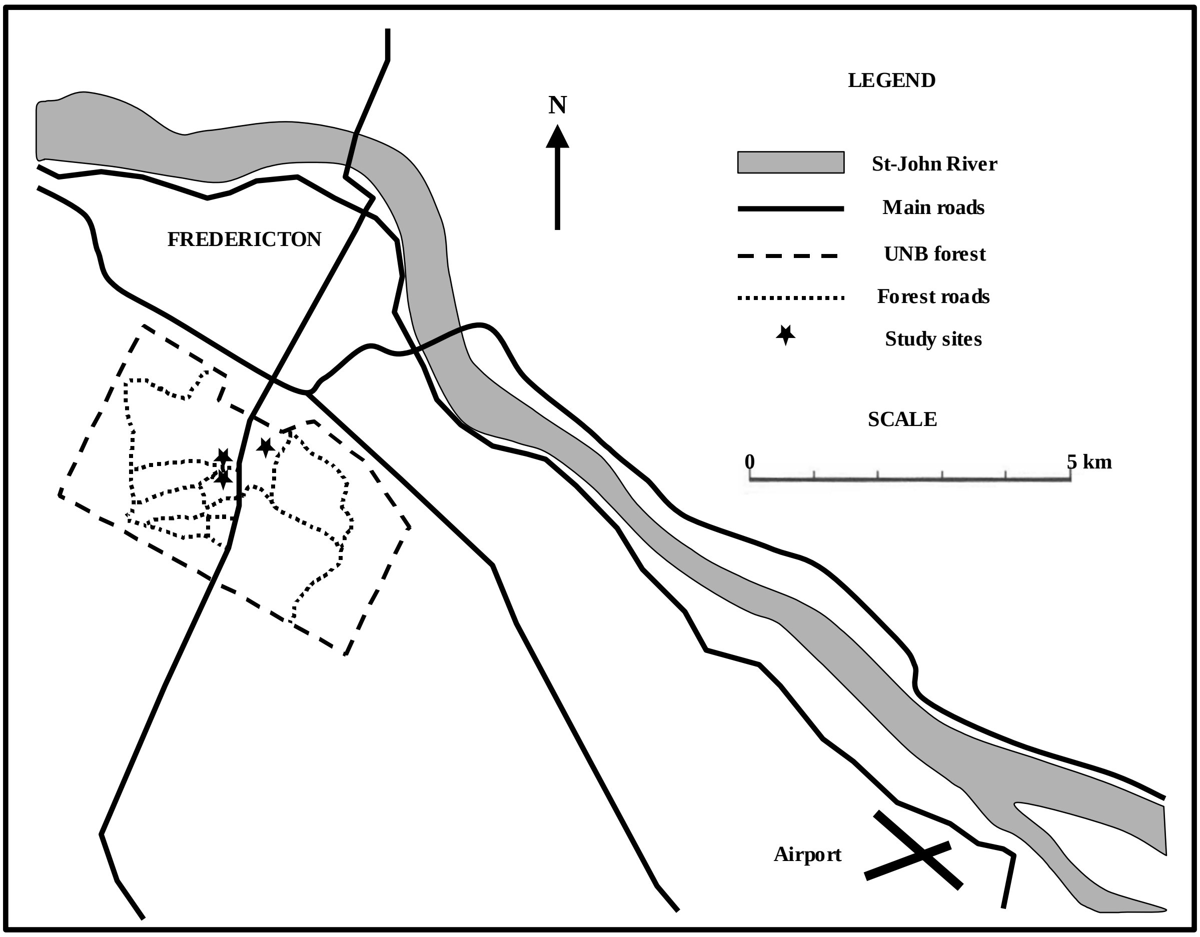

Fig. 35. Location of sites used for snowpack study and model calibration in Fredericton

Moosepit Brook drains a watershed covering an area of 17.7 km^2. This brook is located in Nova Scotia, at a latitude of 44°26'N, at an altitude of 127m, and is covered by a mixed forest (50% coniferous, 50% deciduous). Daily rain, snow, snow depth, air temperature and chemical atmospheric deposition data were obtained from the weather station at Kejimkujik National Park (Figure 36) for the period 1984-2000. Stream discharge was available for Moosepit Brook for the same time period. A generalized description of the soils within the Moosepit Brook catchment (Yanni 1995) was used to estimate texture, OM and soil depth within this catchment.

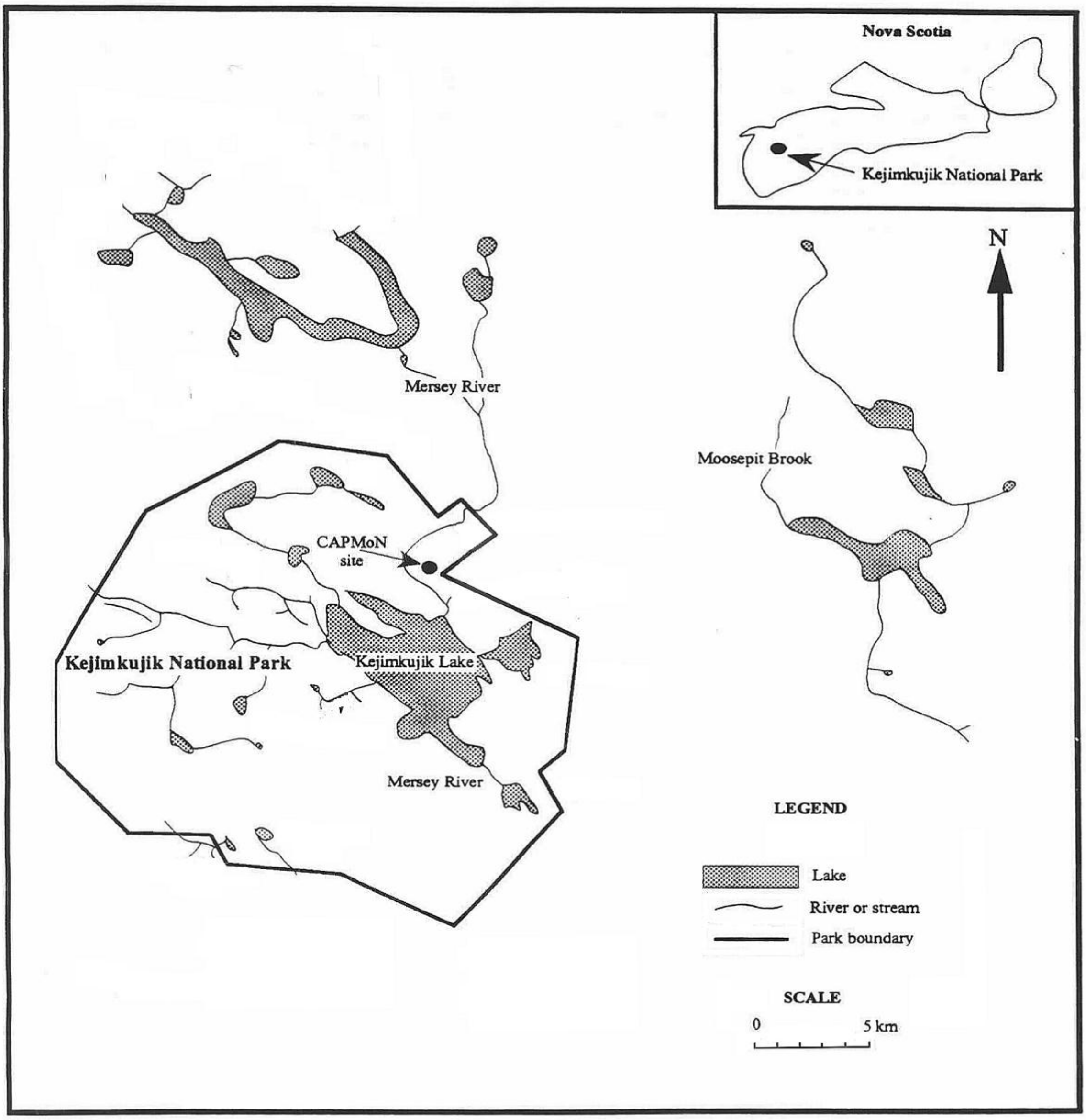

Fig. 36. Location of sites used for stream discharge and snowpack calibration in Nova Scotia. The weather station is situated at the CAPMoN site (after Yanni 1995)

8.3 MODEL CALIBRATION

Model calibration consists in running the model for several years, in order to gain first impressions about general model performance: are all the modeled outputs referring to snowpack accumulations, stream discharge, soil temperatures, soil moisture levels, soil frost penetration realistic? Next, if the calculations are realistic, how closely do they match the field observed data? Next, if there are mismatches, what are the likely causes of these mismatches? Are these based on faulty programming, or based on faulty data or faulty data input? Data could be faulty for several reasons:

- The data as presented are incorrectly used. Only an in-depth analysis of the metadata can reveal how available data should actually be applied within the modeling context.
- The data are correct, but they have not been entered properly. This is very common: often systematic mismatches appear when, e.g., data for snow, rain, air temperature, stream discharge are not exactly cross-referenced to actual dates. In this case, each data entry must be examined carefully, to ensure that model inputs are perfectly synchronized with one another, and with the internal time step of the model calculations.
- Other data input problems refer to ensuring proper data conversion, to conform with the exact data specifications of the model; for example, all data referring to ForHyM fluxes must be converted to the per hectare per year scale.
- Data that have been used may require further quality control checks. Special procedures must be followed to determine values for missing values. The occurrence of missing data is quite common. Often, missing data can be estimated to other con-current variables. For example, if air temperature is missing for one location for certain days, then regression

equations between that weather station and other stations nearby should be used to fill in the data, with reasonable reliability.

In order to produce numerically stable model output, it is best to start hydrological model calculations in early fall, i.e., September, when daily variations in soil moisture and soil temperature are the least in any year. To further improve on model initialization, it is important to check whether the model performs well from the start. Good initial conditions are those that do not produce major or minor oscillations or jumps in any of the model output during the first and subsequent time step calculations. For proper soil moisture and temperature initialization, it is best to run the model for a complete year, and re-initialize the soil moisture and temperature "boxes" in *Stella* with the values obtained in the following September. Invariably, these numbers tend to produce a reasonable initial solution. However, additional checking may be required to avoid initial jumps in the soil moisture and temperature calculations altogether.

Once all data entries are what they should be, further mismatches between model output and field observations could be due to improper model formulation. For example, the model output for stream discharge may not be as peaked as the field-observed values. In this case, adjustments are necessary to direct more water to flow as run-off, or the permeability of the soil has to be increased. In another instance, cumulative discharge may fall below field-observed cumulative discharge. In this case, one should examine whether the AET calculations are done properly, or whether some adjustments are needed in this respect. In another instance, calculated snowmelt may occur too early in spring on account of a high heat transfer coefficient.

Since ion release from the snowpack is assumed to be proportional to ion content within the snowpack, adjustments need to be made to the corresponding proportionality coefficient until the predicted ion concentration in the snowmelt generally agrees with field-observed values.

Since field-observed values are cumulative for each time period between sampling, it is important to compare volume weighted ion concentrations rather than instantaneous ion concentrations.

8.4 RESULTS AND DISCUSSIONS

Best-fitted results for the cumulative discharge for Moosepit Brook are shown in Figure 37 for a period of 15 years. As shown, cumulative calculated and cumulative field observations are generally quite close over the entire calculation period. Some of the short-term discrepancies (annual ripples) are either due to too early snowmelt timing or due to occasional errors in the snowfall data or in the stream discharge data during winter. For some years, calculated stream discharge was a bit higher in the fall and a bit lower in the spring than what was actually measured. This seems to suggest that the Moosepit Brook basin has a slightly higher water retention capacity in the fall, and a slightly smaller water retention capacity in the spring. This is quite possible because the model does not consider water storage within the catchment in any other media than soil and subsoil.

To achieve the long-term correspondence between calculated and field observed stream discharge values in Figure 37, it is necessary to set canopy snow retention to 1 mm and rain interception to 0.25 mm water equivalent per unit of leaf area. Also, since AET is sensitive to soil moisture availability, it was necessary to adjust the relationship between AET and soil moisture conditions as follows:

$$\text{Evapotranspiration} = \left(\frac{\text{Water content} - \text{PWP}}{\text{Pore space} - \text{PWP}} \right)^{\text{Kevap}} * \text{PET} * \text{Veg\%}$$

with $K_{evap} = 0.5$.

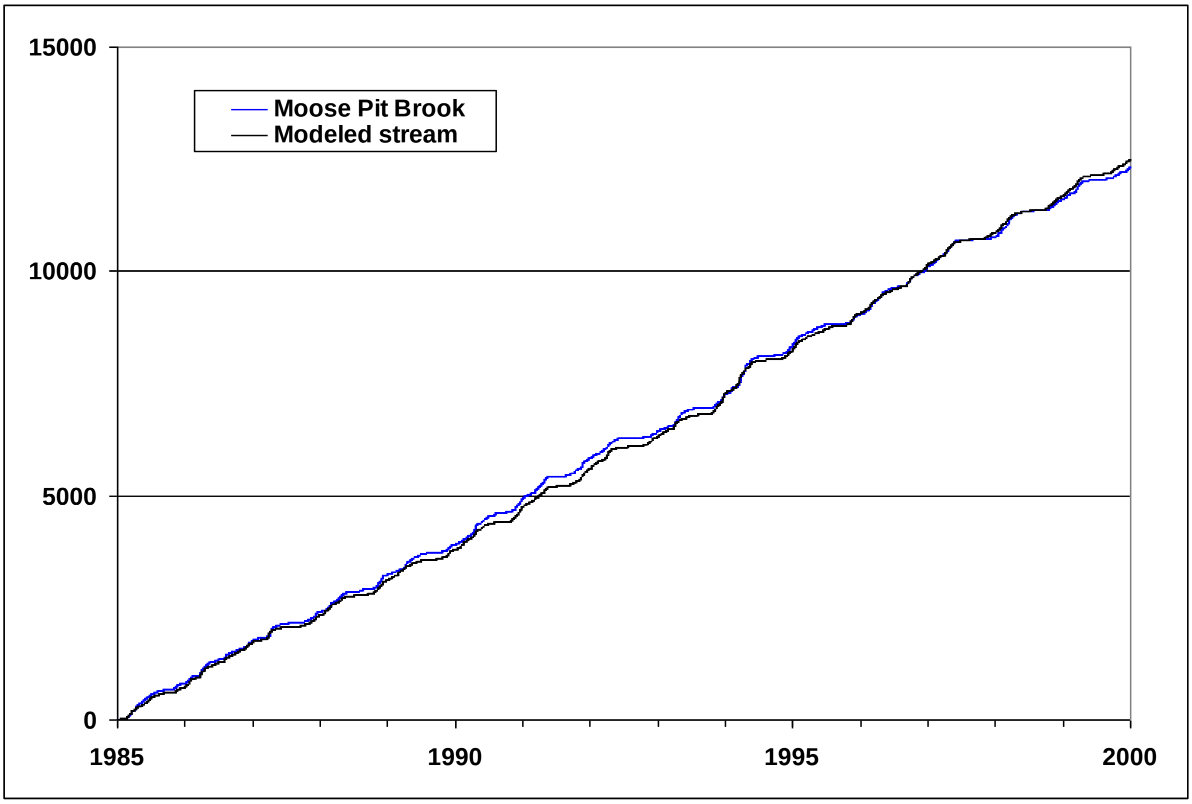

Fig. 37. Comparison of measured and simulated cumulative stream discharge (in mm).

Figure 38 shows modeled and measured snow depth values for KNP. The simulation was done for a situation with no forest, since the snow depth data were measured at the weather station, i.e., outside of the forest. For some years, the model gave acceptable results. For others, the modeled snowpack was not thick enough. It is suspected also that the snow depth data in these years may not be accurate, or may not represent the actual snow amounts on the ground on a per hectare basis. Variations in local wind conditions are likely the greatest source in terms of generating local variations in snow depth. Palmer (1988) showed that wind speed affects snow distributions, and snow fall measurements as well.

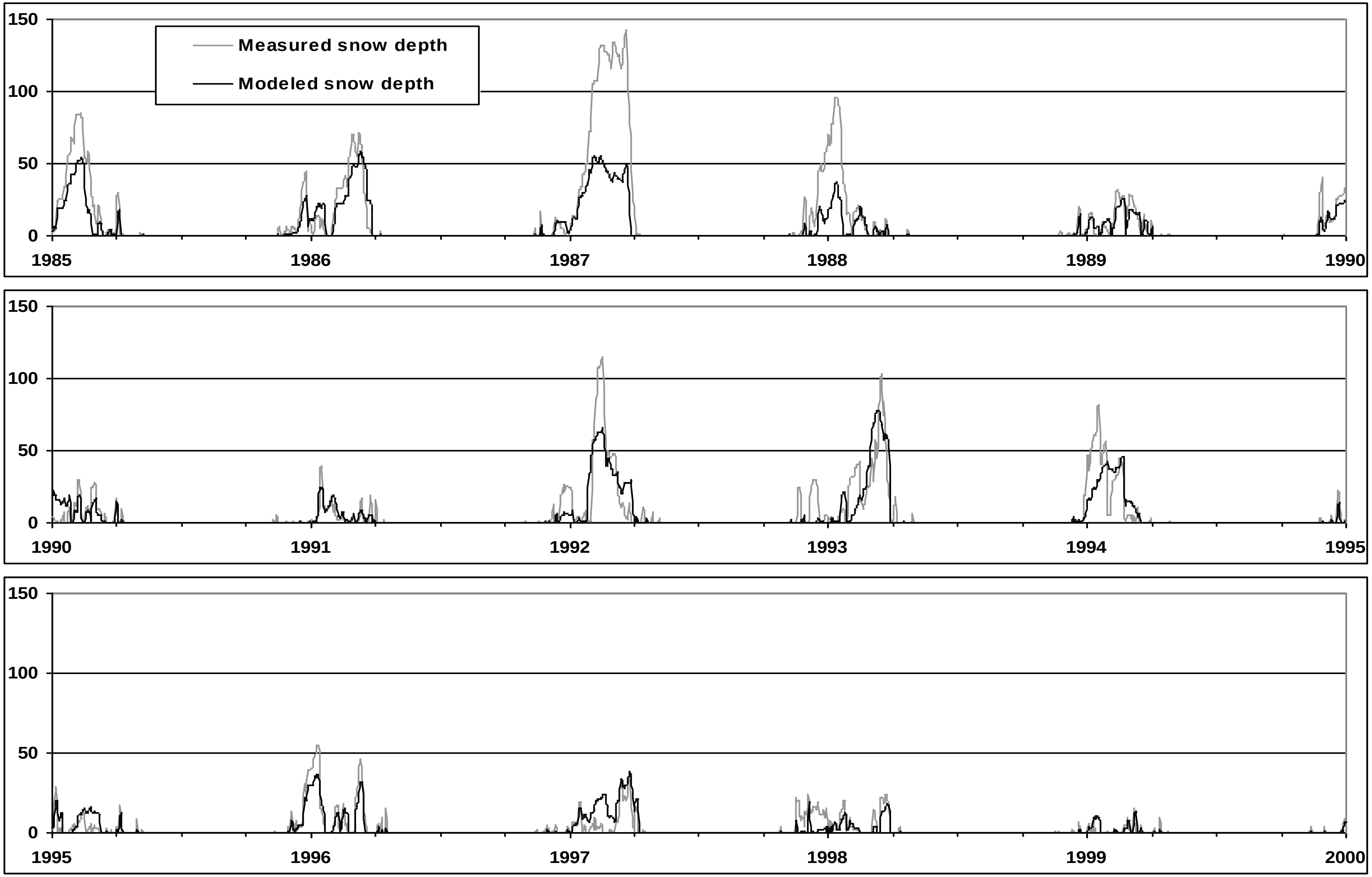

Fig. 38. Comparison of measured and simulated thickness of snowpack (in cm) in Kejimkujik National Park.

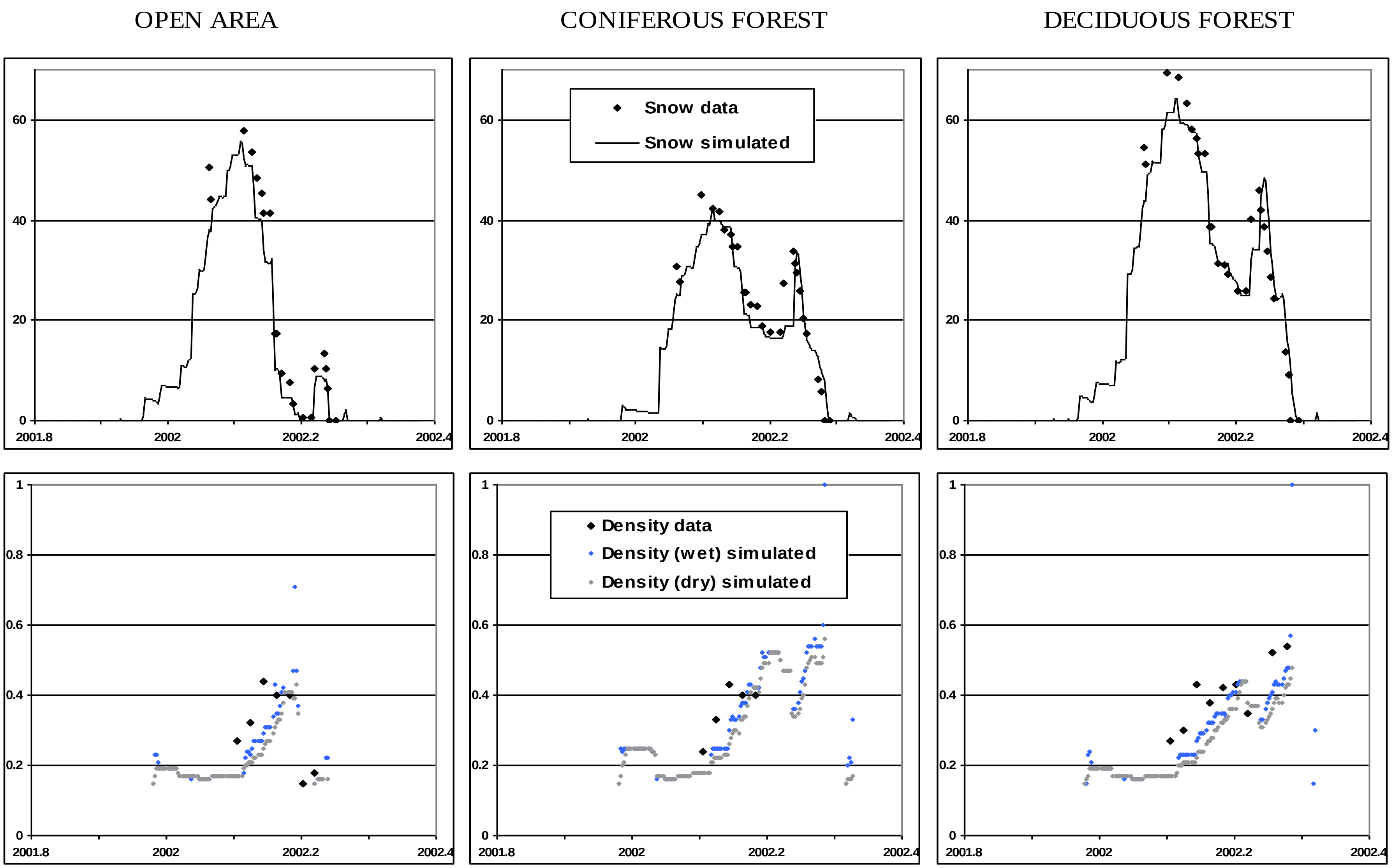

Fig. 39. Snow density (in g cm^{-3}) and thickness of snowpack (in cm), measured and simulated for Fredericton.

Figure 39 shows the modeled and measured snow depth and snow density for Fredericton. Calculated snow depth results are in good agreement with field-observed values for the open area and for the softwood forest. For the hardwood forest, a number of extra considerations needed to be made to get good agreement: this site is a shaded area (on northern slope north of a conifer forest, leading to a local depression further north). As such, this site would receive considerably less solar radiation than a hardwood site on a flat or on a southern slope, especially during winter, when low sun angles prevail during the entire day. The hardwood site also received more snow than the open area: being a hardwood stand near a depression, snow catch could be higher than in the neighboring area because of greater canopy roughness. Good results could only be obtained by assuming that this site must have received 30% more snowfall, and received 3 times less sunshine compared to the open area.

In all of the snow density calculations, it was also important to reduce the field capacity of the snow from 7% (earlier suggested value, similar to that of sand) to 5% of the volume of the snowpack. The density of fresh snow was chosen to be 0.15 g cm^{-3}, as observed. Smaller values such as 0.1 g cm^{-3} were tried, but this increased the overall effectiveness of the snowpack as a heat insulator for the soil, thereby producing higher than observed snowmelt amounts. Rain dropping into the snow added to the overall snowpack density, but the field observations indicated that the snow density should have increased even further than adding rain to the snow alone. Therefore, a snow compaction process was added, and this process would come into effect whenever daily snow water equivalent > 2mm or daily rain>0, such that:

$$\mathrm{Density}_{\mathrm{dry}}(t+dt) = K_{\mathrm{compaction}} * \frac{(SD_{WE} - SM - SE_{WE}) + FS_{WE} + WF}{\dfrac{(SD_{WE} - SM - SE_{WE})}{\mathrm{Density}_{\mathrm{dry}}(t)} + \dfrac{FS_{WE}}{D_{FS}}}$$

with:

SD_{WE} as the water equivalent of the dry snowpack
SM as the amount of snowmelt water
SE_{WE} as the amount of snow lost by evaporation
FS_{WE} as the amount of incoming fresh snow
WF as the amount of liquid water getting frozen in the snowpack
D_{FS} as the density of fresh snow

The compaction parameter gave the best results for the hardwood forest, mixed forest (KNP) and open areas with $K_{compact} = 1.015$. For the softwood condition, $K_{compact} = 1.025$.

For snowmelt, it was further necessary to reduce the rate of sensible heat transfer from the atmosphere to the snowpack (and vice versa), to account for the cold and moist air that normally rests on top of the melting snow when the air temperature is positive. The corresponding reduction factor was calibrated to be 0.4 for the four study locations.

Some adjustments needed to be made to the automatically calculated soil permeabilities in order to reproduce the stream discharge of Moosepit Brook. For example, infiltration flows and runoff flows needed to be adjusted to accommodate for the difference between local and watershed infiltration and runoff. Figures 40 to 43 show calculated and field-observed stream discharge over fifteen years. While there is an overall good agreement of modeled versus field-observed stream discharge, some discrepancies occurred occasionally with respect to:

- timing of snowmelt (1987 and 1993),
- water retention in the fall,
- spring stream discharge.

In order to assess the true nature of these discrepancies, one should carefully examine:

- The data records. For example the fact that there can be considerable variations in precipitation, especially since the weather station is not located within the catchment area.
- The actual terrain of Moosepit Brook, to detect areas that would fill up in the fall before there is any major change in stream discharge further below.

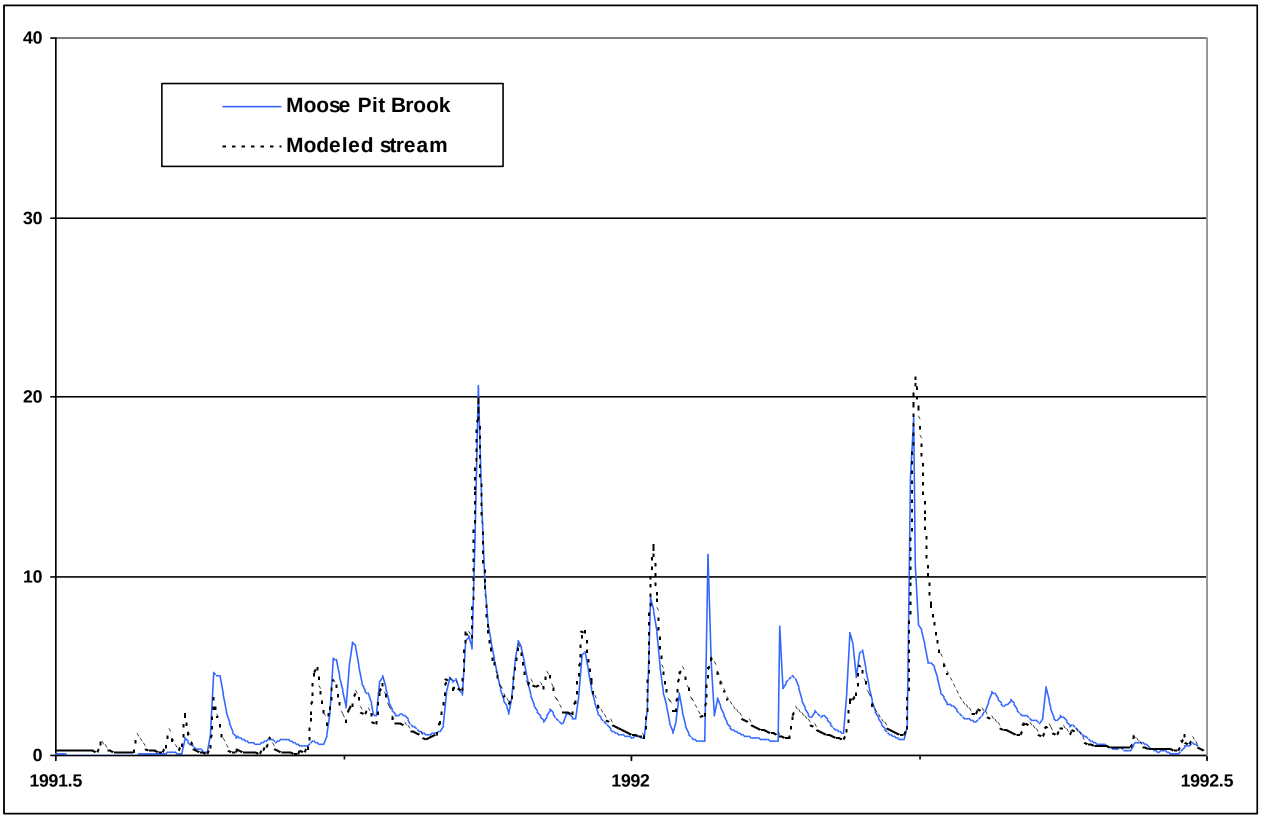

Fig. 40. Simulated and measured daily discharge (in mm day^{-1}) for Moosepit Brook, over a year.

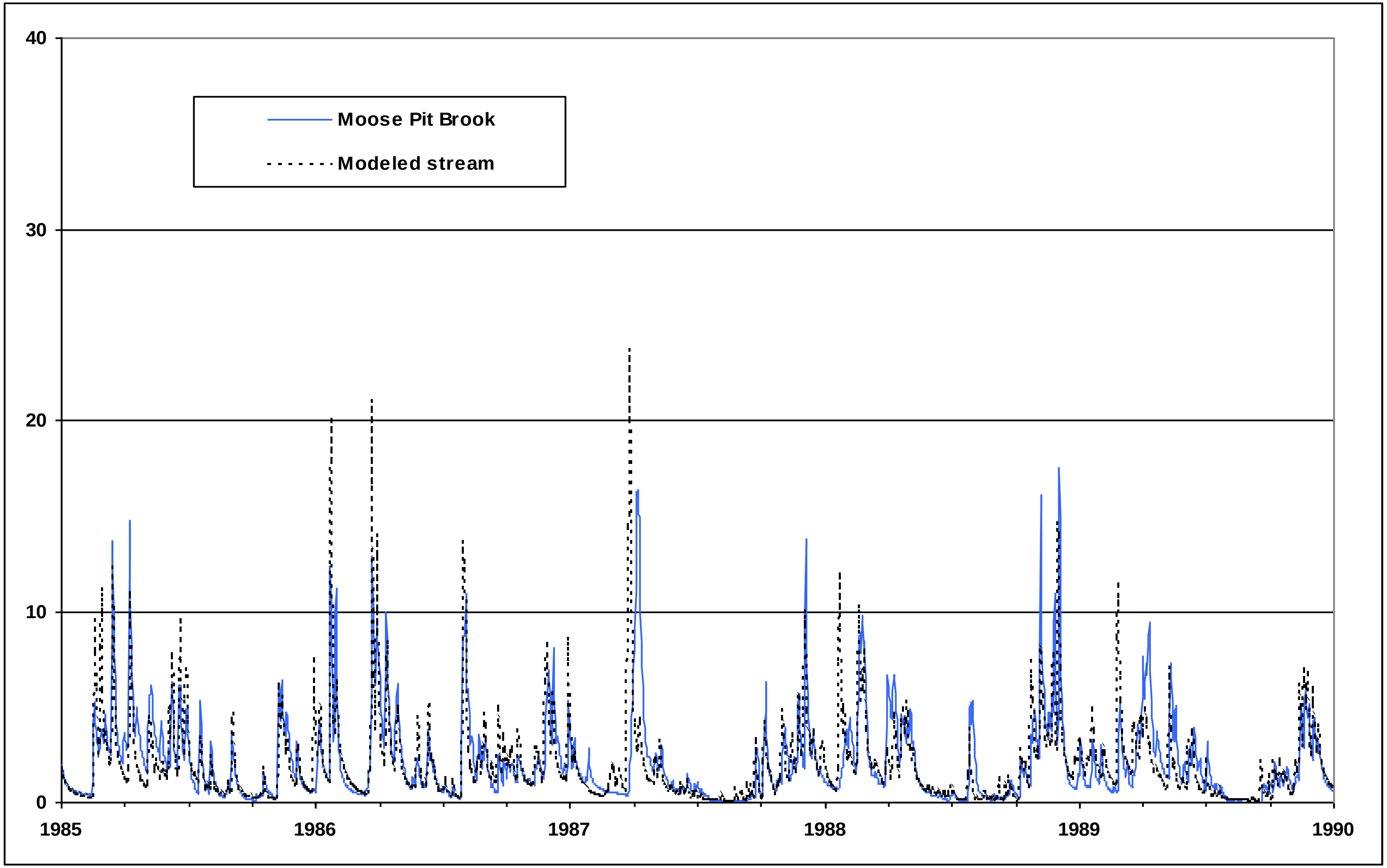

Fig. 41. Simulated and measured daily discharge (in mm day^{-1}) for Moosepit Brook from 1985 to 1990.

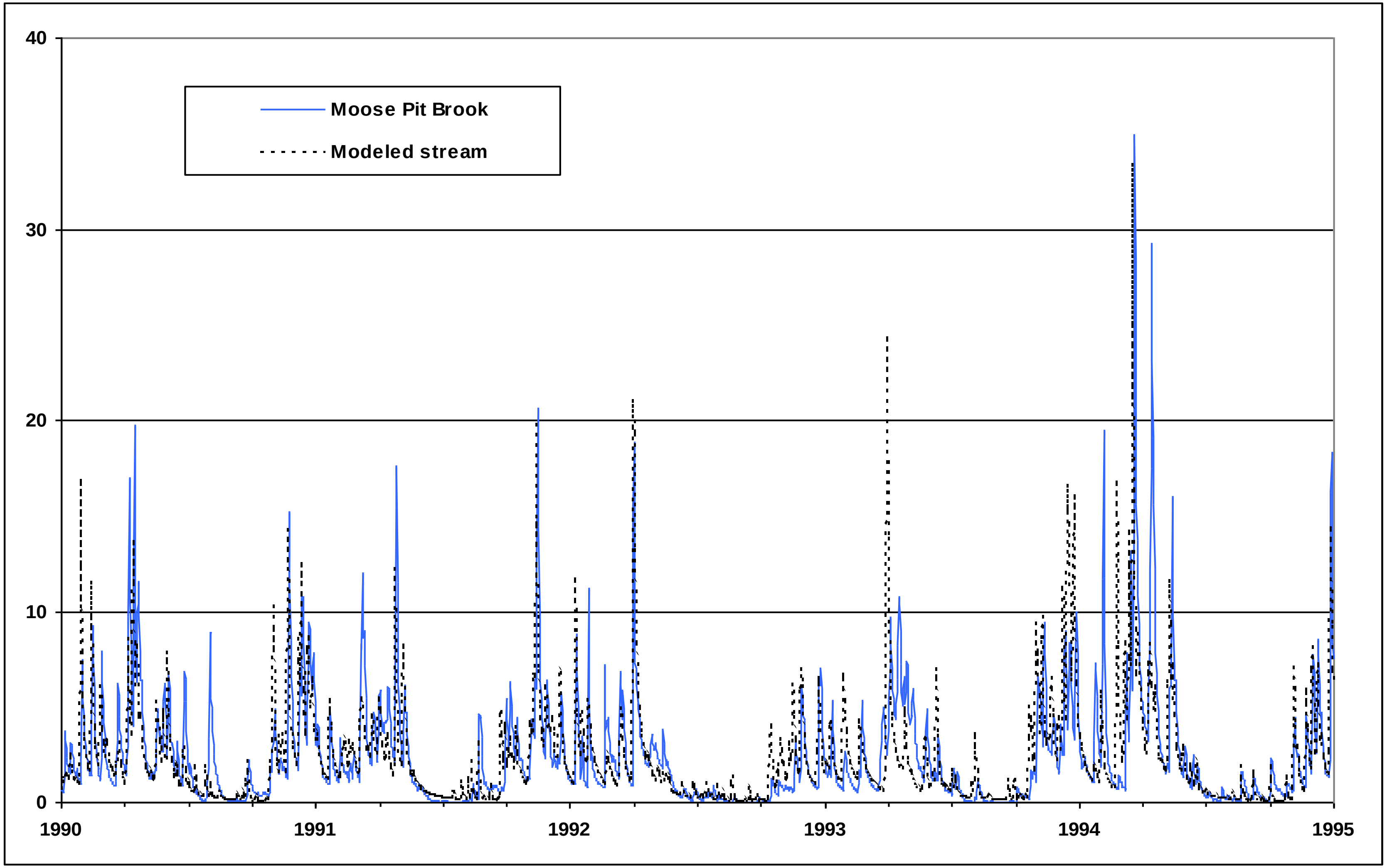

Fig. 42. Simulated and measured daily discharge (in mm day^{-1}) for Moosepit Brook from 1990 to 1995.

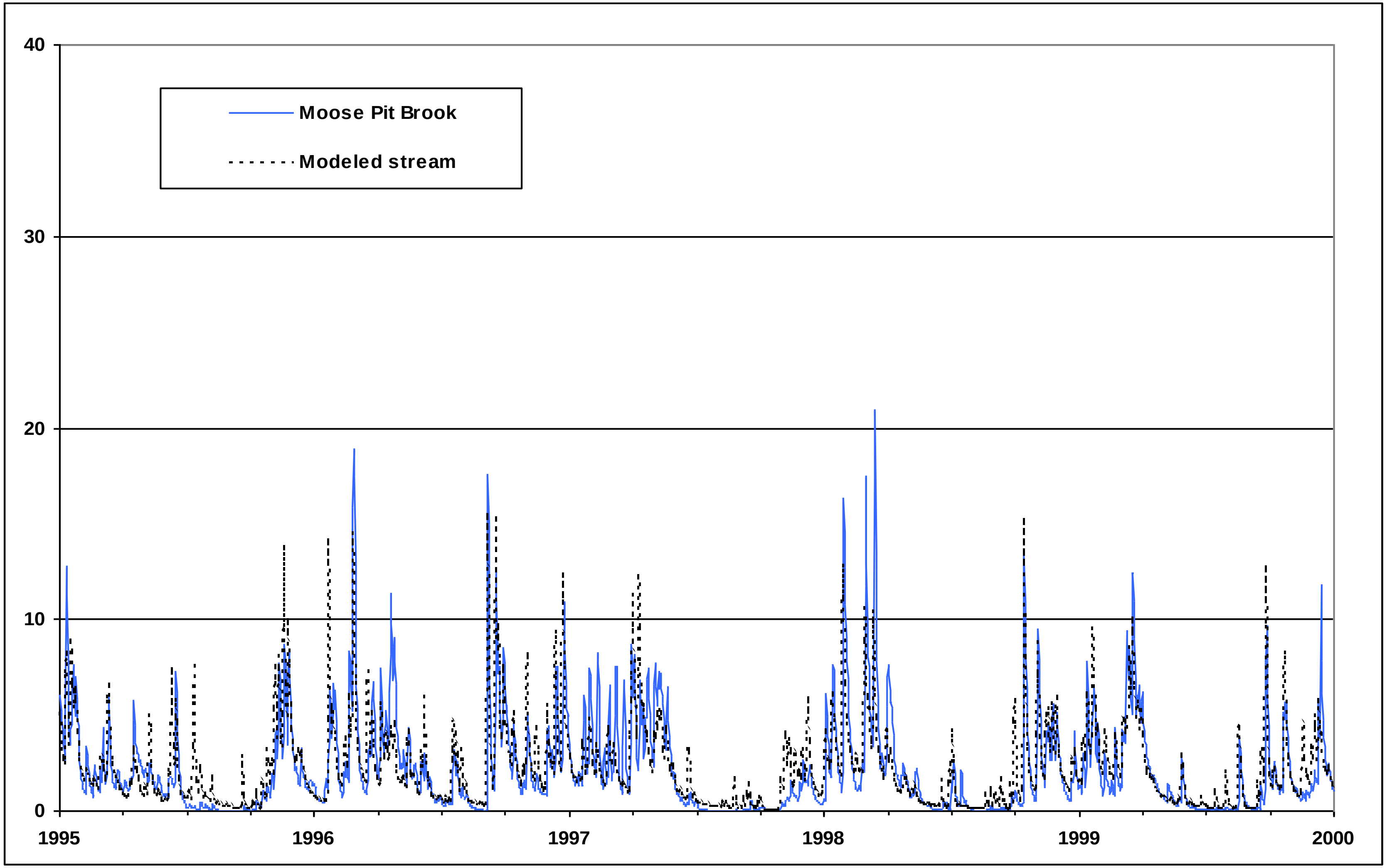

Fig. 43. Simulated and measured daily discharge (in mm day^{-1}) for Moosepit Brook from 1995 to 2000.

The parameter that controls the release of ions from the snowpack during snowmelt events ($K_{release}$) was found to give best results with a value of 0.3 for H^+ and 0.4 for other ions (NO_3^-, SO_4^{2-}, Ca^{2+}) after calibration with the snowmelt data for the open area (cleanest snow).

$$Q_{ionsleaving} = k_{release} * \frac{Water_{throughsnow}}{Water_{throughsnow} + Water_{in\ snow}} * Q_{ions\ in\ snowpack}$$

Table 8 shows the mean ion concentrations that were used to calculate overall ion deposition rates, for comparison. Figures 44 and 45 show the simulated and measured ion concentrations in snowmelt for the three sites. As explained in Chapter 4, high fluctuations in ion concentrations at the end of the snowmelt season are likely due to snow contaminations from rain splashing, and from plant and soil debris.

Table 8. Mean ion concentrations in precipitation

	Concentrations used in model	Concentrations in Fredericton from Simpson (1997)	Mean concentration in Kejimkujik (1985-2000)
pH	4.6	4.51	4.3
NO_3^-	0.02	0.016	0.032
SO_4^{2-}	0.015	0.017	0.041
Ca^{2+}	0.035	0.074	0.0064

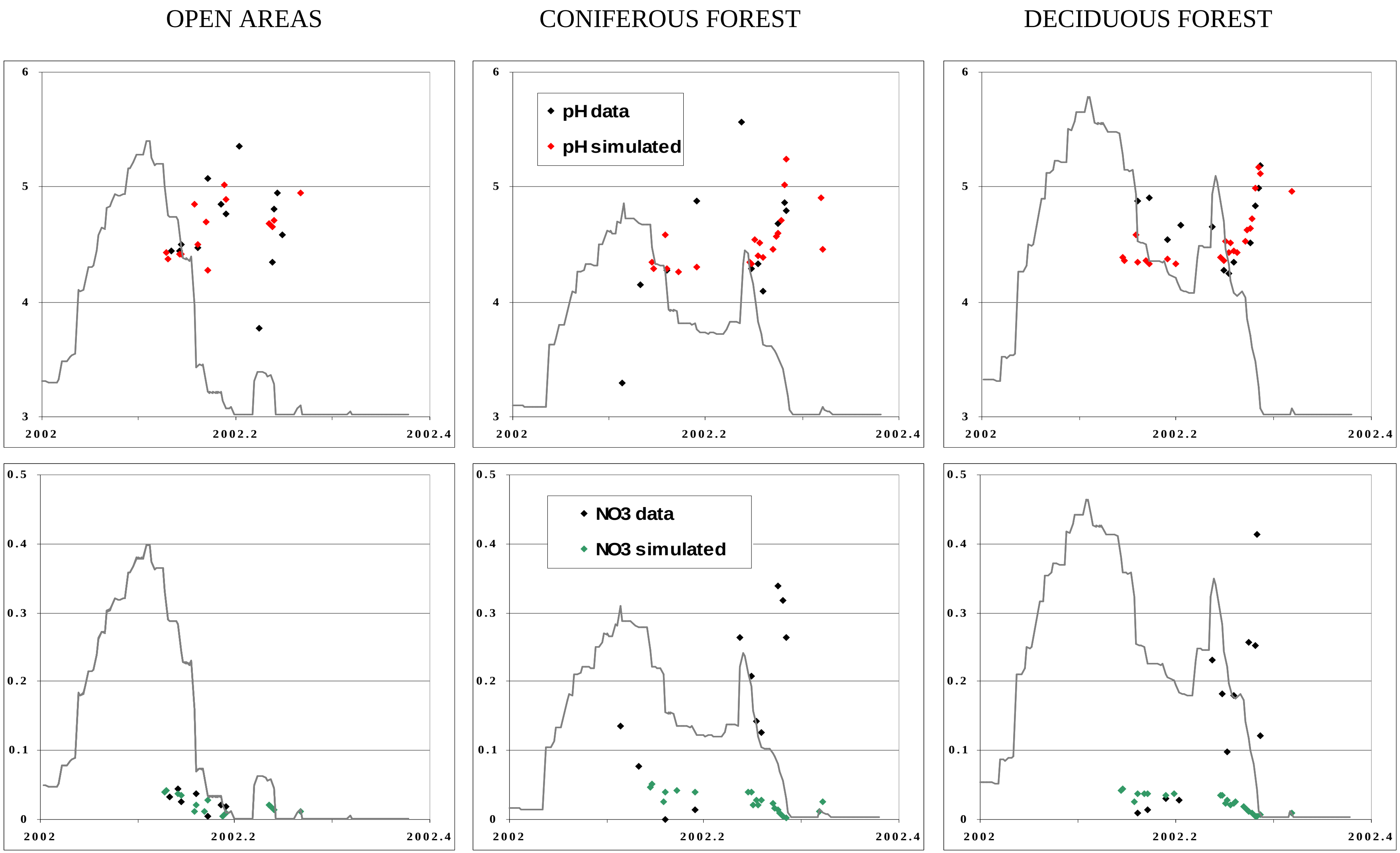

Fig. 44. Measured and simulated pH and NO_3^- content (in meq L^{-1}) of snowmelt in Fredericton.

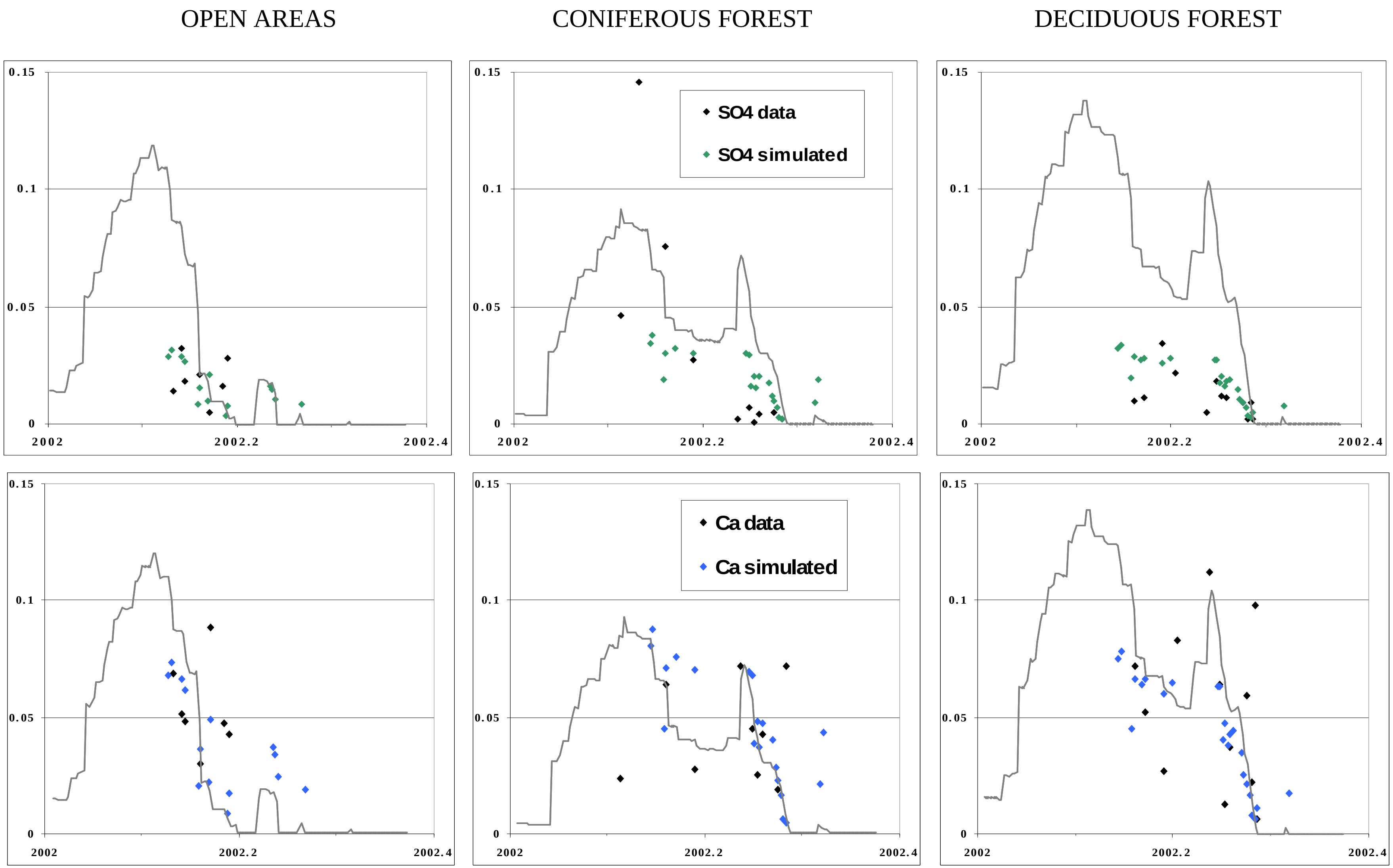

Fig. 45. Measured and simulated SO_4^{2-} and Ca^{2+} content (in meq L^{-1}) of snowmelt in Fredericton.

8.5 REFERENCES

Palmer, D.C. 1988. Correction of snowfall measurements on two small watersheds in New Brunswick. M.Sc.F. thesis, University of New Brunswick, Fredericton, NB.

Simpson, B.N. 1997. Quantifying sustainability for twenty-five forest sites in New Brunswick. B.Sc.F. thesis, University of New Brunswick, Fredericton, NB.

Yanni, S. 1995. Hydrogeochemical assessment of water in forested watersheds at Kejimkujik National Park: discharge rate, chemical composition and ion fluxes. M.Sc.F. thesis, University of New Brunswick, Fredericton, NB.

CHAPTER 9

MODEL APPLICATION: ACID DEPOSITION, CLIMATE CHANGE.

9.1 INTRODUCTION

In this Chapter, the new ForHyM model is used to analyze likely impacts of climate change and change in atmospheric deposition on:

- stream discharge,
- snowpack accumulations and snowmelt, and
- ion and acid release rates from snowpacks during snowmelt

at Kejimkujik National Park, Nova Scotia. All of this is done to derive a basic understanding of how changes in climate and atmospheric deposition may affect the hydrological conditions on the ground at the Park, as expressed by timing and amounts of soil moisture and temperature and snowpack accumulations, and in the forest stream water, by timing and extent of stream discharge and acid pulses. These acid pulses are of concern in relation to the health of fish that over-winter in the streams at the Park and elsewhere (Laudon *et al.* 2002). Some work on acid pulses in these forest streams has already been done (Clair *et al.* 2001), to examine what relationships there may be between precipitation and frequency and magnitude of acid pulses in streams. With this Chapter, these examinations are continued:

- by comparing field-observed stream chemical composition with modeled snowmelt chemical composition,

- by simulating impacts of 20% and 40% decreases of atmospheric ion loads on snowmelt chemical composition,
- By simulating how mean air temperature variations of –2°C, +2°C and +4°C will affect snow depth, cumulative discharge and snowmelt.

Table 9 shows ion concentrations at KNP for atmospheric deposition and in streams.

Table 9. Comparison of mean ion concentrations for precipitation in Kejimkujik National Park and in Moosepit Brook, between Jan 1st 1984 to Dec 31st 2000 (in meq L^{-1})

	H^+	pH	Cl^-	NO_3^-	SO_4^{2-}	Ca^{2+}	Mg^{2+}
Precipitation	0.053	4.28	0.049	0.032	0.046	0.0064	0.011
Stream	0.021	4.69	0.1	0.0003	0.045	0.042	0.044
Ratio stream/prec.	0.39		2	0.0087	0.98	6.5	4.1
Intercepted	61.1%			99.1%			

9.2 SIMULATING ION CONCENTRATIONS IN SNOWMELT

The ForHyM model, as calibrated and discussed in the preceding Chapters, was used to simulate stream discharge and snowpack accumulations for Moosepit Brook, from 1985 to 2000. Also calculated were pH, Ca^{2+}, NO_3^- and SO_4^{2-} concentrations in the snowmelt based on wet deposition alone. The results so obtained were then compared with field-observed values for stream discharge, snowpack depth, precipitation and stream pH, and precipitation and stream ion concentrations (Ca^{2+}, NO_3^-, SO_4^{2-}) as shown on Figures 46 to 49.

As plotted, release of H+ from the snowpack is pulsed, i.e., occurs with each discrete snowmelt event. The fluctuations of pH within the snowmelt amount to about one pH unit. This is similar to the pH fluctuations in the precipitation. Stream pH is also pulsed, but varies little, i.e., acid pulses in streamwater amount to about 0.1 pH unit. This suggests that the stream water is much better buffered than snowmelt. Much of the pH variations in the snowmelt are likely absorbed as the snowmelt water percolates through the soils towards the stream. Also, high levels of dissolved organic matter in the essentially brown and tea-colored stream water would further add to acid buffering potential of the stream water. During late fall and early winter, pH in the stream is the lowest. At these times, the lower soil layers are calculated to be water-saturated due to increased water inputs into the soil and watersheds. Hence, more water is calculated to flow laterally through the upper portions of the soils. In this case, water that percolates through the basin from the upper forest soil layers would contain high amounts of dissolved organic acids, thereby lowering streamwater pH to the lowest annual values. Overall, annual stream water pH fluctuates from as low as 4.2 to 5.0 or higher.

With regard to NO_3^-, it is clear that there must be considerable NO_3^-, absorption or nitrate loss as water percolates from the melting snowpack to the streams. The calculations suggest that some of the sulphate ions are also absorbed. However, the model should increase sulphate and nitrate inputs into the snow by at least a factor of 1.25 or more on account of dry deposition (Yanni 1995). With regard to Ca^{2+}, it appears that the stream has much higher Ca^{2+} concentrations than the snowmelt. Here, water is picking up Ca^{2+} ions as it percolates through the soil (and other ions such as K^+, Mg^{2+}, and Na^+) on account of soil weathering. Seasonally, snowmelt water does not appear to affect stream pH, because stream pH increases gradually from its lowest values in late fall to its highest values the following year in early fall. Sulphate levels appear least in the summer, and highest at the end of each year (beginning of winter). Calcium shows the greatest seasonal variations, being lowest in the spring, and highest in the fall. The latter is likely due to soil weathering, because soil temperatures would be least in the spring and highest in the fall. Sulphate ions are likely controlled biologically, showing highest uptake in the summer, and greatest release during winter. Stream pH follows a trend similar to Ca^{2+}, and should be highest when the rate of soil weathering is the highest.

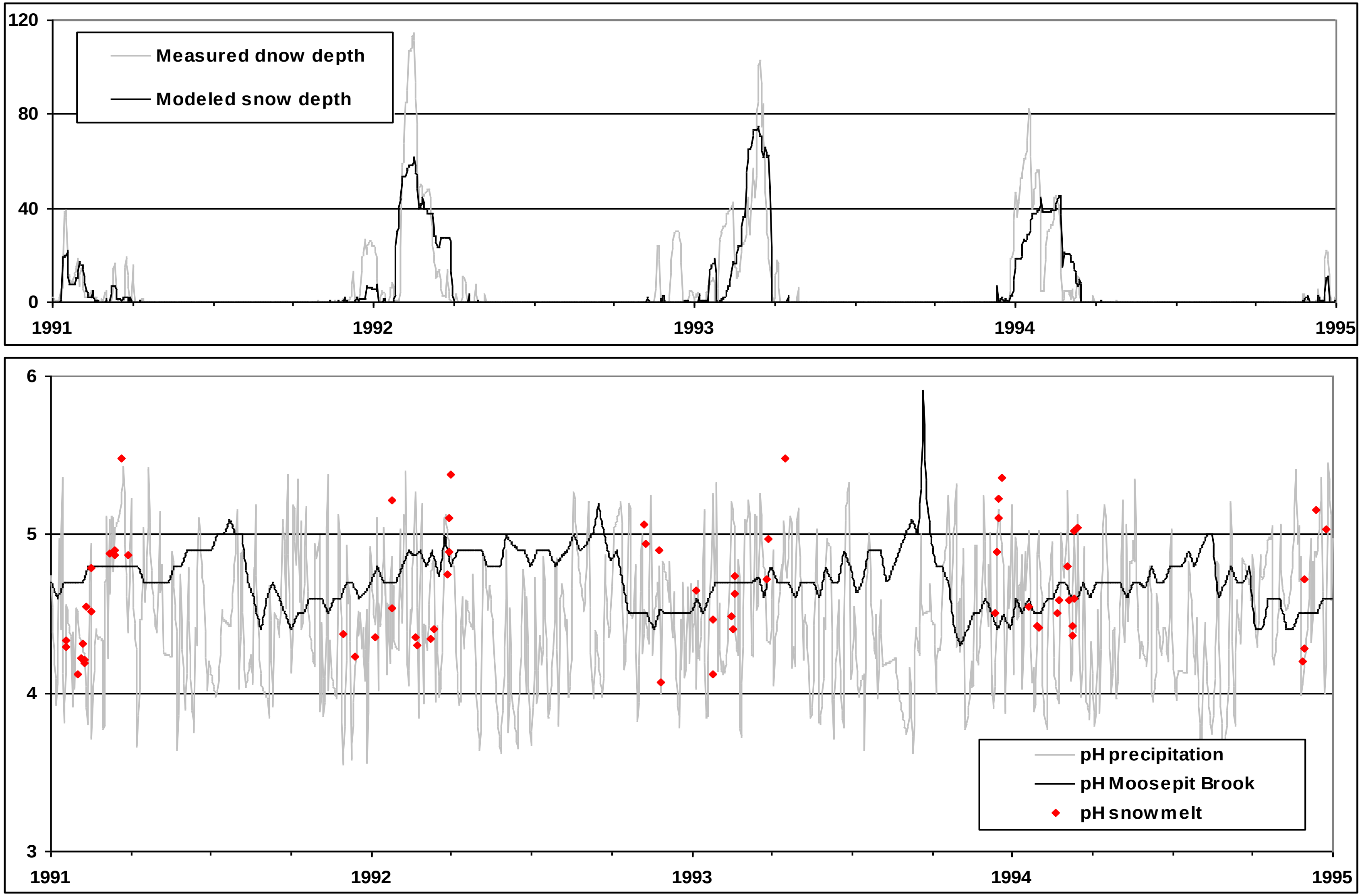

Fig. 46. Comparison of modeled pH of snowmelt and pH of Moosepit Brook.

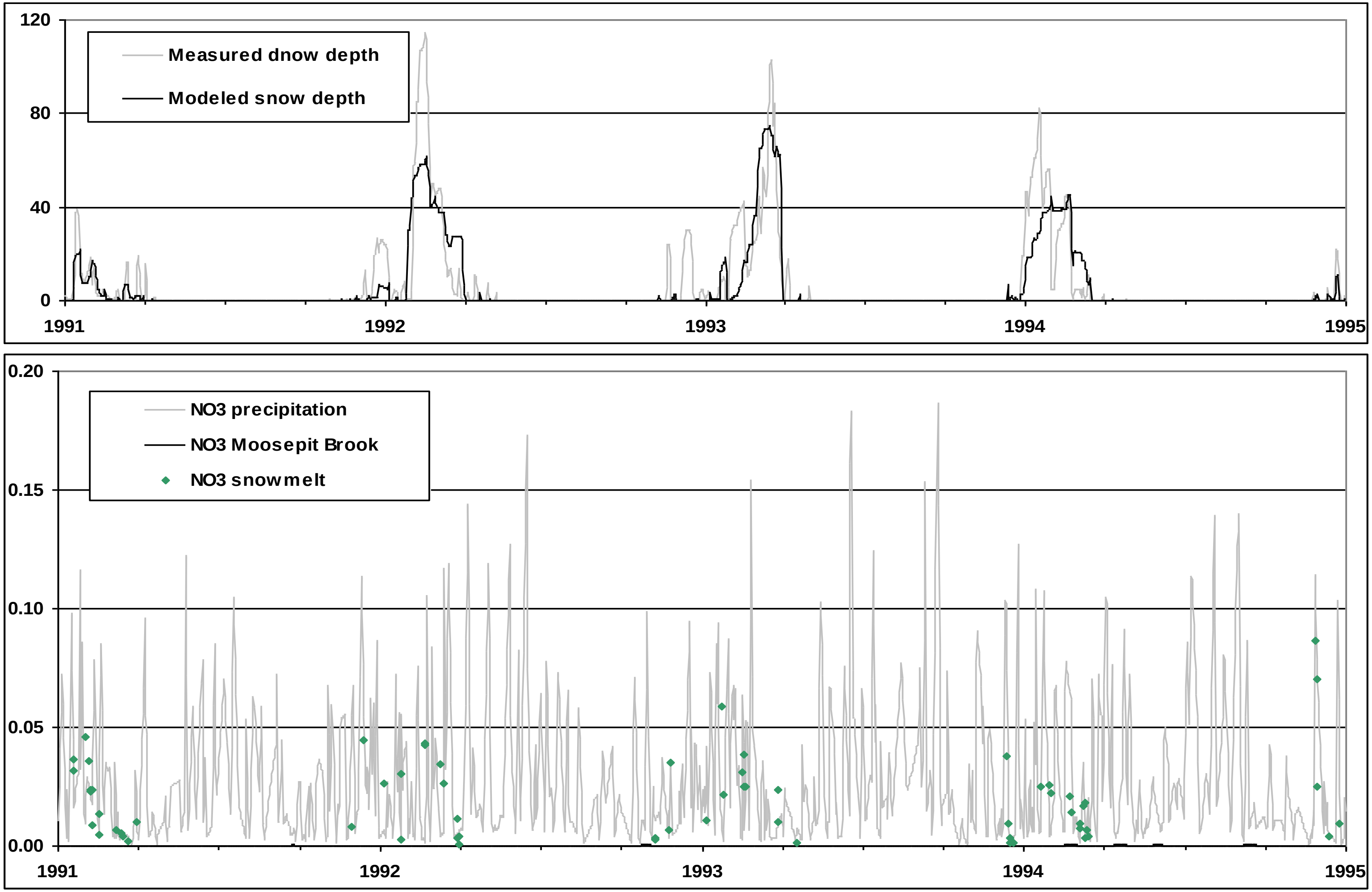

Fig. 47. Comparison of nitrate concentrations in modeled snowmelt and in Moosepit Brook (close to zero) (in meq L^{-1}).

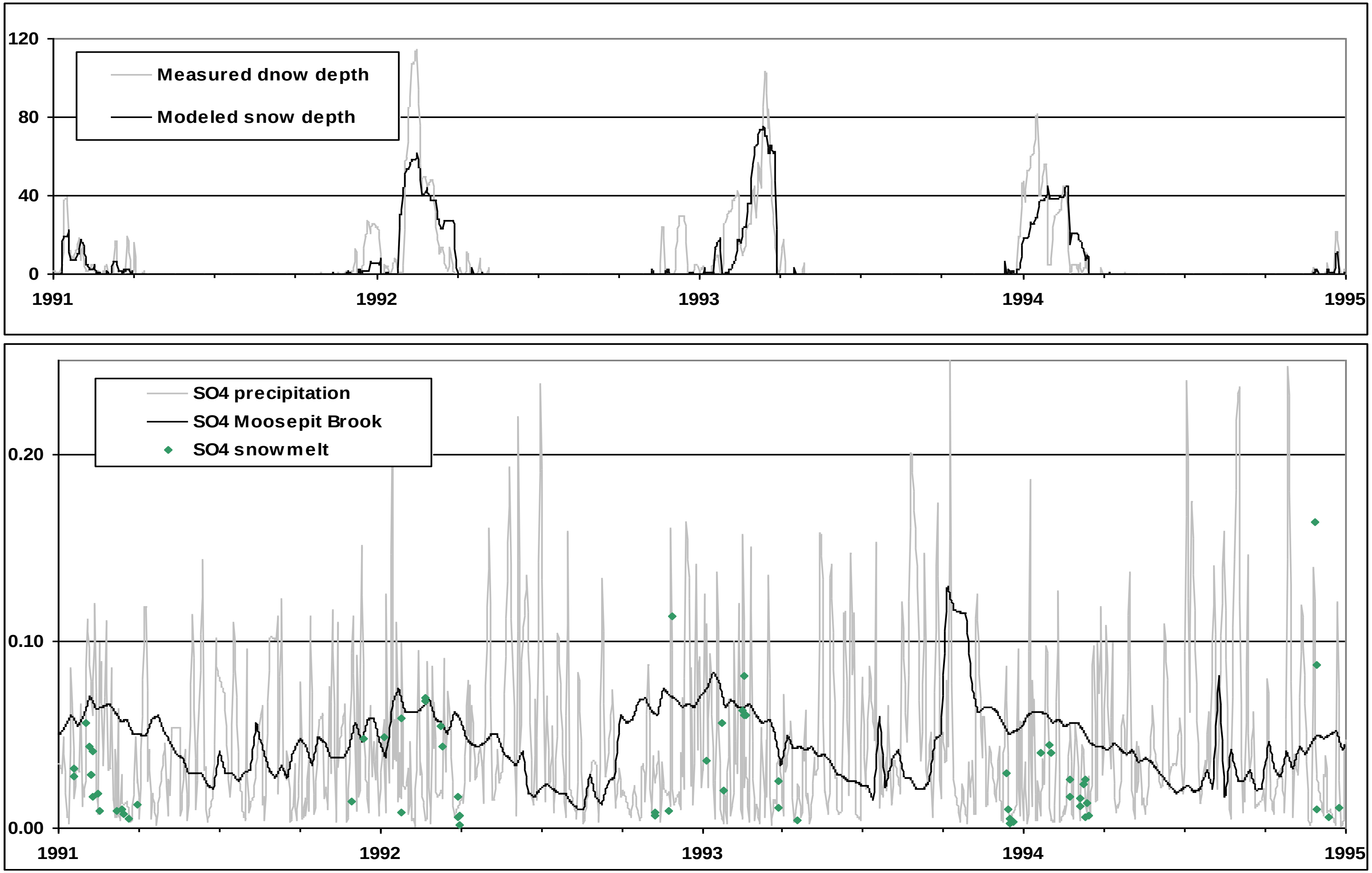

Fig. 48. Comparison of sulphate concentrations in modeled snowmelt and in Moosepit Brook (in meq L^{-1}).

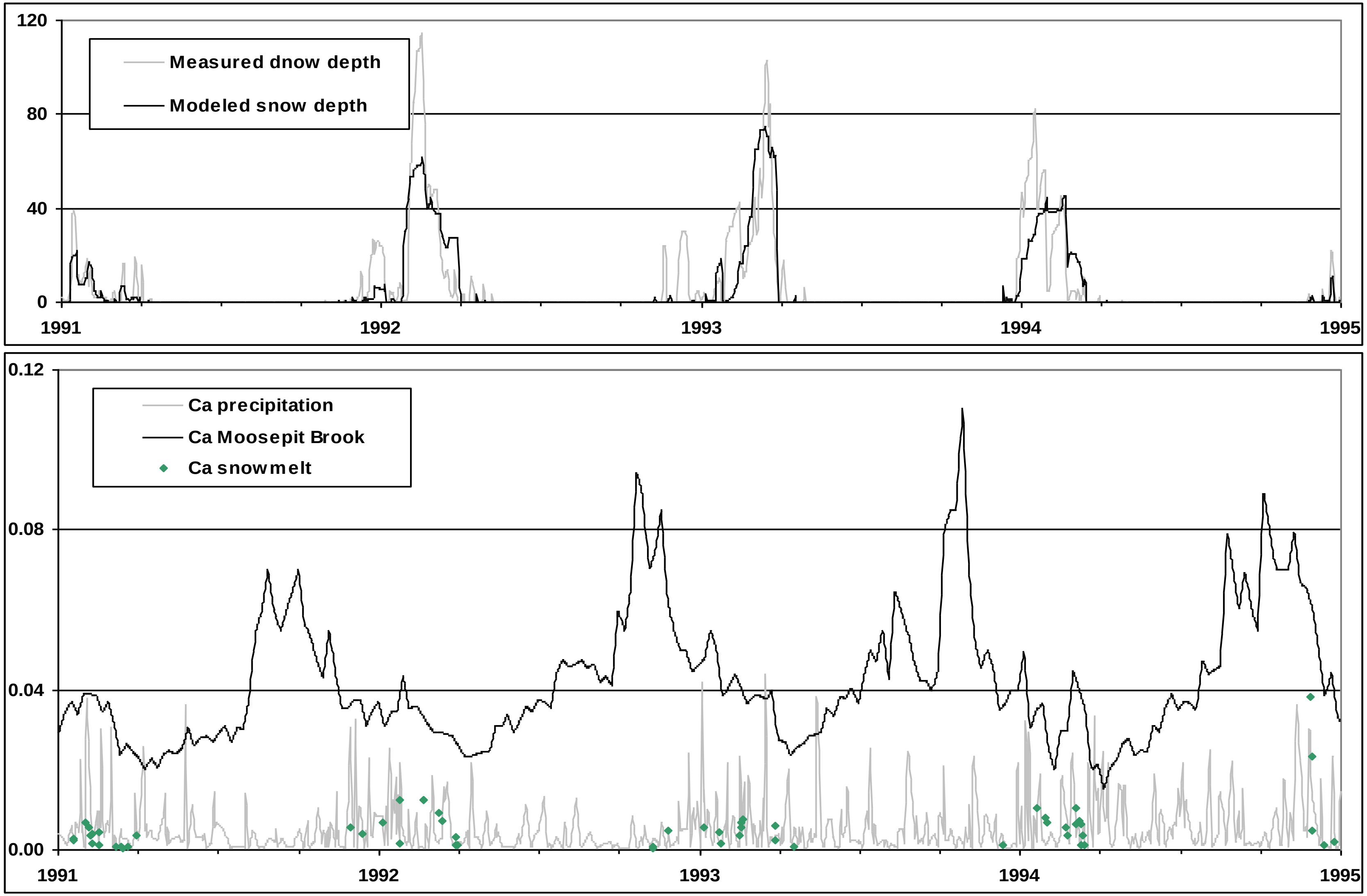

Fig. 49. Comparison of calcium concentrations in modeled snowmelt and in Moosepit Brook (in meq L^{-1}).

9.3 SIMULATING IMPACTS OF DECREASED ACID DEPOSITION ON ION CONCENTRATIONS IN SNOWMELT

In Nova Scotia, acid deposition has decreased by approximately 50% since 1983 (Laudon *et al.* 2002). However, acid pulses in stream water still occur at times of large rainfall events in the fall, and during extensive snowmelt events in winter and early spring. It is thought that the acidity of the precipitation is transmitted to the stream water because the soils of the area are shallow, and – in principle - should have little acid buffering capacity (Laudon *et al.* 2002).

The following calculations are done to see to what extent reductions in atmospheric deposition would lead to reductions in snow melt acid loads, and to relate the resulting concentrations for H^+, sulfate and nitrate ions to the corresponding stream water concentrations. The simulations were done using daily snow, rain, temperature, and ion deposition data from Kejimkujik National Park for the period 1985-2000. Ion concentrations were decreased by 20% and 40% as compared to the actual data.

Over the fifteen years studied, 288 snowmelt events occurred, among which 121 were characterized by a pH smaller than 4.5, which is fairly low (mean pH of the stream is 4.7 as shown in Table 9). This number dropped to 98 events with 20% reduction, and to 49 events with 40% reduction. The most acidic event had a pH of 3.68; this would be improved to pH 3.77 and 3.9, and the median value of snowmelt events would increase from 4.55 to 4.65 and 4.77 with 20 and 40% acid deposition reductions, respectively. Table 10 shows a summary of the model output for mean pH.

Table 10. Characteristics of pH of modeled snowmelt events between Jan 1st 1985 to Dec 31st 1999

	pH normal deposition	pH -20% deposition	pH -40% deposition
pH Minimum	3.68	3.77	3.9
pH Maximum	5.74	5.83	5.96
pH Median	4.55	4.65	4.77
Number of events with pH < 4.5	121	98	49

Figure 50 shows the frequency distribution of pH values in all snowmelt events over the 1985-2000 period. Altogether, the calculations imply a systematic shift towards higher pH values with increased reductions in acid deposition.

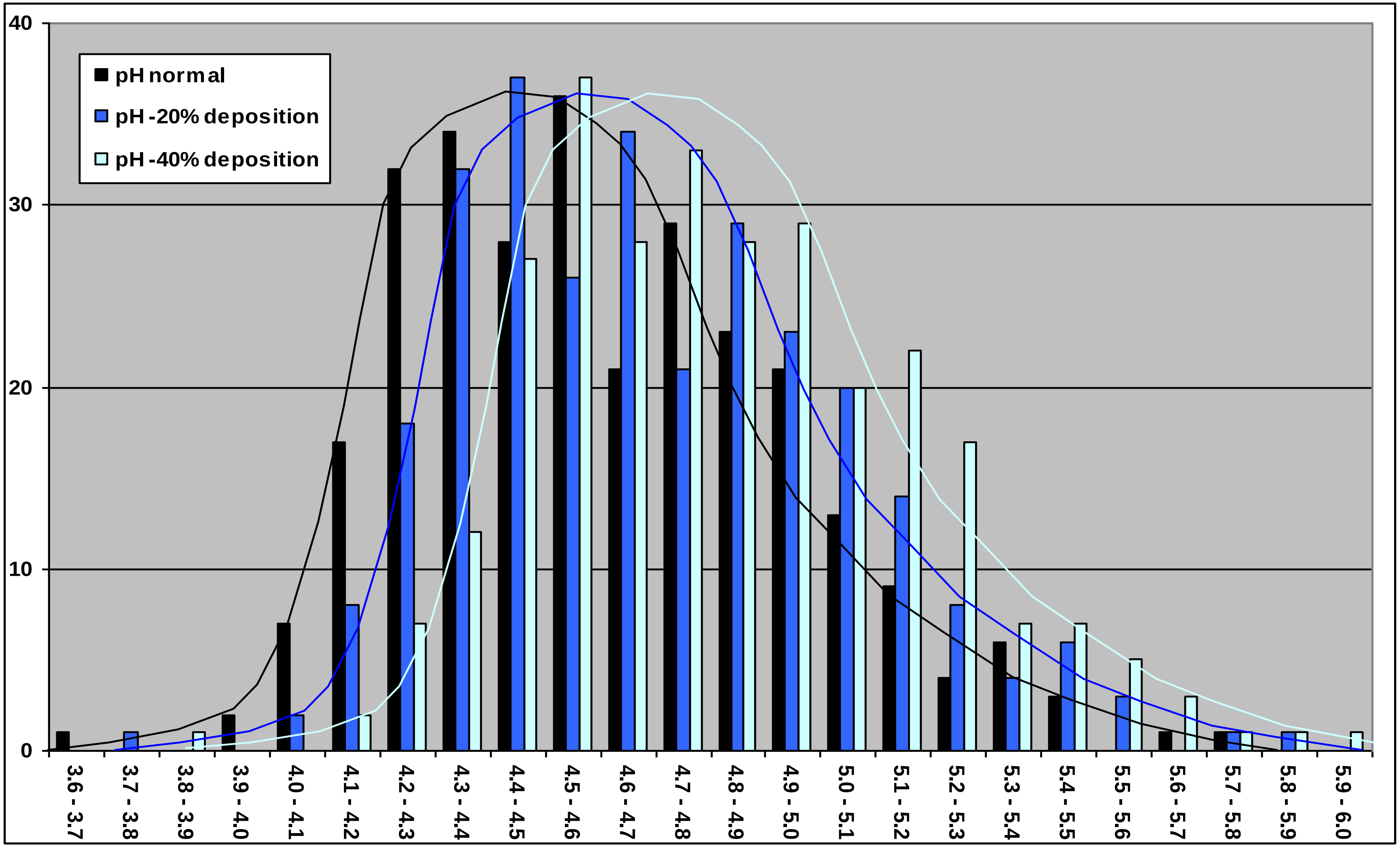

Fig. 50. Number of acidic snowmelt events simulated with actual data and under hypotheses of decreased deposition, between Jan 1st 1985 and Dec 31st 1999.

9.4 SIMULATING IMPACTS OF CLIMATE CHANGE ON SNOWMELT, STREAM DISCHARGE AND ION CONCENTRATIONS IN SNOWMELT

Due to increasing emissions of gases like CO_2, CH_4 and N_2O into the atmosphere, global circulation models have predicted a gradual increase in global air temperatures ranging from 1.4 to 5.8°C by 2100 as compared to current levels (Intergovernmental Panel on Climate Change 2001). This will affect the nature of precipitation (less snow and more rain), and precipitation events will likely become be more variable as well. The hydrologic cycle will also be affected, with dryer summers expected in mid-continental areas. In Canada, winter snowpacks will melt earlier and spring floods will likely be of smaller magnitude because of more frequent snowmelt events throughout the winter (Clair *et al.* 1998). To visualize how changes in air temperature may affect model outcome, it was decided to run the model for four climate change scenarios, by changing current air temperatures through simple addition of –2, 0, +2 and +4°C on top of the historical air temperature record from 1985 to 2000. Precipitation values were kept unchanged, except for ensuring rain/snow switch occurs between –1 and +1°Cin each run. For each simulation it was necessary to run the model twice, to correctly re-initialize the soil temperatures for each run.

With increased air temperature, there should be fewer snowfalls and more rainfalls, which should lead to significantly smaller snowpacks, as illustrated on Figure 51, Snowpack also melts quicker due to warmer air and warmer and more frequent rain. Table 11 presents the mean number of days per year with snow coverage, as calculated with the model.

Table 11. Mean number of days per year with snow on the ground in Moosepit Brook watershed predicted by the model between Jan 1st 1985 to Dec 31st 1999

Simulated -2°C	Simulated normal	Simulated +2°C	Simulated +4°C
131	106	85	64

Calculated cumulative discharge should diminish with increasing air temperature due to greater evapotranspiration, as is shown on Figure 52. Figure 53 details the quantitative difference in cumulative discharge between the four air temperature runs: over the long term, the difference is quite linear, and is equal to a loss of about 50mm per year per degree of increase in air temperature. Also, with increasing air temperature, peak discharge events should occur earlier in the spring, due to earlier melting of the snow on the ground.

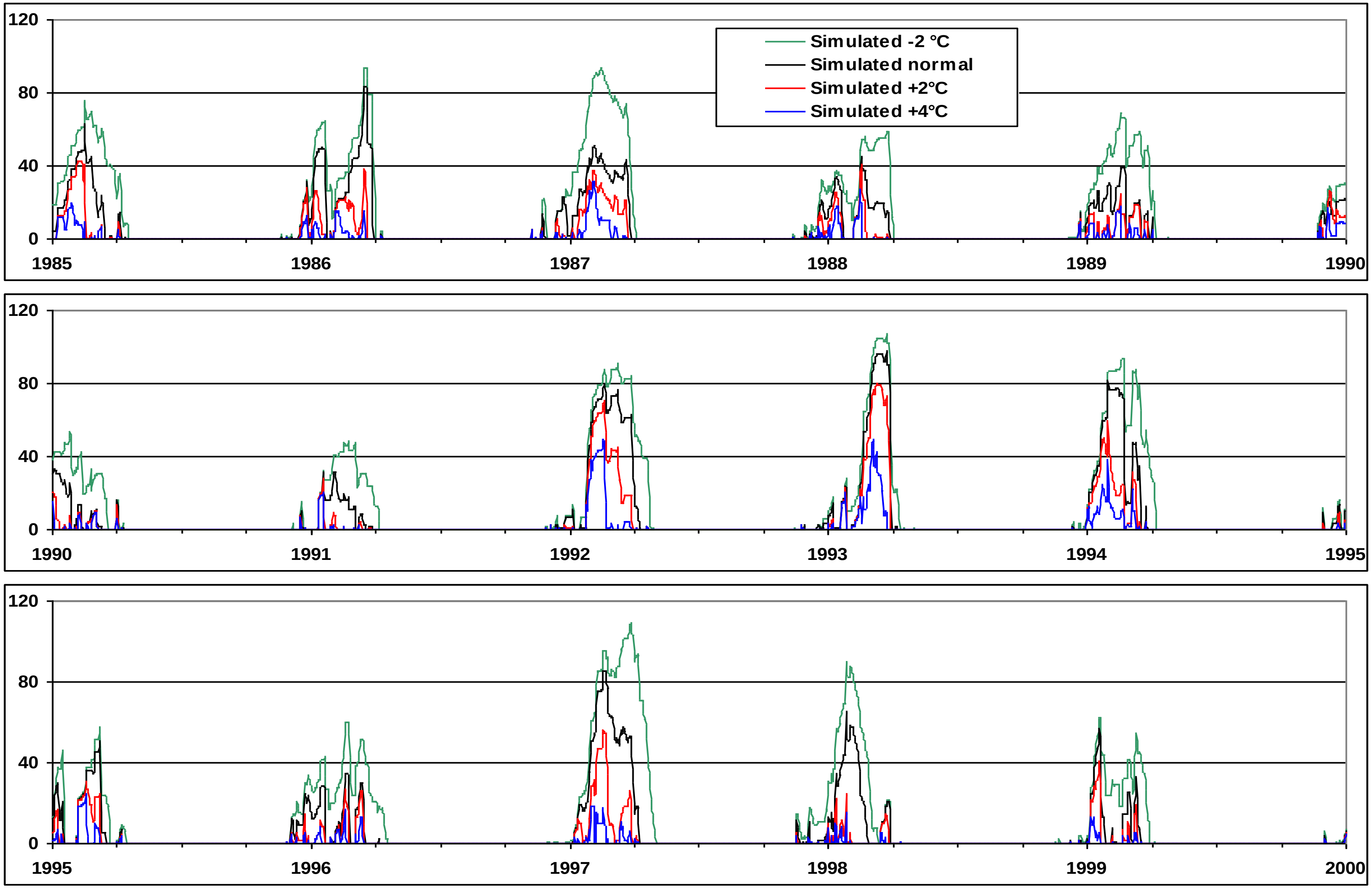

Fig. 51. Climate change simulations for snowpack (in cm) in Moosepit Brook watershed.

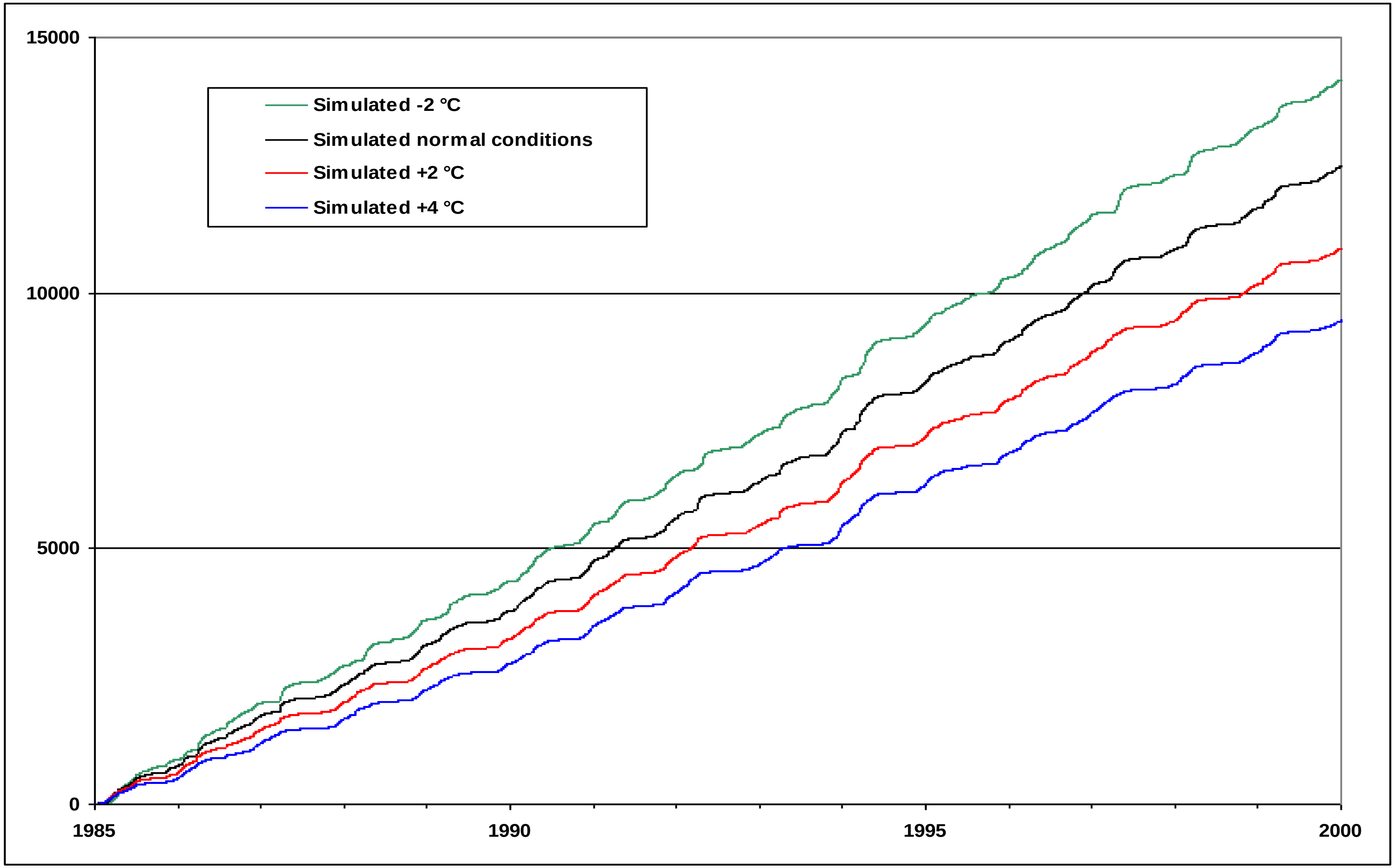

Fig. 52. Evolution of cumulative simulated discharge (in mm) with climate change.

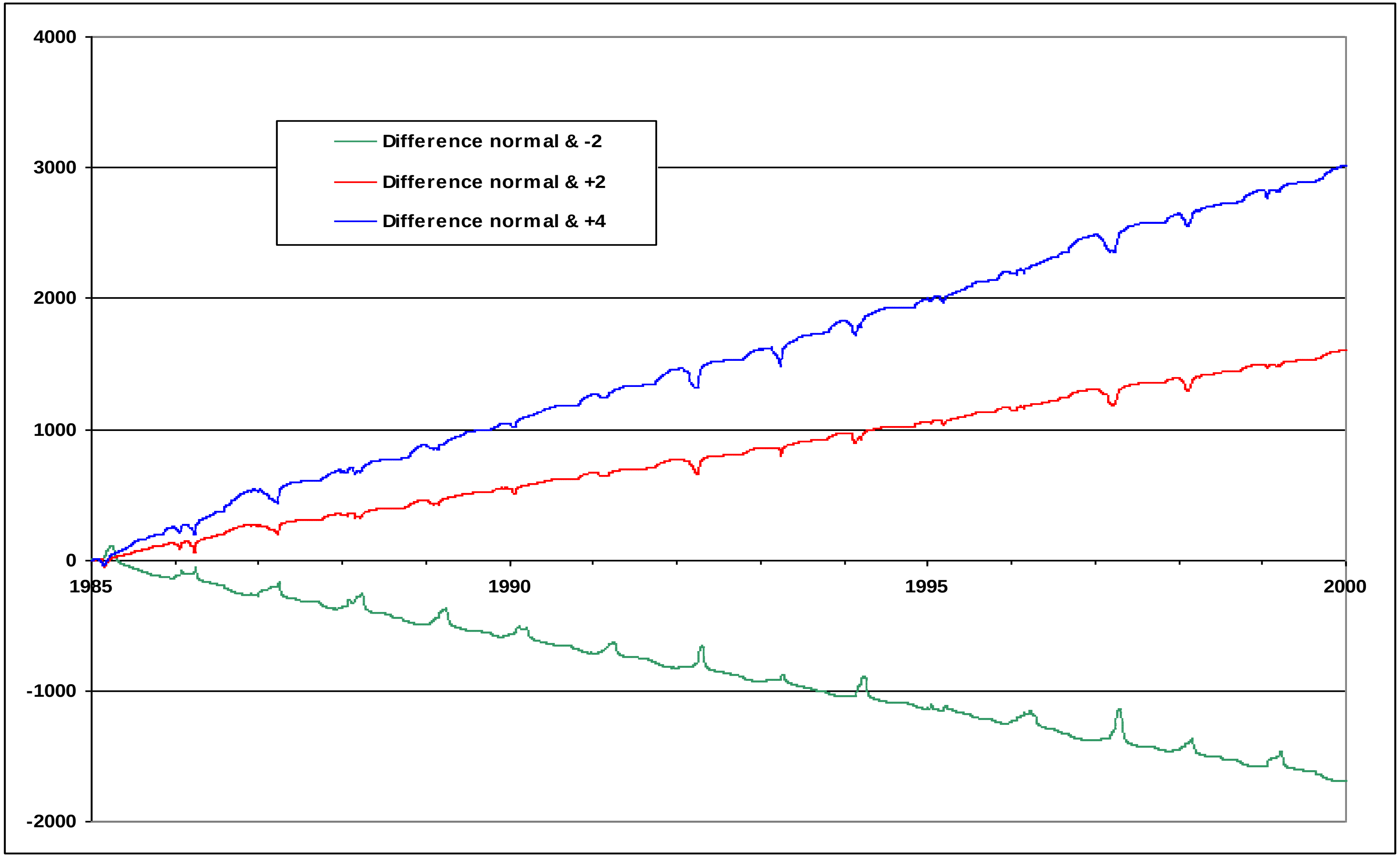

Fig. 53. Cumulative difference in discharge (in mm) between current conditions and hypothetic conditions.

Daily discharge plotted on Figure 54 shows the expected consequences of warmer temperatures. For the +4°C run, base flow would clearly be lower in the summer, but higher in the winter due to more frequent snowmelt events. Peaks in discharge should be smaller in the fall due to more evapotranspiration during the summer, and discharge peaks should become more frequent in the winter because of reduced snowpack accumulations.

A summary about model output concerning pH in snowmelt events is presented in Table 12 for each of the 4 runs. The median value should gradually increase with rising air temperature. These results can be compared with the pH distribution plotted on Figure 55.

Table 12. Characteristics of pH of modeled snowmelt events between Jan 1st 1985 to Dec 31st 1999, with normal conditions and with climate change conditions

	-2°C hypothesis	Normal	+2°C hypothesis	+4°C hypothesis
pH Min	3.85	3.86	3.9	3.98
pH Max	5.8	5.92	5.73	5.87
pH Median	4.44	4.53	4.64	4.74
Number of events with pH < 4.5	222	138	85	60
Total number of events	359	292	265	253

Figure 55 details how the frequency distribution of pH of the snowmelt events should vary with increasing climate warming: in general, higher air temperatures in winter should lead to higher snowmelt pH. In addition, increased winter air temperatures should strongly diminish the occurrence of low pH snowmelt events. In other words, warm winters will have fewer episodes of severe acid pulse events than cold winters.

Figure 56 shows the daily model output for a four-year period. As indicated, the timing of peaks in snowmelt ion concentrations should also change. For example, the simulations with -2°C produce the largest snowpacks in each year, and melting would be delayed and quite sudden in the spring. Also, snowmelt ion concentrations would be quite high in early snowmelt runoff, due to the large ion loads within the snowpack. In contrast, the simulations with +4°C should lead to much reduced snow accumulations on the ground, to early snow melt events all along the winter. Also, increased snowmelt occurrences should deplete the ion loads of the snowpacks more quickly each time. Therefore, concentrations in snowmelt would be even more strongly influenced by daily deposition, especially from rain, and rain – as calculated - should become more frequent than snow. Since concentrations of ions in precipitation are very variable, snowmelt concentrations should follow the same trend.

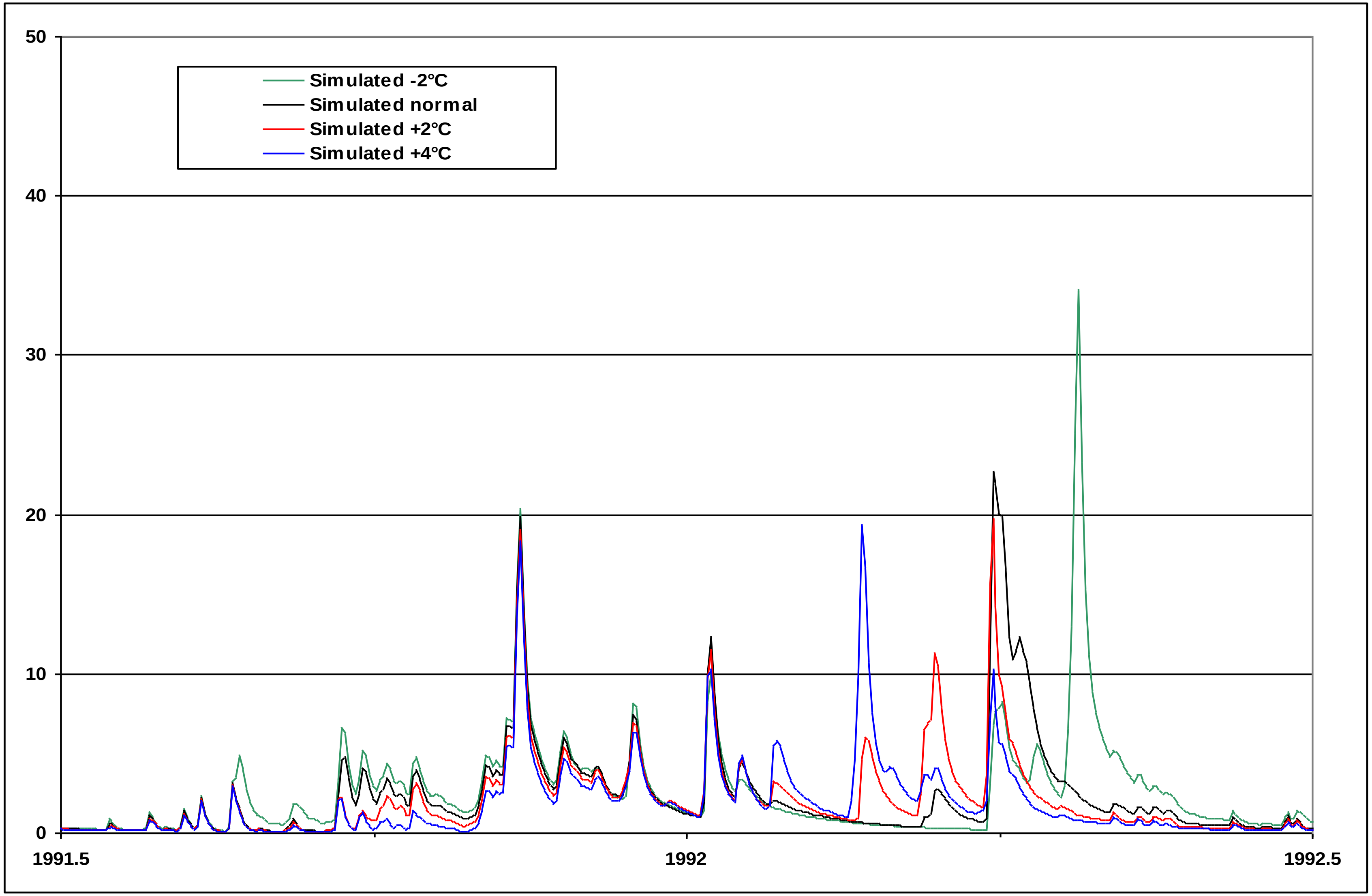

Fig. 54. Simulated daily stream discharge (in mm day^{-1}) for various climatic conditions.

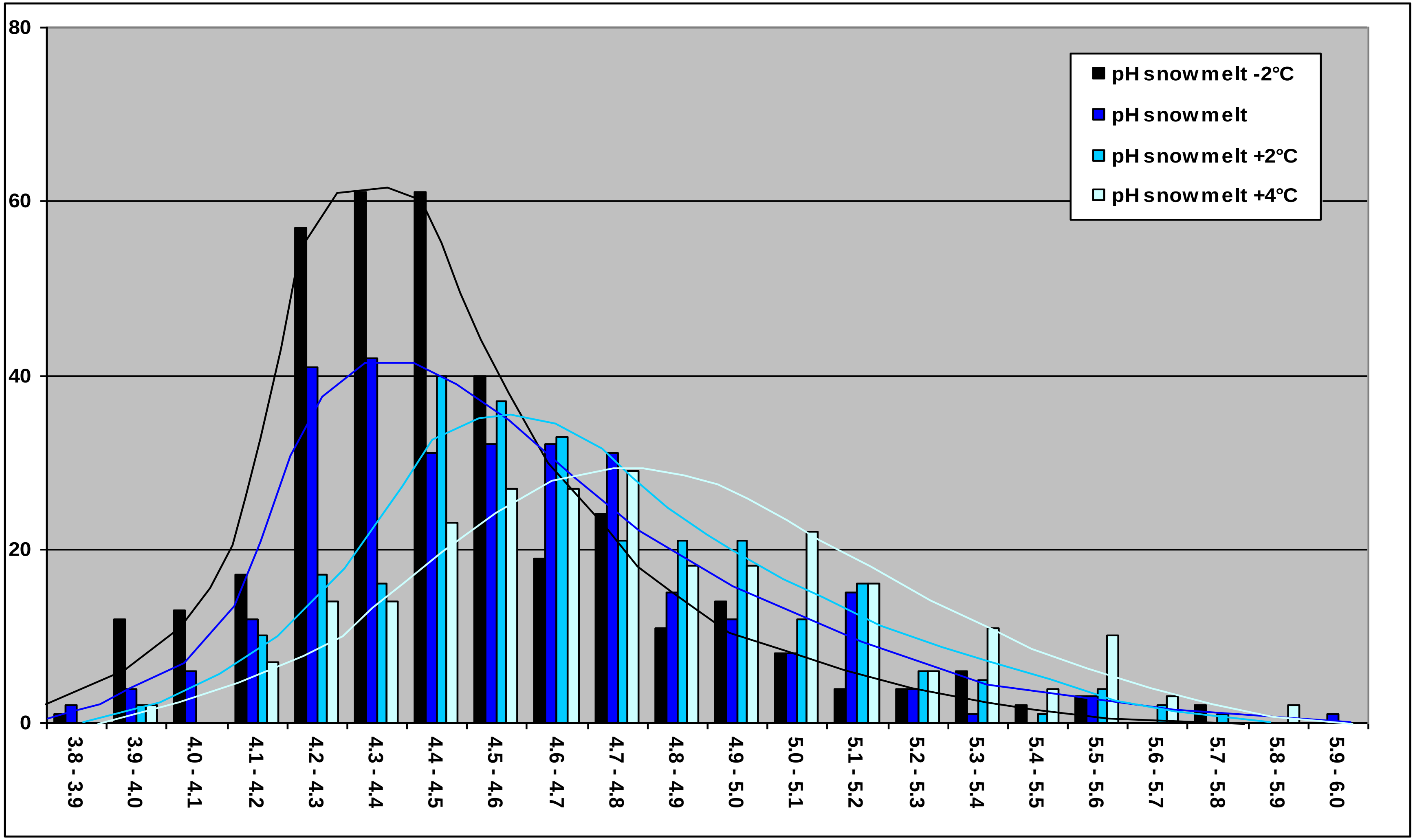

Fig. 55. Distribution of pH of snowmelt events simulated with measured deposition in normal conditions and under hypotheses of air temperature change, between Jan 1st 1985 and Dec 31st 1999.

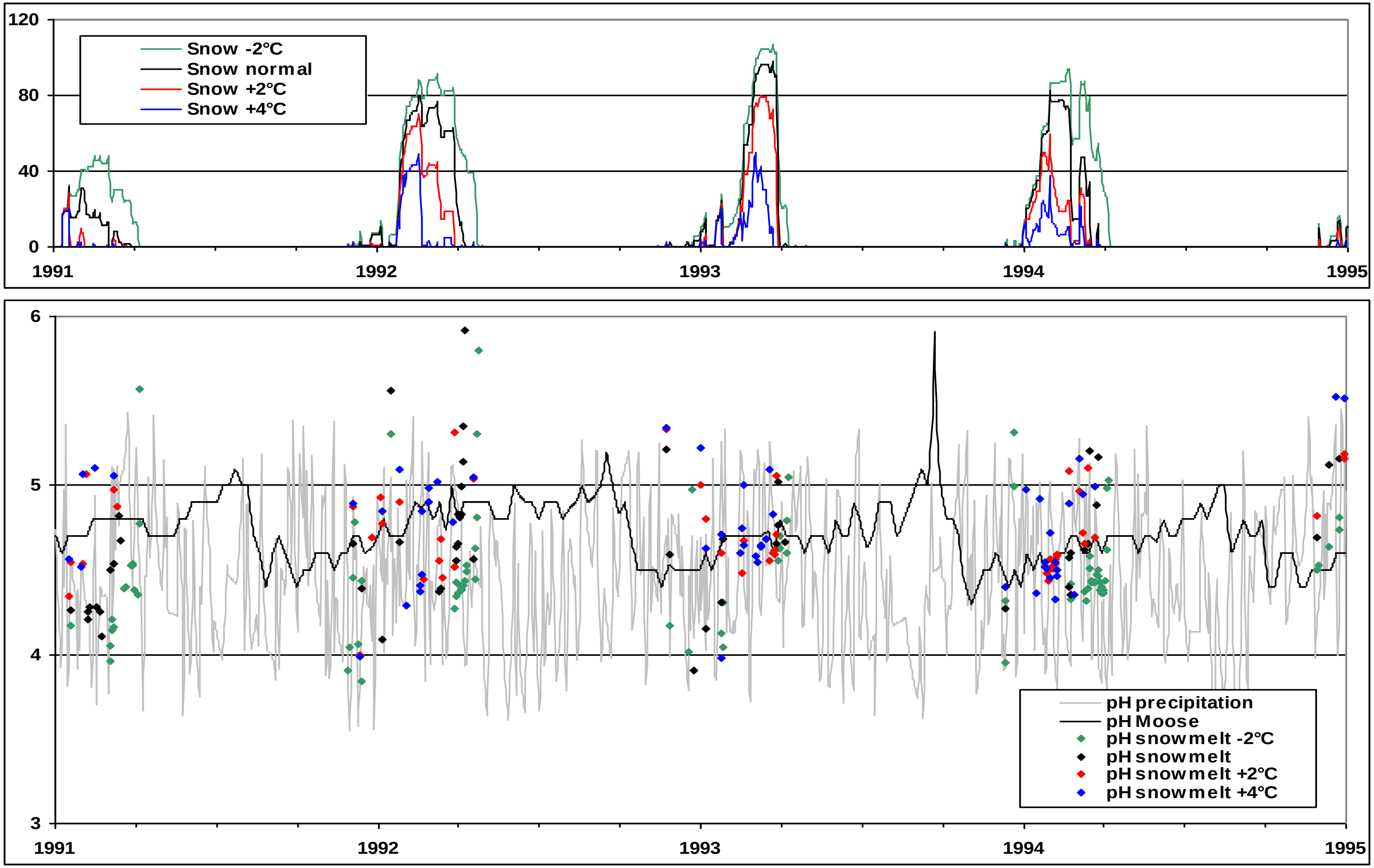

Fig. 56. Evolution of pH of snowmelt (in meq L^{-1}) with climate change.

9.5 REFERENCES

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CHAPTER 10

CONCLUDING REMARKS

10.1 THESIS SUMMARY

The following summarizes the major components of this Thesis:

1. A literature review was conducted to provide background and hydrological concepts for the field and modeling research of this Thesis, with special reference to physical and chemical properties and transformations of snow within snowpacks, including processes that affect acid and ion release from snowpack during successive snowmelt events (Chapters 2 and 3).

2. Field research was conducted to establish data for snowpack depth, density, temperature, pH and ion loads of snowpacks and snowpack-generated snowmelt for three forest conditions (softwood, hardwood, open). The data so generated were obtained from carefully selected mini-catchments that take advantage of the natural mound-and-pit micro-topography that is generally associated with forest soils (Chapter 4).

3. The process of modeling the flow of water and heat through a forest stand and/or forest catchment was introduced in Chapter 5 by way of an overview of the forest hydrology model ForHyM.

4. A statistical analysis was done to reliably link soil hydrologic parameters such as permeability, density, porosity, field capacity and permanent wilting point to basic soil characteristics such as texture, organic matter content and soil depth (Chapter 6).

5. In Chapter 7, more details are provided about ForHyM in terms of the latest revisions arising from the research for this Thesis, especially in reference to
   - a new formulation for snowmelt, and for ice formation in snow and soil,
   - re-parametrization of all parameters that are needed to model the flow and retention of heat and water through snow and soil, in reference to snow as a 3-component system (ice, water, air) and soil as a 5-component system (mineral, organic matter, water, ice, air),
   - connection of the flow of heat with the flow of water, especially during times of phase change of water into ice and vice versa,
   - the building of a user-friendly model structure, and building a new interface to enhance the process of model calibration and model application.

6. Details concerning model calibration regarding snowmelt and stream discharge are provided in Chapter 8. The emphasis in this calibration is making the calibration values for parameters pertaining to water retention in snow, heat transfer between atmosphere and snow, snow compaction, and ion release from snowpack as portable as possible, i.e., portable from softwood to hardwood to open conditions, at a scale ranging from small mini-catchments to a large watershed such as the Moosepit Brook catchment in Nova Scotia. In the process of model calibration, it was discovered that H+ ions would be released from the accumulated snowpack impulses, with essentially one pulse per

snowmelt event. It was also found that the variations in snowmelt pH would be about 10 times larger than the corresponding variations in streamwater pH.

7. In Chapter 9, the calibrated model was applied to evaluate effects of climate change (change in air temperature from –2 to +4°C) and reductions in atmospheric deposition (from 0 to 40%) of snowpack, snowmelt, ion release from snow, and stream discharge behavior at the locations of our field studies in Fredericton (UNB Forest) and Nova Scotia (Moosepit Brook).

10.2 ORIGINAL CONTRIBUTIONS

1. Data collection and analysis showed that the pH of snowmelt is generally lower than the pH of the total snowpack itself. This supports earlier observations and theories that most ions leave the snowpack during early snowmelt. A mechanism was introduced into ForHyM to model the release of ions from the snow. The mechanism is related to, but also differs from the model-internal mechanism that quantifies the amount of snowmelt that would occur given certain combinations of air temperature, soil temperature, snowpack temperature, and amount of rain received.

2. Snowpack density measurements showed a gradual increase of snow density during the winter, from 0.15 g cm^{-3} for fresh snow to values around 0.5 g cm^{-3} for ripe snow in the spring. This change can now be modeled, with ForHyM.

3. The reliability and portability of the ForHyM model was improved substantially by a thorough analysis of all formulations within the model. This included:
   - A numerically reliable generalization of all the pertinent parameters that are used to quantify the flow and retention of water and heat in snowpacks and in the soil, year-round, on a daily and possibly hourly scale.
   - Soil parameters such as saturation point, field capacity, permanent wilting point, pore space, bulk density, soil permeability at saturation now depend on sand, silt, clay and OM content as well as depth (instead of only clay before).
   - Heat flow and heat retention parameters such as heat capacity and heat conductivity now depend on all 5 soil constituents.
   - A number of adjustments made to account for daily and hourly variations regarding the quantification of incoming and outgoing radiation, based on slope, aspect, albedo, sun angle, altitude, etc. These calculations are essential for estimating the temperature at the ground surface be it the soil or forest floor in summer, or the snowpack in winter.

4. All the important equations in the model are now referenced to the literature or to this thesis.

5. Simulations of snowmelt chemical composition in response to decreased acid deposition (0, 20% and 40% decreases of H^+, NO_3^- and SO_4^{2-}) show that the decrease of the ion load in the snowmelt would be proportional to the deposition change.

6. The model application with respect to simulating the effects of decreased or increased air temperatures show that, with increasing air temperatures, there would – not surprisingly –

be less snow on the ground, more frequent snowmelts, etc. While these results are in line with general expectations, it is important to know that such changes can be quantified with ForHyM with unprecedented detail.

10.3 SUGGESTIONS FOR FURTHER WORK

1. *To model ion concentrations in soil solution and in stream water*. Doing this would require the modeling of chemical reactions that take place in the soil, and of nutrient cycling of ions like nitrates, sulphate, calcium, magnesium etc. This cycling occurs as vegetation absorbs nutrients from the soil, and dead leaves and branches fall on the ground and decompose. Moreover, this includes modeling soil weathering and soil leaching.

2. *To use ForHyM within the geo-spatial context of routing water from uplands to wetlands, by way of Geographic Information System (GIS) modeling software.* The current ForHyM model produces estimates for vertical soil percolation, and for lateral run-off and interflow as well. It also provides a means to determine fluctuations in water table as these would vary with weather. From a simplest perspective, one would start by dividing drainage basin into areas of homogeneous slope, aspect, forest cover and soil composition. Lateral flows from one surface unit would then be directed to flow into the next surface unit below.

3. *To link ForHyM with the output of global circulation models (GCM).* Global circulation models provide predictions for air temperatures and precipitation, at various geo-spatial scales (e.g., a 50 km grid) and temporal scales (e.g., monthly, or daily). ForHyM would turn GCM air temperature and precipitation output at each grid point into detailed

calculations regarding (e.g.): soil moisture and temperature, snowpack accumulations, snowmelt, evapotranspiration, ice in the ground, year-round.

4. Similarly, with ForHyM, historical weather data can be used to calculate, e.g., the past occurrences of soil frost and soil moisture deficits, to evaluate past occurrences of forest decline and/or crop failure, in relation to past local climate. In addition, ForHyM should find considerable use in the evaluation of soil chemical processes that are, for the most part, dependent on soil moisture and soil temperature conditions, e.g., the decomposition of soil organic matter, the weathering of soil minerals, etc.

5. ForHyM could find many uses in the scientific assessment of sustainable forest management by means of criteria and indicators. For example, how much of a particular watershed could be harvested without risking deleterious effects on water tables and flooding on downslope locations.

6. ForHyM could also find many field applications regarding decision-making with respect to scheduling outdoors activities, especially on terrain that gets wet and soft based on certain weather and soil conditions. This would be of special interest in forestry and agriculture, to minimize soil, compaction, rutting and soil erosion.

7. Much needs to be done in terms of examining snowpack, snowmelts and snowpack ion loads across regions, to determine whether the current model calibration remains applicable elsewhere, or to what extent which parameters need to be changed to account for variations not yet captured with the current model formulation, or with the rather limited set of data that was used for model calibrations. It would be interesting to

systematically examine snowpacks and snowmelt on ridge tops, in depressions, and on northern and southern slopes (same angle, between, e.g., 10 to 20 degrees). In this, it would be good to get hourly reading for air temperature and precipitation on the spot, to specifically deal with the hourly application of ForHyM.

8. Experiments could be done to determine snow albedo based on snowpack aging and quantity of dust on the snow.

9. Detailed examinations should be done to determined actual and modeled relationships between soil moisture percolation, and related fluctuations in water tables, based on local hydro-geological substrate conditions.

10. In terms of ForHyM, there are several issues that one may wish to address:

    - *How would model output (stream discharge, actual evaporation) be affected by the evapotranspiration model?* In its current formulation, ForHyM uses air temperature as the only predictor for assessing potential evapotranspiration. In other models, other weather indicators such as wind speed, relative humidity, incoming radiation, or net radiation are part of the model input.
    - *Should the soil be divided into more layers?* If so, what would be the overall gain of doing so?
    - *What needs to be done to improve the seasonal dynamics of forest vegetation in terms of leaf growth and leaf fall.* At this stage of the model, leaf growth is set to depend on cumulative air temperature in the spring, and leaf fall depends on return of a criterion

for day length in the fall. A detailed literature review should be undertaking to improve the universality of the equations used, by climate, latitude, and altitude.

- *Are there other input processes that need to be considered?* For example, for catchments where the forest vegetation intercepts fog and clouds, specific mechanisms need to be put in place to model effective fog and cloud droplet interception, in order to achieve a realistic water input/output mass balance at these locations. Similarly, the model should, eventually, accept condensation at dawn as another part of water input.
- *What is the role of the surrounding terrain on model outcome?* Clearly, a shading factor should be introduced to automatically diminish the solar energy input into snow/soil surfaces for areas that are shaded (e.g., depressions, northern aspects, open conditions north of forest, etc.). Temperature probes could be placed in various forest environments and open areas to determine how air temperature close to the ground and the snow surface compare with temperature recorded at the closest weather station, to check overall applicability of model input.

APPENDIX 1

MORE DETAILS ABOUT DEFINITIONS OF SOIL PARAMETERS

1. RELATIONSHIP BETWEEN BULK DENSITY (DB), PARTICLE DENSITY (DP) AND PORE SPACE

$$Dp = \frac{W_{solid}}{V_{solid}}$$

$$Db = \frac{W_{solid}}{V_{total}}$$

$$Db = \frac{M_{solid}}{V_{total}}$$

$$V_{pores} = V_{total} - V_{solid} = V_{total} * (1 - \frac{V_{solid}}{V_{total}}) = V_{total} * (1 - \frac{Db}{Dp})$$

So that: $$\frac{V_{pores}}{V_{total}} = 1 - \frac{Db}{Dp}$$

2. CALCULATION OF PARTICLE DENSITY (DP)

$$Dp = \frac{W_{solid}}{V_{solid}} = \frac{W_{OM} + W_{min}}{V_{OM} + V_{min}} = \frac{W_{OM} + W_{min}}{\frac{W_{OM}}{D_{OM}} + \frac{W_{min}}{D_{min}}}$$

With density of minerals D_{min} and density of OM D_{OM}:

$$D_{min} = \frac{W_{min}}{V_{min}}$$

$$D_{OM} = \frac{W_{OM}}{V_{OM}}$$

So that $$Dp = \frac{1}{\frac{W_{OM}}{D_{OM} * (W_{OM} + M_{min})} + \frac{W_{min}}{D_{min} * (W_{OM} + W_{min})}}$$

And finally $$\frac{1}{Dp} = \frac{OM}{D_{OM}} + \frac{1 - OM}{D_{min}}$$

Where $$OM = \frac{W_{OM}}{W_{OM} + W_{min}}$$

3. RELATIONSHIPS BETWEEN SOIL WATER CONTENTS BY WEIGHT AND BY VOLUME

$$SPw = \frac{W_{Water}^{SatPt}}{W_{solid}} = \frac{V_{water}^{SatPt} * D_{water}}{V_{total} * Db} = SPv * \frac{1}{Db}$$

so that: $$SPv = SPw * Db$$

and similarly:

$$FCv = FCw * Db$$

$$PWPv = PWPw * Db$$

APPENDIX 2

MODEL EQUATIONS

1. SOIL PARAMETERS

```
Clay_%_texture[Layer] =
if Texture[Layer]=1 then 0.04 else
if Texture[Layer]=2 then 0.06 else
if Texture[Layer]=3 then 0.06 else
if Texture[Layer]=4 then 0.10 else
if Texture[Layer]=5 then 0.13 else
if Texture[Layer]=6 then 0.18 else
if Texture[Layer]=7 then 0.26 else
if Texture[Layer]=8 then 0.34 else
if Texture[Layer]=9 then 0.34 else
if Texture[Layer]=10 then 0.41 else
if Texture[Layer]=11 then 0.46 else
if Texture[Layer]=12 then 0.50 else 0
```

```
Clay_fr[Layer] = Clay_%_texture[Layer]*(1-OM_%[Layer])
```

```
Db[Layer] = (1.234+(1.366-.723*Sand_fr[Layer])*(1-exp(-
0.011*Depth_mid_layer[Layer])))/(1+7.21*OM_fr[Layer])
```

```
Depth_FF_effect = Thickn_cm[FF]+Falling_leaf
```

```
Depth_mid_layer[FF] = Thickn_cm[FF]/2
Depth_mid_layer[A] = Thickn_cm[A]/2
Depth_mid_layer[B] = Thickn_cm[B]/2+Thickn_cm[A]
Depth_mid_layer[C] = Thickn_cm[C]/2+Thickn_cm[B]+Thickn_cm[A]
```

```
Dp[Layer] = 1/(OM_%[Layer]/1.1+(1-OM_%[Layer])/2.6)
```

```
FCv[Layer] = FCw%SPw[Layer]*SPv[Layer]
```

```
FCw%SPw[Layer] = (1-EXP(-0.32*(1-0.32*Sand_fr[Layer])-8.82*Clay_fr[Layer]))*(OM_fr[Layer]+(1-
OM_fr[Layer])/(1+8.03*EXP(-2.45*Db[Layer])))
```

```
FC_mm[Layer] = FCv[Layer]*Thickness_cm[Layer]*10
```

```
FC_soil_mm = FC_mm[A]+FC_mm[B]
```

```
K_sat_hydro[Layer] = 10^(-0.912+7.22*LOG10(2.6-Db[Layer])+1.99*Sand_fr[Layer])
```

```
K_sat_mm_yr[Layer] = K_sat_hydro[Layer]*10*24*365.25
```

```
OM_%[FF] = 1
```

OM_%[A] = 0
OM_%[B] = 0
OM_%[C] = 0

OM_fr[Layer] = OM_%[Layer]

PWPv[Layer] = PWPw%SPw[Layer]*SPv[Layer]

PWPw%FCw[Layer] = 1/(1+(FCw%SPw[Layer]*SPw[Layer])^1.93/(1-EXP(-1.13*Clay_fr[Layer]-0.0766*Sand_fr[Layer]-4.06*OM_fr[Layer]))^2.63)

PWPw%SPw[Layer] = PWPw%FCw[Layer]*FCw%SPw[Layer]

PWP_mm[Layer] = PWPv[Layer]*Thickness_cm[Layer]*10

PWP_Soil_mm = PWP_mm[A]+PWP_mm[B]

Sand_%_texture[Layer] = if Texture[Layer]=1 then 0.92 else
if Texture[Layer]=2 then 0.07 else
if Texture[Layer]=3 then 0.82 else
if Texture[Layer]=4 then 0.66 else
if Texture[Layer]=5 then 0.22 else
if Texture[Layer]=6 then 0.43 else
if Texture[Layer]=7 then 0.60 else
if Texture[Layer]=8 then 0.10 else
if Texture[Layer]=9 then 0.32 else
if Texture[Layer]=10 then 0.52 else
if Texture[Layer]=11 then 0.07 else
if Texture[Layer]=12 then 0.25 else 0

Sand_fr[Layer] = Sand_%_texture[Layer]*(1-OM_%[Layer])

SPv[Layer] = 1-Db[Layer]/Dp[Layer]

SPw[Layer] = SPv[Layer]/Db[Layer]

Texture[FF] = 0
Texture[A] = 6
Texture[B] = 4
Texture[C] = 4

Thickness_cm[FF] = Depth_FF_effect
Thickness_cm[A] = Thickn_cm[A]
Thickness_cm[B] = Thickn_cm[B]
Thickness_cm[C] = Thickn_cm[C]

Thickn_cm[FF] = 5
Thickn_cm[A] = 21
Thickn_cm[B] = 21

Thickn_cm[C] = 150

Thickn_Soil_mm = (Thickness_cm[A]+Thickness_cm[B])*10

Vpores_soil_mm = V_pores_mm[A]+V_pores_mm[B]

V_pores_cm[Layer] = SPv[Layer]*Thickness_cm[Layer]

V_pores_mm[Layer] = V_pores_cm[Layer]*10

2. VEGETATION

DegreeDay_cum(t) = DegreeDay_cum(t - dt) + (DegreeDay_m - Degree_reset) * dt
INIT DegreeDay_cum = 1600
INFLOWS:
DegreeDay_m = if T_air>0 then 12*T_air*Days_per_m else 0
OUTFLOWS:
Degree_reset = if T_air<0 then DegreeDay_cum/dt else 0

New_Litter(t) = New_Litter(t - dt) + (Leaf_fall - Litter_comp) * dt
INIT New_Litter = 0
INFLOWS:
Leaf_fall = if (Daylength>10.18 and Daylength<11.89 and Month_number>8) then LAI_max_decid*Deciduous_fr else 0
OUTFLOWS:

Litter_comp = New_Litter*Ppt/10*30.45

Conifer_fr = 0.5

Cutting_time = 10000

Daylength = (2*Sunset_angle/15)

Deciduous_fr = 0.5

Deciduous_LAI = if Month_number<8 then Leafout_switch*LAI_max_decid else (1-Leaffall_switch)*LAI_max_decid

DegreeDays = IF DegreeDay_cum>=DELAY(DegreeDay_cum, DT) THEN DegreeDay_cum ELSE 0

Depth_FF_effect = Thickn_cm[FF]+Falling_leaf

Falling_leaf = 10*New_Litter

Growth_function = if time >Cutting_time then Residues_k/(Residues_k+(1-Residues_k)*exp(-Growth_rate*(time-Cutting_time))) else 1

Growth_rate = 1.37*0

LAI = (Growth_function*(LAI_max_conif*Conifer_fr+Deciduous_fr*Deciduous_LAI))

LAI_max_conif = 8

LAI_max_decid = 5.5

Month_number = INT((time-INT(time))*12+1)

Ppt = Rain_fall+Snow_fall

Residues_k = 1

SAI = Growth_function*SAI_max*(Conifer_fr+Deciduous_fr)

SAI_max = 2

Thickn_cm[FF] = 5
Thickn_cm[A] = 21

Thickn_cm[B] = 21
Thickn_cm[C] = 150

T_air = T_air_Keji_82_02

VAI = (LAI+SAI/2)

VAI_max = (LAI_max_conif*Conifer_fr+LAI_max_decid*Deciduous_fr+SAI_max/2)

Veg_% = VAI/VAI_max*(Conifer_fr+Deciduous_fr)

Days_per_m = GRAPH(Month_number)
(1.00, 31.0), (2.00, 28.0), (3.00, 31.0), (4.00, 30.0), (5.00, 31.0), (6.00, 30.0), (7.00, 31.0), (8.00, 31.0), (9.00, 30.0), (10.0, 31.0), (11.0, 30.0), (12.0, 31.0), (13.0, 31.0)

Leaffall_switch = GRAPH(Daylength)
(10.2, 1.00), (11.9, 0.00)

Leafout_switch = GRAPH(DegreeDays)
(350, 0.00), (538, 0.105), (725, 0.305), (913, 0.655), (1100, 1.00)

3. ENERGY

En_ice[Layer](t) = En_ice[Layer](t - dt) + (En_freeze[Layer] - En_thaw[Layer]) * dt
INIT En_ice[Layer] = 0
INFLOWS:
En_freeze[Layer] = IF (T_'[Layer]<0 and En_ice[Layer]<En_limit[Layer]) THEN min(En_avail_freeze[Layer]/dt,(En_limit[Layer]-En_ice[Layer])/dt) ELSE 0
OUTFLOWS:
En_thaw[Layer] = IF (T_'[Layer]>0 AND En_ice[Layer]>0) THEN En_avail_thaw[Layer]/dt ELSE 0

Ice_in_FF_we(t) = Ice_in_FF_we(t - dt)
INIT Ice_in_FF_we = 0

Ice_in_soil_we(t) = Ice_in_soil_we(t - dt)
INIT Ice_in_soil_we = 0

Ice_in_subsoil_we(t) = Ice_in_subsoil_we(t - dt)
INIT Ice_in_subsoil_we = 0

Snowpack(t) = Snowpack(t - dt)
INIT Snowpack = 0

T_A(t) = T_A(t - dt) + (dT_AA) * dt
INIT T_A = 4.5
INFLOWS:
dT_AA = (T_'_A-T_A)/DT+T_frz_thw[A]/dt

T_B(t) = T_B(t - dt) + (dT_BB) * dt
INIT T_B = 4.5
INFLOWS:
dT_BB = (T_'_B-T_B)/DT+T_frz_thw[B]/dt

T_C(t) = T_C(t - dt) + (dT_CC) * dt
INIT T_C = 4.5
INFLOWS:
dT_CC = (T_'_C-T_C)/DT+T_frz_thw[C]/dt

```
T_FF(t) = T_FF(t - dt) + (dT_FF) * dt
INIT T_FF = 4.5
INFLOWS:
dT_FF = (T_'_FF-T_FF)/DT+T_frz_thw[FF]/dt

T_FF_surf(t) = T_FF_surf(t - dt) + (dT_FF_surf) * dt
INIT T_FF_surf = 4.5
INFLOWS:
dT_FF_surf = (T_'_FF_surf-T_FF_surf)/DT

T_Snowpack(t) = T_Snowpack(t - dt) + (dT_Snowpack) * dt
INIT T_Snowpack = 0
INFLOWS:
dT_Snowpack = (T_'_Snowpack-T_Snowpack)/DT+T_frz_thw_Snowpack/dt

T_Sub1(t) = T_Sub1(t - dt) + (dT_Sub1) * dt
INIT T_Sub1 = 4.5
INFLOWS:
dT_Sub1 = (T_'_Sub1-T_Sub1)/DT

T_Sub10(t) = T_Sub10(t - dt) + (dT_Sub10) * dt
INIT T_Sub10 = 7
INFLOWS:
dT_Sub10 = (T_'_Sub10-T_Sub10)/DT

T_Sub11(t) = T_Sub11(t - dt) + (dT_Sub11) * dt
INIT T_Sub11 = 7
INFLOWS:
dT_Sub11 = (T_'_Sub_11-T_Sub11)/DT

T_Sub12(t) = T_Sub12(t - dt) + (dT_Sub12) * dt
INIT T_Sub12 = 7
INFLOWS:
dT_Sub12 = (T_'_Sub12-T_Sub12)/DT

T_Sub2(t) = T_Sub2(t - dt) + (dT_Sub2) * dt
INIT T_Sub2 = 5.5
INFLOWS:
dT_Sub2 = (T_'_Sub2-T_Sub2)/DT

T_Sub3(t) = T_Sub3(t - dt) + (dT_Sub3) * dt
INIT T_Sub3 = 6
INFLOWS:
dT_Sub3 =  (T_'_Sub3-T_Sub3)/DT

T_Sub4(t) = T_Sub4(t - dt) + (dT_Sub4) * dt
INIT T_Sub4 = 6.5
INFLOWS:
dT_Sub4 = (T_'_Sub4-T_Sub4)/DT

T_Sub5(t) = T_Sub5(t - dt) + (dT_Sub5) * dt
INIT T_Sub5 = 7
INFLOWS:
dT_Sub5 = (T_'_Sub_5-T_Sub5)/DT

T_Sub6(t) = T_Sub6(t - dt) + (dT_Sub6) * dt
INIT T_Sub6 = 7
INFLOWS:
dT_Sub6 = (T_'_Sub6-T_Sub6)/DT
```

T_Sub7(t) = T_Sub7(t - dt) + (dT_Sub7) * dt
INIT T_Sub7 = 7
INFLOWS:
dT_Sub7 = (T_'_Sub_7-T_Sub7)/DT

T_Sub8(t) = T_Sub8(t - dt) + (dT_Sub8) * dt
INIT T_Sub8 = 7
INFLOWS:
dT_Sub8 = (T_'_Sub_8-T_Sub8)/DT

T_Sub9(t) = T_Sub9(t - dt) + (dT_Sub9) * dt
INIT T_Sub9 = 7
INFLOWS:
dT_Sub9 = (T_'_sub_9-T_Sub9)/DT

Water_in_FF(t) = Water_in_FF(t - dt)
INIT Water_in_FF = FC_mm[FF]*1.2

Water_in_soil(t) = Water_in_soil(t - dt)
INIT Water_in_soil = FC_soil_mm*1.02

Water_in_subsoil(t) = Water_in_subsoil(t - dt)
INIT Water_in_subsoil = FC_mm[C]*1.05

Water_retention(t) = Water_retention(t - dt)
INIT Water_retention = 0

aA = bA+cA+Htcap[A]/dt

aB = bB+cB+Htcap[B]/dt

aC = bC+cC+Htcap[C]/dt

Adj_solar_rad = 1

Air_vol_fr[Layer] = Air_vol_fr_%SP[Layer]*SPv[Layer]

Air_vol_fr_%SP[Layer] = 1-Ice_vol_fr_%SP[Layer]-Water_vol_fr_%SP[Layer]

Albedo_ff = Albedo_ff_wet+(EXP(.003286*Zenith_angle^1.5)-1)/100

Albedo_ff_wet = IF FF_wetness>.5 THEN .07 ELSE .14*(1-FF_wetness)

Albedo_snow = 0.9

Albedo_surf = IF Snowdepth_cm>0.1 THEN Albedo_snow*0+0.9 ELSE Albedo_ff

a_FF = bFF+cFF+Htcap_FFmain/dt

a_FF_surf = b_FF_surf+c_FF_surf+Htcap_FF_surf/dt

a_snowpack = if Snowdepth_cm>0.1 then b_snowpack+c_snowpack+Htcap_snow/dt else 0

a_Sub = bSub+c_Sub+Htcap_Subsoil/dt

a_Sub1 = bSub1+cSub1+Htcap_Subsoil/dt

bA = K_A_B/(Thickness_cm[A]/2+Thickness_cm[B]/2)

bB = K_B_C/(Thickness_cm[B]/2+Thickness_cm[C]/2)

bC = K_C_Sub/(Thickness_cm[C]/2+Thickness_sub_layers/2)

bFF = K_FF_A/((Thickness_cm[FF]-1)/2+Thickness_cm[A]/2)

bSub = bSub1

bSub1 = K_Sub/Thickness_sub_layers

b_FF_surf = K_FF_FF/((Thickness_cm[FF]-1)/2+1/2)

b_snowpack = if Snowdepth_cm>0.1 then K_FF_snow/(Snowdepth_cm/2+1/2) else 0

cA = bFF

cB = bA

cC = bB

cFF = b_FF_surf

Clay_fr[Layer] = Clay_%_texture[Layer]*(1-OM_%[Layer])

CosSunAngle = -tan(Latitude*3.14159265/180)*tan(Declination)

Cos_zen_day =
1/3.14159265*(Sunset_angle*3.14159265/180*SIN(Latitude*3.14159265/180)*SIN(Declination)+SIN(Sunset_angl
e*3.14159265/180)*COS(Latitude*3.14159265/180)*COS(Declination))

cSub1 = bC

c_FF_surf = if Snowdepth_cm>0.1 then K_FF_snow/(Snowdepth_cm/2+1/2) else K_FF_snow/(1/2)

c_snowpack = if Snowdepth_cm>0.1 then K_snow/(Snowdepth_cm/2) else 0

c_Sub = bSub1

Daylength = (2*Sunset_angle/15)

Db[Layer] = (1.234+(1.366-.723*Sand_fr[Layer])*(1-exp(-
0.011*Depth_mid_layer[Layer])))/(1+7.21*OM_fr[Layer])

Declination = (23.45*3.14159265/180)*sin((360/365)*(284+Julian_day)*(3.14159265/180))

d_A = Htcap[A]*T_A/dt

d_B = Htcap[B]*T_B/dt

d_C = Htcap[C]*T_C/dt

d_FF = Htcap_FFmain*T_FF/dt

d_FFsurf = Htcap_FF_surf*T_FF_surf/dt

D_ice = 0.9

d_snowpack = IF Snowdepth_cm>0.1 THEN Htcap_snow*T_Snowpack/dt ELSE 0

d_Sub1 = Htcap_Subsoil*T_Sub1/dt

d_Sub_10 = Htcap_Subsoil*T_Sub10/dt

d_Sub_11 = Htcap_Subsoil*T_Sub11/dt

d_Sub_12 = Htcap_Subsoil*T_Sub12/dt

d_Sub_2 = Htcap_Subsoil*T_Sub2/dt

d_Sub_3 = Htcap_Subsoil*T_Sub3/dt

d_Sub_4 = Htcap_Subsoil*T_Sub4/dt

d_Sub_5 = Htcap_Subsoil*T_Sub5/dt

d_Sub_6 = Htcap_Subsoil*T_Sub6/dt

d_Sub_7 = Htcap_Subsoil*T_Sub7/dt

d_Sub_8 = Htcap_Subsoil*T_Sub8/dt

d_Sub_9 = Htcap_Subsoil*T_Sub9/dt

En_available[Layer] = Htcap[Layer]*(T_'[Layer]-0)

En_avail_freeze[Layer] = IF T_'[Layer]<0 THEN (if water_avail[Layer]>=potential_wat_freeze[Layer] then -En_available[Layer] else Latent_heat_of_fusion*water_avail[Layer]) ELSE 0

En_avail_S = if T_'_Snowpack<0 then -Htcap_snow*T_'_Snowpack else 0

En_avail_thaw[Layer] = IF T_'[Layer]>0 THEN (if ice_avail[Layer]>=potential_ice_melt[Layer] then En_available[Layer] else Latent_heat_of_fusion*ice_avail[Layer]) ELSE 0

En_freeze_S = Water_freeze*Latent_heat_of_fusion

En_limit[Layer] = V_limit_cm[Layer]*Latent_heat_of_fusion

Extinction_f = 0.45/2

FCv[Layer] = FCw%SPw[Layer]*SPv[Layer]

FF_wetness = if SPv[FF]<=Ice_vol_fr[FF] then 1 else Water_vol_fr[FF]/(SPv[FF]-Ice_vol_fr[FF])

Frequency = 86400*365/pi

H2O_cm[FF] = Water_in_FF/10
H2O_cm[A] = (Water_in_soil/10)*Thickness_cm[A]/(Thickness_cm[A]+Thickness_cm[B])
H2O_cm[B] = (Water_in_soil/10)*Thickness_cm[B]/(Thickness_cm[A]+Thickness_cm[B])
H2O_cm[C] = Water_in_subsoil/10

Heat_ff = max(0,cFF*T_FF+c_snowpack*T_Snowpack)

Heat_top = if T_top_snow>=0 and Snowdepth_cm>0.1 then max(0, Solar_rad_yr+Stef_botzm_c*(Ta_adj+273)^3*(4-Albedo_surf/2)*Ta_adj+Q_airFF_k*Ta_adj+c_snowpack*T_Snowpack-Snow_evaporation*68) else 0

Htcap[Layer] = Htcap_vol[Layer]*Thickness_cm[Layer]

Htcap_Air = .003

Htcap_FFmain = Htcap_vol[FF]*(Thickness_cm[FF]-1)

Htcap_FF_surf = Htcap_vol[FF]*1

Htcap_Ice = .45

Htcap_Mineral = .48

Htcap_OM = .6

Htcap_snow = Htcap_water*(Water_retention/10)+Htcap_Ice*(Snowpack/10)*D_ice

Htcap_Subsoil = Vol_Htcap_sub*Thickness_sub_layers

Htcap_vol[Layer] =
Htcap_Air*Air_vol_fr[Layer]+Htcap_Ice*Ice_vol_fr[Layer]+Htcap_Mineral*Min_vol_fr[Layer]+Htcap_OM*OM_vol_fr[Layer]+Htcap_water*Water_vol_fr[Layer]

Htcap_water = 1

ice_avail[Layer] = Ice_cm[Layer]

Ice_cm[FF] = Ice_in_FF_we/10
Ice_cm[A] = (Ice_in_soil_we/10)*Thickness_cm[A]/(Thickness_cm[A]+Thickness_cm[B])
Ice_cm[B] = (Ice_in_soil_we/10)*Thickness_cm[B]/(Thickness_cm[A]+Thickness_cm[B])
Ice_cm[C] = Ice_in_subsoil_we/10

Ice_vol_fr[Layer] = Ice_vol_fr_%SP[Layer]*SPv[Layer]

Ice_vol_fr_%SP[Layer] = V_ice_cm3[Layer]/V_pores_cm3[Layer]

Ice_we_cm[Layer] = En_ice[Layer]/Latent_heat_of_fusion

Julian_day = 365*(time-int(time))

K[Layer] = Therm_K[Layer]*2.4/1000*Sec_per_yr

Ke[Layer] = Ke_frozen[Layer]*Ice_vol_fr[Layer]+Ke_unf[Layer]*(1-Ice_vol_fr[Layer])

Ke_frozen[Layer] = Saturation_%[Layer]

Ke_unf[Layer] = ((EXP(13.6*Saturation_%[Layer])/(2.3+EXP(13.6*Saturation_%[Layer])))^3-((1-Saturation_%[Layer])/(1+2.3))^3)*Saturation_%[Layer]^0.5

Ke_unf_sub = ((EXP(13.6*Saturation_%_sub)/(2.3+EXP(13.6*Saturation_%_sub)))^3-((1-Saturation_%_sub)/(1+2.3))^3)*Saturation_%_sub^0.5

Ko[Layer] = K_om^OM_fr[Layer]*K_min^(1-OM_fr[Layer])

Ks[Layer] = K_quartz^Quartz_%[Layer]*Ko[Layer]^(1-Quartz_%[Layer])

K_A_B =
K[A]*K[B]*(Thickness_cm[A]/2+Thickness_cm[B]/2)/(K[A]*Thickness_cm[B]/2+K[B]*Thickness_cm[A]/2)

K_B_C =
K[B]*K[C]*(Thickness_cm[B]/2+Thickness_cm[C]/2)/(K[B]*Thickness_cm[C]/2+K[C]*Thickness_cm[B]/2)

K_C_Sub =
K[C]*Sub_K*(Thickness_cm[C]/2+Thickness_sub_layers/2)/(K[C]*Thickness_sub_layers/2+Sub_K*Thickness_cm[C]/2)

K_dry[Layer] = (137*Db[Layer]+64.7)/(2700-947*Db[Layer])

K_FF_A = K[FF]*K[A]*(Thickness_cm[A]/2+(Thickness_cm[FF]-1)/2)/(K[FF]*Thickness_cm[A]/2+K[A]*(Thickness_cm[FF]-1)/2)

K_FF_FF = K[FF]

K_FF_snow = IF Snowdepth_cm>0.1 THEN Snow_K*K[FF]*(Snowdepth_cm/2+1/2)/(Snow_K*1/2+K[FF]*Snowdepth_cm/2) else K[FF]

K_heat_transfer = 0.4

K_Heat_transfer_conv = 50000

K_ice = 2.2

K_min = 2

K_om = 0.25

K_quartz = 7.7

K_sat[Layer] = Ks[Layer]^(1-n[Layer])*K_ice^(n[Layer]-wu[Layer])*K_water^wu[Layer]

K_sat_sub = Ks[C]^(1-n_sub)*K_ice^(n_sub-wu_sub)*K_water^wu_sub

K_snow = IF Snowdepth_cm>0.1 THEN Snow_K ELSE 0

K_Sub = Sub_K

K_water = 0.57

Latent_heat_of_fusion = 79.67

Latitude = 44.434

Min_vol_fr[Layer] = (1-OM_%[Layer])*Db[Layer]/2.6

n[Layer] = SPv[Layer]

n_sub = SPv[C]

OM_%[FF] = 1
OM_%[A] = 0
OM_%[B] = 0
OM_%[C] = 0

OM_fr[Layer] = OM_%[Layer]

OM_vol_fr[Layer] = OM_%[Layer]*Db[Layer]/1.1

P0 = 0

potential_ice_melt[Layer] = if T_'[Layer]>0 then En_available[Layer]/Latent_heat_of_fusion else 0

potential_wat_freeze[Layer] = En_available[Layer]/(-Latent_heat_of_fusion)

Pot_wat_freeze_snow = En_avail_S/Latent_heat_of_fusion

P_A = bA/(aA-cA*P_FF)

P_B = bB/(aB-cB*P_A)

P_C = bC/(aC-cC*P_B)

P_FF = bFF/(a_FF-cFF*P_FF_surf)

P_FF_surf = IF Snowdepth_cm>0.1 then b_FF_surf/(a_FF_surf-c_FF_surf*P_snowpack) else

b_FF_surf/(a_FF_surf-c_FF_surf*P0)

P_snowpack = if Snowdepth_cm>0.1 then b_snowpack/(a_snowpack-c_snowpack*P0) else 0

P_Sub1 = bSub1/(a_Sub1-cSub1*P_C)

P_Sub10 = bSub/(a_Sub-c_Sub*P_Sub9)

P_Sub11 = bSub/(a_Sub-c_Sub*P_Sub10)

P_Sub12 = bSub/(a_Sub-c_Sub*P_Sub11)

P_Sub2 = bSub/(a_Sub-c_Sub*P_Sub1)

P_Sub3 = bSub/(a_Sub-c_Sub*P_Sub2)

P_Sub4 = bSub/(a_Sub-c_Sub*P_Sub3)

P_Sub5 = bSub/(a_Sub-c_Sub*P_Sub4)

P_Sub6 = bSub/(a_Sub-c_Sub*P_Sub5)

P_Sub7 = bSub/(a_Sub-c_Sub*P_Sub6)

P_Sub8 = bSub/(a_Sub-c_Sub*P_Sub7)

P_Sub9 = bSub/(a_Sub-c_Sub*P_Sub8)

Q0 = T_surface

Quartz_%[Layer] = 0.652*Sand_fr[Layer]+0.204*Clay_fr[Layer]+0*OM_fr[Layer]+0.470*(1-Sand_fr[Layer]-Clay_fr[Layer]-OM_fr[Layer])

Q_AA = (d_A+cA*Q_FF)/(aA-cA*P_FF)

Q_airFF_k = K_Heat_transfer_conv*(1-EXP(MIN(6.5, VAI)-6.8))

Q_BB = (d_B+cB*Q_AA)/(aB-cB*P_A)

Q_CC = (d_C+cC*Q_BB)/(aC-cC*P_B)

Q_FF = (d_FF+cFF*Q_FF_surf)/(a_FF-cFF*P_FF_surf)

Q_FF_surf = IF Snowdepth_cm>0.1 then (d_FFsurf+c_FF_surf*Q_snowpack)/(a_FF_surf-c_FF_surf*P_snowpack)

else (d_FFsurf+c_FF_surf*Q0)/(a_FF_surf-c_FF_surf*P0)

Q_snowpack = if Snowdepth_cm>0.1 then (d_snowpack+c_snowpack*Q0)/(a_snowpack-c_snowpack*P0) else 0

Q_Sub1 = (d_Sub1+cSub1*Q_CC)/(a_Sub1-cSub1*P_C)

Q_Sub10 = (d_Sub_10+c_Sub*Q_Sub9)/(a_Sub-c_Sub*P_Sub9)

Q_Sub11 = (d_Sub_11+c_Sub*Q_Sub10)/(a_Sub-c_Sub*P_Sub10)

Q_Sub12 = (d_Sub_12+c_Sub*Q_Sub11)/(a_Sub-c_Sub*P_Sub11)

Q_Sub2 = (d_Sub_2+c_Sub*Q_Sub1)/(a_Sub-c_Sub*P_Sub1)

Q_Sub3 = (d_Sub_3+c_Sub*Q_Sub2)/(a_Sub-c_Sub*P_Sub2)

Q_Sub4 = (d_Sub_4+c_Sub*Q_Sub3)/(a_Sub-c_Sub*P_Sub3)

Q_Sub5 = (d_Sub_5+c_Sub*Q_Sub4)/(a_Sub-c_Sub*P_Sub4)

Q_Sub6 = (d_Sub_6+c_Sub*Q_Sub5)/(a_Sub-c_Sub*P_Sub5)

Q_Sub7 = (d_Sub_7+c_Sub*Q_Sub6)/(a_Sub-c_Sub*P_Sub6)

Q_Sub8 = (d_Sub_8+c_Sub*Q_Sub7)/(a_Sub-c_Sub*P_Sub7)

Q_Sub9 = (d_Sub_9+c_Sub*Q_Sub8)/(a_Sub-c_Sub*P_Sub8)

Rain_H = Rain_Throughfall*T_rain

Rain_Throughfall = Rain_through_canopy

Sand_fr[Layer] = Sand_%_texture[Layer]*(1-OM_%[Layer])

Saturation_%[Layer] = (Water_vol_fr[Layer]+Ice_vol_fr[Layer])/SPv[Layer]

Saturation_%_sub = (Water_vol_fr[C]+Ice_vol_fr[C])/SPv[C]

Sec_per_yr = 86400*365

Snowdepth_cm = Snowpack/SnowDens/10

SnowD_ann = 3.87

Snowmelt = if (Heat_ff+Heat_top+Rain_H)<0 then 0 else (Heat_ff+Heat_top+Rain_H)/(Latent_heat_of_fusion)*10

Snow_dens2 = Delay_wet_snow

Snow_evaporation = IF Snowpack>0 THEN 12*Surface_evap_fct ELSE 0

Snow_K = Sec_per_yr*.001/4.184*(.358+26.4*Snow_dens2^2)

Solar_rad_daily = SoRad_day_Keji

Solar_rad_yr = exp(-Extinction_f*(VAI))/5*(Solar_rad_daily*365*(1-Albedo_surf))*Adj_solar_rad

SPv[Layer] = 1-Db[Layer]/Dp[Layer]

Stef_botzm_c = 81.7*10^(-12)*Sec_per_yr/60

Sub_K = Therm_K_sub*2.4/1000*Sec_per_yr

SunAngle0 = arctan(sqrt(1-CosSunAngle*CosSunAngle)/CosSunAngle)*180/3.14159265

Sunset_angle = IF SunAngle0>0 THEN SunAngle0 ELSE 180+SunAngle0

Surface_evap_fct = (min(4,LAI/4)-1)*(min(4,LAI/4)-1)*(1-min(2,SAI_max)/8)*PET/2

Ta_adj = IF Temperature_code=1 THEN T_air ELSE

(if (Snowdepth_cm>0.1 and T_air>0) then T_air*K_heat_transfer else T_air+((-.11+.96*T_air-.000080*T_air*T_air*T_air)-T_air)*LOGN(1+MIN(6.5,VAI))/LOGN(1+6.5))

Temperature_code = 0

Therm_K[Layer] = (K_sat[Layer]-K_dry[Layer])*Ke[Layer]+K_dry[Layer]

Therm_K_sub = (K_sat_sub-K_dry[C])*Ke_unf_sub+K_dry[C]

Thickness_cm[FF] = Depth_FF_effect
Thickness_cm[A] = Thickn_cm[A]
Thickness_cm[B] = Thickn_cm[B]
Thickness_cm[C] = Thickn_cm[C]

Thickness_sub_layers = 100

T_'[FF] = T_'_FF
T_'[A] = T_'_A
T_'[B] = T_'_B
T_'[C] = T_'_C

T_'_A = P_A*T_'_B+Q_AA

T_'_B = P_B*T_'_C+Q_BB

T_'_C = P_C*T_'_Sub1+Q_CC

T_'_FF = P_FF*T_'_A+Q_FF

T_'_FF_surf = IF Snowdepth_cm>0.1 THEN MIN(0,P_FF_surf*T_'_FF+Q_FF_surf) ELSE P_FF_surf*T_'_FF+Q_FF_surf

T_'_Snowpack = MIN(0, P_snowpack*T_'_FF_surf+Q_snowpack)

T_'_Sub1 = P_Sub1*T_'_Sub2+Q_Sub1

T_'_Sub10 = P_Sub10*T_'_Sub_11+Q_Sub10

T_'_Sub12 = P_Sub12*T_bottom+Q_Sub12

T_'_Sub2 = P_Sub2*T_'_Sub3+Q_Sub2

T_'_Sub3 = P_Sub3*T_'_Sub4+Q_Sub3

T_'_Sub4 = P_Sub4*T_'_Sub_5+Q_Sub4

T_'_Sub6 = P_Sub6*T_'_Sub_7+Q_Sub6

T_'_Sub_11 = P_Sub11*T_'_Sub12+Q_Sub11

T_'_Sub_5 = P_Sub5*T_'_Sub6+Q_Sub5

T_'_Sub_7 = P_Sub7*T_'_Sub_8+Q_Sub7

T_'_Sub_8 = P_Sub8*T_'_sub_9+Q_Sub8

T_'_sub_9 = P_Sub9*T_'_Sub10+Q_Sub9

T_air = T_air_Keji_82_02

T_ampd = 12.12

T_ann = 6.7

T_bottom = T_ann+(T_ampd/2)*(1-exp(-SnowD_ann/SQRT(0.00032*Frequency/0.2)))

T_frz_thw[Layer] = if T_'[Layer]<0 then En_freeze[Layer]/Htcap[Layer]*dt

else (if T_'[Layer]>0 then -En_thaw[Layer]/Htcap[Layer]*dt else 0)

T_frz_thw_Snowpack = if Snowdepth_cm> 0.1 then En_freeze_S/Htcap_snow else 0

T_rain = max(0, T_air)

T_surface = IF Snowdepth_cm>0.1 THEN min(0,T_top_snow) ELSE T_top_NoSnow

T_top_NoSnow = if Snowdepth_cm>0.1 then 0 else
(Solar_rad_yr+Stef_botzm_c*(Ta_adj+273)^3*4*Ta_adj+Q_airFF_k*Ta_adj+cFF*T_FF-
68*12*Surface_evap_fct)/(4*Stef_botzm_c*(Ta_adj+273)^3+Q_airFF_k+cFF)

T_top_snow = IF Snowdepth_cm>0.1 THEN min(0,
(Solar_rad_yr+Stef_botzm_c*(Ta_adj+273)^3*4*Ta_adj+Q_airFF_k*Ta_adj+c_snowpack*T_Snowpack-
Snow_evaporation*68)/(4*Stef_botzm_c*(Ta_adj+273)^3+Q_airFF_k+c_snowpack)) else 0

VAI = (LAI+SAI/2)

Vol_Htcap_sub =
Htcap_Air*Air_vol_fr[C]+Htcap_Mineral*Min_vol_fr[C]+Htcap_OM*OM_vol_fr[C]+Htcap_water*(Water_vol_fr
[C]+Ice_vol_fr[C])

V_ice_cm3[Layer] = 1*Ice_we_cm[Layer]*(1/D_ice)

V_limit_cm[Layer] = FCv[Layer]*Thickness_cm[Layer]*D_ice/1

V_pores_cm[Layer] = SPv[Layer]*Thickness_cm[Layer]

V_pores_cm3[Layer] = 1*V_pores_cm[Layer]

V_water_cm3[Layer] = H2O_cm[Layer]*1

water_avail[Layer] = H2O_cm[Layer]

Water_freeze = min(Pot_wat_freeze_snow,Wat_ret_cm)

Water_vol_fr[Layer] = Water_vol_fr_%SP[Layer]*SPv[Layer]

Water_vol_fr_%SP[Layer] = min(V_water_cm3[Layer]/V_pores_cm3[Layer],1-Ice_vol_fr_%SP[Layer])

Wat_freeze[Layer] = En_freeze[Layer]/Latent_heat_of_fusion

Wat_ret_cm = Water_retention/10

Wat_thaw[Layer] = En_thaw[Layer]/Latent_heat_of_fusion

wu[Layer] = if (Water_vol_fr[Layer] = 0 and Ice_vol_fr[Layer] =0) then 1*SPv[Layer] else (Water_vol_fr[Layer]/(Water_vol_fr[Layer]+Ice_vol_fr[Layer]))*SPv[Layer]

wu_sub = 1*SPv[C]

Zenith_angle = arctan(sqrt(1-Cos_zen_day^2)/Cos_zen_day)*180/3.14159265

4. HYDROLOGY

Delay_dry_snow(t) = Delay_dry_snow(t - dt) + (D_dry_new - D_dry_reset) * dt
INIT Delay_dry_snow = 0.1
INFLOWS:
D_dry_new = D_dry_snow/dt
OUTFLOWS:
D_dry_reset = Delay_dry_snow/dt

Delay_wet_snow(t) = Delay_wet_snow(t - dt) + (D_wet_new - D_wet_reset) * dt
INIT Delay_wet_snow = D_fresh_snow
INFLOWS:
D_wet_new = D_wet_snow/dt
OUTFLOWS:
D_wet_reset = Delay_wet_snow/dt

Ice_in_FF_we(t) = Ice_in_FF_we(t - dt) + (Freezing_in_FF - Thawing_in_FF) * dt
INIT Ice_in_FF_we = 0
INFLOWS:
Freezing_in_FF = Wat_freeze[FF]*10
OUTFLOWS:
Thawing_in_FF = Wat_thaw[FF]*10

Ice_in_soil_we(t) = Ice_in_soil_we(t - dt) + (Freezing_in_soil - Thawing_in_soil) * dt
INIT Ice_in_soil_we = 0
INFLOWS:
Freezing_in_soil = (Wat_freeze[A]+Wat_freeze[B])*10
OUTFLOWS:
Thawing_in_soil = (Wat_thaw[A]+Wat_thaw[B])*10

Ice_in_subsoil_we(t) = Ice_in_subsoil_we(t - dt) + (Freezing_in_subsoil - Thawing_in_subsoil) * dt
INIT Ice_in_subsoil_we = 0
INFLOWS:
Freezing_in_subsoil = Wat_freeze[C]*10
OUTFLOWS:
Thawing_in_subsoil = Wat_thaw[C]*10

Rain_in_canopy(t) = Rain_in_canopy(t - dt) + (Rain - Rain_Interception - Rain_Throughfall) * dt
INIT Rain_in_canopy = 0
INFLOWS:
Rain = Rain_fall*30.45*12
OUTFLOWS:
Rain_Interception = MIN(Rain_in_canopy/DT, 12*PET)
Rain_Throughfall = Rain_through_canopy

Snowpack(t) = Snowpack(t - dt) + (Snow_Throughfall + Water_freeze_flow - Snow_evaporation - Snowmelt_flow) * dt
INIT Snowpack = 0
INFLOWS:
Snow_Throughfall = Snow_through_canopy
Water_freeze_flow = Water_freeze*10/dt
OUTFLOWS:
Snow_evaporation = IF Snowpack>0 THEN 12*Surface_evap_fct ELSE 0

Snowmelt_flow = min(Snowpack/dt, Snowmelt)

Snow_in_canopy(t) = Snow_in_canopy(t - dt) + (Snow - Snow_Throughfall - Snow_Interception) * dt
INIT Snow_in_canopy = 0
INFLOWS:
Snow = 12*Snow_fall*30.45
OUTFLOWS:
Snow_Throughfall = Snow_through_canopy
Snow_Interception = MIN(Snow_in_canopy/DT, 12*PET)

Surface(t) = Surface(t - dt) + (Water_runoff - Infiltration - Surface_Runoff) * dt
INIT Surface = 0
INFLOWS:
Water_runoff = Water_through_snow
OUTFLOWS:
Infiltration = Inf_rate
Surface_Runoff = Runoff_rate

Water_in_FF(t) = Water_in_FF(t - dt) + (Infiltration + Thawing_in_FF - Perc_FF - Interflow_in_FF - Evaporation_forest_floor - Freezing_in_FF) * dt
INIT Water_in_FF = FC_mm[FF]
INFLOWS:
Infiltration = Inf_rate
Thawing_in_FF = Wat_thaw[FF]*10
OUTFLOWS:
Perc_FF = Perc_forest_floor
Interflow_in_FF = Flow_slope_forest_floor
Evaporation_forest_floor = EvTr_FF+Tr_FF
Freezing_in_FF = Wat_freeze[FF]*10

Water_in_soil(t) = Water_in_soil(t - dt) + (Perc_FF + Thawing_in_soil - Perc_Soil - Interflow_in_Soil - Evaporation_Soil - Freezing_in_soil) * dt
INIT Water_in_soil = FC_soil_mm
INFLOWS:
Perc_FF = Perc_forest_floor
Thawing_in_soil = (Wat_thaw[A]+Wat_thaw[B])*10
OUTFLOWS:

Perc_Soil = Perc_from_soil
Interflow_in_Soil = Flow_slope_soil
Evaporation_Soil = EvTr_soil
Freezing_in_soil = (Wat_freeze[A]+Wat_freeze[B])*10

Water_in_subsoil(t) = Water_in_subsoil(t - dt) + (Perc_Soil + Thawing_in_subsoil - Interflow_in_subsoil - Freezing_in_subsoil) * dt
INIT Water_in_subsoil = FC_mm[C]
INFLOWS:
Perc_Soil = Perc_from_soil
Thawing_in_subsoil = Wat_thaw[C]*10
OUTFLOWS:
Interflow_in_subsoil = Flow_subsoil_max
Freezing_in_subsoil = Wat_freeze[C]*10

Water_retention(t) = Water_retention(t - dt) + (Snowmelt_flow + Rain_Throughfall - Water_runoff - Water_freeze_flow) * dt
INIT Water_retention = 0
INFLOWS:
Snowmelt_flow = min(Snowpack/dt, Snowmelt)
Rain_Throughfall = Rain_through_canopy
OUTFLOWS:

Water_runoff = Water_through_snow

Water_freeze_flow = Water_freeze*10/dt

Adj_flow_slope_C = 1

Adj_flow_slope_FF = 1

Adj_flow_slope_soil = 1

Adj_infiltration_soil = 1

adj_infiltration_surface = 1

adj_runoff_slope_surface = 1

AET_FF_part = 1-Root_depth_k^Thickness_cm[FF]

Area = 17700

compaction = if (Snow_Throughfall*dt>2 or Rain_Throughfall*dt>0) then K_compaction else 1

Daylength = (2*Sunset_angle/15)

Diffusion_surface = LOGN((Reference_H-Zero_plan_H)/rough_H)

Diffusion_velocity = Von_Karman_k^2*Wind_speed/Diffusion_surface^2

Distance_Coast = 80

D_dry_snow = if WE_Snow>WE_Out_snow then min(D_ice,compaction*(WE_Snow-WE_Out_snow+WE_Fresh_snow+WE_Freeze)/((WE_Snow-WE_Out_snow)/D_dry_snow_delayed+WE_Fresh_snow/D_fresh_snow)) else D_fresh_snow

D_dry_snow_delayed = max(0.1,Delay_dry_snow)

D_fresh_snow = 0.15

D_ice = 0.9

D_wet_snow = if WE_Snow>WE_Out_snow then min(1,(WE_Snow-WE_Out_snow+WE_Fresh_snow+WE_Freeze+WE_Liquid)/((WE_Snow-WE_Out_snow)/D_dry_snow_delayed+WE_Fresh_snow/D_fresh_snow)) else D_fresh_snow

empty_Pore_C = Max(0,V_pores_mm[C]-Water_in_subsoil-Ice_in_subsoil_we/D_ice)

empty_Pore_FF = MAX(0, V_pores_mm[FF]-Water_in_FF-Ice_in_FF_we/D_ice)

empty_Pore_Soil = Max(0, Vpores_soil_mm-Water_in_soil-Ice_in_soil_we/D_ice)

EvTr_FF = if Water_in_FF>PWP_mm[FF] then

PET_FF*Veg_%*((Water_in_FF-PWP_mm[FF])/(V_pores_mm[FF]-PWP_mm[FF]))^K_evap else 0

EvTr_soil = if Water_in_soil>PWP_Soil_mm then PET_Soil*Veg_%*((Water_in_soil-PWP_Soil_mm)/(Vpores_soil_mm-PWP_Soil_mm))^K_evap else 0

FCv_snow = 0.05

FC_canopy_rain_mm = k_leaf_intercep_rain*VAI

FC_canopy_snow_mm = k_leaf_intercep_snow*VAI

FC_C_adj_mm = if Ice_in_subsoil_we<FC_mm[C] then FC_mm[C]-Ice_in_subsoil_we else 0

FC_FF_adj_mm = if Ice_in_FF_we<FC_mm[FF] then FC_mm[FF]-Ice_in_FF_we else 0

FC_mm[Layer] = FCv[Layer]*Thickness_cm[Layer]*10

FC_snow_mm = if (Snowpack-FCv_snow)>0 then FCv_snow*((Snowpack-WE_Out_snow_2)/SnowDens) else 0

FC_soil_adj_mm = if Ice_in_soil_we<FC_soil_mm then FC_soil_mm-Ice_in_soil_we else 0

FC_soil_mm = FC_mm[A]+FC_mm[B]

Flow_availability_C = if Water_in_subsoil>FC_C_adj_mm then ((Water_in_subsoil-FC_C_adj_mm)/(V_pores_mm[C]-FC_C_adj_mm)) else 0

Flow_slope_FF_max = (K_sat_mm_yr[FF]*Water_column_%_FF)*Gradient_slope*(Water_column_%_FF*1*Thickn_cm[FF])*Adj_flow_slope_FF

Flow_slope_forest_floor = min(Flow_slope_FF_max,Water_left_FF)

Flow_slope_soil = min(Flow_slope_Soil_max,Water_left_soil)

Flow_slope_Soil_max = (max(K_sat_mm_yr[A],K_sat_mm_yr[B])*Water_column%_soil)*Gradient_slope*(Water_column%_soil*1*(Thickn_cm[A]+Thickn_cm[B]))*Adj_flow_slope_soil

Flow_subsoil_max = (K_sat_mm_yr[C]*Flow_availability_C)*Gradient_slope*(Flow_availability_C*1)*Adj_flow_slope_C

Fog_Code = 1

Fog_drip = Fog_hrs_per_month*Fog_drip_rate

Fog_drip_k = Unit_conversion*Fog_wc*(Diffusion_velocity+Sedimentation_v)

Fog_drip_rate = Fog_drip_k*LOGN(1+(EXP(1)-1)*VAI/6.5)

Fog_hrs_per_month = Fog_mist_hrs*(1/3+0.1*2/3)

Fog_mist_hrs = if T_air<0 then 0 else

Fog_Code*MAX(0,EXP(4.631-.003711*Distance_Coast+Fog_month_F-.004664*(T_air+3.5)^2)-1)

Fog_month_F = .09322*(Season_month-14)*(Season_month-14)+.1269*SIN((Season_month+4.25)*2*3.14159265/7)

Fog_wc = .1

Gradient_slope = 1/100

Growth_function = if time >Cutting_time then Residues_k/(Residues_k+(1-Residues_k)*exp(-Growth_rate*(time-Cutting_time))) else 1

Inf_rate = MIN(Water_surface_free*adj_infiltration_surface, Space_in_FF)

In_C_max =
(min(min(K_sat_mm_yr[A],K_sat_mm_yr[B]),K_sat_mm_yr[C])*Water_column%_soil)*Water_column%_soil*(1*1)

In_Soil_max =
(min(K_sat_mm_yr[FF],min(K_sat_mm_yr[A],K_sat_mm_yr[B]))*Water_column_%_FF)*Water_column_%_FF*(1*1)*Adj_infiltration_soil

K_compaction = 1

K_evap = 0.5

k_leaf_intercep_rain = 0.25

k_leaf_intercep_snow = 1

K_sat_mm_yr[Layer] = K_sat_hydro[Layer]*10*24*365.25

LAI = (Growth_function*(LAI_max_conif*Conifer_fr+Deciduous_fr*Deciduous_LAI))

No_PET_days = Fog_mist_hrs/24

Output_C = Interflow_in_subsoil

Output_FF = Perc_FF+Interflow_in_FF+Evaporation_forest_floor

Output_Soil = Perc_Soil+Interflow_in_Soil+Evaporation_Soil

Perc_forest_floor = MIN(In_Soil_max,Space_in_Soil)

Perc_from_soil = MIN(In_C_max,Space_in_C)

PET = PET_k*(0.1651*(Daylength/12)*PET_day_month*Vapour)

PETadj = max(0,12*PET)

PET_day_month = 365/12-No_PET_days

PET_FF = AET_FF_part*PETadj

PET_k = .049*T_July+.140

PET_Soil = PETadj*(1-AET_FF_part)

Ppt = Rain_fall+Snow_fall

PWP_mm[Layer] = PWPv[Layer]*Thickness_cm[Layer]*10

PWP_Soil_mm = PWP_mm[A]+PWP_mm[B]

Rain_fall = Rain_Keji

Rain_through_canopy = max(0,Rain-(FC_canopy_rain_mm-Rain_in_canopy)/dt)

Reference_H = Tree_H+2

rough_H = 10^(LOG10(Reference_H)-.98)

Runoff_rate = Water_left_surf*Gradient_slope*adj_runoff_slope_surface

SAI_max = 2

Season_month = IF (time-INT(time))*12 >8 THEN (time-INT(time))*12 ELSE (time-INT(time))*12+12

Sedimentation_v = .02

SnowDens = Delay_dry_snow

Snowdepth_cm = Snowpack/SnowDens/10

Snowmelt = if (Heat_ff+Heat_top+Rain_H)<0 then 0 else (Heat_ff+Heat_top+Rain_H)/(Latent_heat_of_fusion)*10

Snow_dens2 = Delay_wet_snow

Snow_fall = Snow_Keji

Snow_through_canopy = if Rain>0 then Snow_in_canopy/dt+Snow else max(0,Snow-(FC_canopy_snow_mm-Snow_in_canopy)/dt)

Space_in_C = Output_C+empty_Pore_C/dt

Space_in_FF = Output_FF+empty_Pore_FF/dt

Space_in_Soil = Output_Soil+empty_Pore_Soil/dt

Streamflow_mm_day = Streamflow_mm_yr/365.25

Streamflow_mm_yr = (Surface_Runoff+Interflow_in_FF+Interflow_in_Soil+Interflow_in_subsoil)

Streamflow_p = (Surface_Runoff+Interflow_in_FF+Interflow_in_Soil+Interflow_in_subsoil)/12+Stream_delay

Stream_delay = 0.43+.044*LOGN(Area)*2

Surface_evap_fct = (min(4,LAI/4)-1)*(min(4,LAI/4)-1)*(1-min(2,SAI_max)/8)*PET/2

Tree_H = 15*Growth_function

Tree_root_depth_code = 4

Tr_FF = (1-Veg_%)*PET_FF*(Water_in_FF/V_pores_mm[FF])^K_evap

T_air = T_air_Keji_82_02

T_July = 18.8

Unit_conversion = 10*3600/10000

VAI = (LAI+SAI/2)

Veg_% = VAI/VAI_max*(Conifer_fr+Deciduous_fr)

Von_Karman_k = .41

Vpores_soil_mm = V_pores_mm[A]+V_pores_mm[B]

V_pores_mm[Layer] = V_pores_cm[Layer]*10

Water_column%_soil = if Water_in_soil>FC_soil_adj_mm then (Water_in_soil-FC_soil_adj_mm)/(Vpores_soil_mm-FC_soil_adj_mm)else 0

Water_column_%_FF = if Water_in_FF>FC_FF_adj_mm then ((Water_in_FF-FC_FF_adj_mm)/(V_pores_mm[FF]-FC_FF_adj_mm)) else 0

Water_freeze = min(Pot_wat_freeze_snow,Wat_ret_cm)

Water_free_FF = if Water_in_FF>FC_FF_adj_mm then (Water_in_FF-FC_FF_adj_mm)/dt else 0

Water_left_FF = max(0,Water_free_FF-Perc_forest_floor)

Water_left_soil = max(0,Water_soil_free-Perc_from_soil)

Water_left_surf = max(0,Water_surface_free-Inf_rate)

Water_soil_free = if Water_in_soil>FC_soil_adj_mm then (Water_in_soil-FC_soil_adj_mm)/dt else 0

Water_surface_free = Surface/dt+Water_runoff

Water_through_snow = if Snowdepth_cm>0 then max(0,Wat_ret_input+Water_retention/dt-FC_snow_mm/dt) else Wat_ret_input+Water_retention/dt

Wat_freeze[Layer] = En_freeze[Layer]/Latent_heat_of_fusion

Wat_ret_input = Snowmelt_flow+Rain_Throughfall

Wat_thaw[Layer] = En_thaw[Layer]/Latent_heat_of_fusion

WetLandEff = ((Ppt*30.45-PET)-Streamflow_p)*WetlandRatio

WetlandRatio = 0.5

WE_Freeze = Water_freeze_flow*dt

WE_Fresh_snow = Snow_Throughfall*dt

WE_Liquid = Water_retention

WE_Out_snow = (Snowmelt_flow+Snow_evaporation)*dt

WE_Out_snow_2 = (Snowmelt_flow+Snow_evaporation)*dt

WE_Snow = Snowpack

Wind_speed = 5

Zero_plan_H = Tree_H*2/3

Root_depth_k = GRAPH(Tree_root_depth_code)
(0.00, 0.00), (0.5, 0.77), (1.00, 0.89), (1.50, 0.9), (2.00, 0.91), (2.50, 0.92), (3.00, 0.93), (3.50, 0.94), (4.00, 0.95), (4.50, 0.96), (5.00, 0.97), (5.50, 0.98), (6.00, 0.99)

Vapour = GRAPH(T_air)
(-30.0, 0.3), (-25.0, 0.48), (-20.0, 0.884), (-15.0, 1.39), (-10.0, 2.14), (-5.00, 3.25), (0.00, 4.85), (5.00, 6.80), (10.0, 9.40), (15.0, 12.8), (20.0, 17.3), (25.0, 23.1), (30.0, 29.5), (35.0, 36.0), (40.0, 43.0)

5. IONS

Ca_retention(t) = Ca_retention(t - dt) + (Ca_input - Ca_leaving) * dt
INIT Ca_retention = 0
INFLOWS:
Ca_input = ((Snow_Throughfall/10)*(Ca_conc_in_snow/1000)
+(Rain_Throughfall/10)*Ca_conc_in_rain/1000)
OUTFLOWS:
Ca_leaving = Ratio_Ca*(Ca_retention/dt+Ca_input)*Adj_Ca_out

H_retention(t) = H_retention(t - dt) + (H_input - H_leaving) * dt
INIT H_retention = 0
INFLOWS:
H_input = ((Snow_Throughfall/10)*(H_conc_in_snow/1000)
+(Rain_Throughfall/10)*H_conc_in_rain/1000)
OUTFLOWS:
H_leaving = RatioH*(H_retention/dt+H_input)*Adj_H_out

NO3_retention(t) = NO3_retention(t - dt) + (NO3_input - NO3_leaving) * dt
INIT NO3_retention = 0
INFLOWS:
NO3_input = ((Snow_Throughfall/10)*(NO3_conc_in_snow/1000)
+(Rain_Throughfall/10)*NO3_conc_in_rain/1000)
OUTFLOWS:
NO3_leaving = Ratio_NO3*(NO3_retention/dt+NO3_input)*Adj_NO3_out

SO4_retention(t) = SO4_retention(t - dt) + (SO4_input - SO4_leaving) * dt
INIT SO4_retention = 0
INFLOWS:
SO4_input = ((Snow_Throughfall/10)*(SO4_conc_in_snow/1000)
+(Rain_Throughfall/10)*SO4_conc_in_rain/1000)
OUTFLOWS:
SO4_leaving = Ratio_SO4*(SO4_retention/dt+SO4_input)*Adj_SO4_out

Adj_Ca_out = 1

Adj_H_out = 1

Adj_NO3_out = 1

Adj_SO4_out = 1

Ca_conc_in_rain = Ca_ppt_Keji_meq_L

Ca_conc_in_snow = Ca_ppt_Keji_meq_L

CA_conc_snowmelt = if (Water_through_snow>0.0001 and Snowdepth_cm>1) then
Ca_leaving/(Water_through_snow/10)*1000 else 0

Ca_ppt_Keji_meq_L = Ca_ppt_Keji_84_00*2/40.03

FC_snow_mm = if (Snowpack-FCv_snow)>0 then FCv_snow*((Snowpack-WE_Out_snow_2)/SnowDens) else 0

H_conc_in_rain = 10^(-pH_rain)

H_conc_in_snow = 10^(-pH_snow)

H_conc_snowmelt = if (Water_through_snow>0.0001 and Snowdepth_cm>1) then
H_leaving/(Water_through_snow/10)*1000 else 0

NO3_conc_in_rain = NO3_ppt_Keji_meq_L

NO3_conc_in_snow = NO3_ppt_Keji_meq_L

NO3_conc_snowmelt = if (Water_through_snow>0.0001 and Snowdepth_cm>1) then NO3_leaving/(Water_through_snow/10)*1000 else 0

NO3_ppt_Keji_meq_L = NO3_ppt_Keji_84_00/62.01

pH_ppt_Keji_84_01 = if time<1988 then pH_ppt_84_87 else

if time<1992 then pH_ppt_88_91 else

if time<1996 then pH_ppt_92_95 else pH_ppt_96_99

pH_rain = pH_ppt_Keji_84_01

pH_snow = pH_ppt_Keji_84_01

pH_snowmelt = if H_conc_snowmelt>0 then -log10(H_conc_snowmelt) else 0

Rain_Throughfall = Rain_through_canopy

RatioH = if (Snowdepth_cm>0.1 and FC_snow_mm>0) then Water_through_snow/(Water_through_snow+FC_snow_mm/dt) else 1

Ratio_Ca = if (Snowdepth_cm>0.1 and FC_snow_mm>0) then Water_through_snow/(Water_through_snow+FC_snow_mm/dt) else 1

Ratio_NO3 = if (Snowdepth_cm>0.1 and FC_snow_mm>0) then Water_through_snow/(Water_through_snow+FC_snow_mm/dt) else 1

Ratio_SO4 = if (Snowdepth_cm>0.1 and FC_snow_mm>0) then Water_through_snow/(Water_through_snow+FC_snow_mm/dt) else 1

Snowdepth_cm = Snowpack/SnowDens/10

Snow_Throughfall = Snow_through_canopy

SO4_conc_in_rain = SO4_ppt_Keji_meq_L

SO4_conc_in_snow = SO4_ppt_Keji_meq_L

SO4_conc_snowmelt = if (Water_through_snow>0.0001 and Snowdepth_cm>1) then SO4_leaving/(Water_through_snow/10)*1000 else 0

SO4_ppt_Keji_meq_L = SO4_ppt_Keji_84_00*2/96.06

Water_through_snow = if Snowdepth_cm>0 then max(0,Wat_ret_input+Water_retention/dt-FC_snow_mm/dt) else Wat_ret_input+Water_retention/dt