Semi-distributed water balance dynamics in a small boreal forest basin

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Abstract

Information on water balance dynamics is an essential component of studies of the role of the boreal forest in surface-atmosphere interactions and climate change. The water balance of a small boreal forest basin in northern Manitoba was examined using a semi-distributed approach to assess basin sensitivity to climate change, provide a framework for distributed hydrological modelling, and explore data aggregation and micro-to-meso scaling of hydroclimatological variables. Black spruce forest with a highly variable canopy density was the main land cover in the basin. Spring snowmelt dominated basin runoff, while summer outputs were largely via evaporation. Annual differences in spring runoff were controlled by variations in snow water equivalent, rainfall timing and magnitude, thaw depth, and antecedent water content in surface stores and upland soils. Water storage in small wetlands and ephemeral surface depressions in the open-canopy black spruce forest and its subsequent loss via evaporation was a fundamental component of the basin water balance. However, its role could be overlooked by inappropriate spatial lumping of landscape units when scaling-up variables or in the production of depressionless digital terrain models. Hydrological consequences of climate warming in this part of the boreal forest include: (i) increased evaporation following spring snowmelt in open black spruce areas; (ii) decreased surface and soil water storage on basin slopes; and (iii) reduced streamflow response to spring runoff and summer and fall rainstorms. © 1999 Elsevier Science B.V. All rights reserved.

Keywords: Boreal forest; Water balance; Snowmelt; Depression storage; Scaling; Boreal Ecosystem-Atmosphere Study

1. Introduction

The boreal forest is a circumpolar ecotone of coniferous and deciduous tree species (predominantly coniferous) covering over 14.7 million km² or 11% of the world’s land mass (Bonan and Shugart, 1989). Forty percent lies within Canada, where it comprises the largest forest region covering an area of ~3 million km² (Zoltai et al., 1988). The extent of the boreal forest ecotone and increased interest in global climate change has focused attention on physical processes in the boreal forest and their potential influence on northern hemisphere and global climatology through exchanges of CO₂, CH₄, energy and water (Bonan and Shugart, 1989; Bonan et al., 1995). This interest has facilitated the Boreal Ecosystem-Atmosphere Study (BOREAS) in Canada (Sellers et al., 1995) and the Northern Hemisphere Land Surface Climate Processes Experiment (NOPEX) in Scandinavia (Halldin et al., 1998). Hydrological processes and their interaction with the active layer in the discontinuous permafrost zone are particularly important since they affect physical, biological, and...
chemical processes controlling biogeochemical cycles and ecosystem feedback mechanisms (Goulden et al., 1998; Running et al., 1999).

The water balance is a fundamental characterization of the regional environment (Dingman, 1973); however, there have been few basin-scale studies of water balance dynamics in boreal forests, especially those focusing on linkages between uplands and wetlands. Most boreal water balance literature is from northern Europe and the former USSR (Goode et al., 1977) and has focused primarily on wetland subsystems. An exception is Rothwell’s (1982) water balance study in northern Alberta. Other North American studies are limited to the work of Bay (1969) and Verry and Boelter (1978) in the temperate zone of northern Minnesota. The lack of boreal water balance research in Canada is surprising given the extent of the boreal forest, its potentially important role in global climate change, and its increased utilization by the forestry sector.

Water balance studies must advance beyond a “black-box” approach to show a region’s sensitivity to climate change, and elucidate the spatial and temporal variation in water balance components within a basin. Such information may indicate areas where process-based research is required and supplies a framework on which to develop spatially distributed hydrologic models. Dolph et al. (1992) note that the capability to simulate the “spatial magnitude and extent of hydrologic processes and properties” is a fundamental requirement for simulating ecosystem response to climate change. New approaches are also required to accommodate boreal forest heterogeneity and the dynamic storage capacities related to permafrost. Woo et al. (1983) emphasised that single point measurements are inadequate to represent basin gains and losses for even a barren arctic landscape with continuous permafrost, and that a shift toward more spatially distributed information was required. Hence, knowledge of processes within, and linkages between, the assemblage of micro-landscapes in the boreal forest is required for successful scaling-up from point (micro) to basin (meso) to regional (macro) scales in an effort to show the sensitivity of the boreal forest water balance to predicted climate change.

The objective of this study is to elucidate the water balance dynamics of a boreal forest basin using a semi-distributed approach to address landscape heterogeneity. We present the spring and spring-to-autumn water balance at basin and sub-basin scales. We relate the water balances to basin hydrologic conditions, and provide a framework for the study of runoff processes and development of a distributed hydrologic model. Micro-to-meso scaling of hydroclimatological measurements are used to derive improved estimates of basin water balances, and methods of data aggregation are discussed.

2. Study area

The study was conducted in a small (2.4 km²) boreal forest basin located 40 km northwest of Thompson, Manitoba (55° 55.5’ N, 98° 25’ W) in a zone of discontinuous permafrost (Fig. 1). The low-relief basin, extending from 253 to 276 m a.s.l. and containing a series of interconnected wetlands (fens), discharges into a large fen (a BOREAS tower flux measurement site). Micrelief between hummocks and hollows in the fens is in the order of 0.2 m. Larger fens become ice-free during the summer while permafrost is maintained under smaller fens and basin slopes. Forest cover is predominantly black spruce (Picea mariana). Lesser amounts of larch (Larix laricina) occur in wet areas, with paper birch (Betula papyrifera) and trembling aspen (Populus tremuloides) in mesic areas, and jack pine (Pinus banksiana) in dry areas. In dry areas where trees do not form a closed-crown canopy, ground cover is dominated by Cladina spp., particularly reindeer lichen (C. rangifera), with an overstory of labrador tea (Ledum groenlandicum). These species are increasingly replaced by feather moss (Pleurozium schreberi), sphagnum mosses (Sphagnum spp.), and some scrub birch (Betula glandulosa) and willow (Salix spp.) in areas with more closed canopies and/or greater soil moisture. The latter are more prominent at wetland fringes. Characteristic species of larger wetlands include sedges (Carex spp.), mosses (Sphagnum spp.), bog laurel (Kalmia spp.), bog birch (Betula glandifera), and buckbean (Menyanthes trifoliata). Wetlands contain typic and humic mesisols, while the hillslopes are mantled by orthic grey luvisols and orthic eutric brunisols. The basin is underlain by glacio-lacustrine sediment and thus disconnected.
Fig. 1. Study basin location and measurement sites.
from the regional groundwater system. Mean annual precipitation for the basin is ca. 0.45 m, mean annual snowfall depth is ca. 1.7 m. January mean daily temperature is ca. \(-25^\circ\text{C}\) and July mean daily temperature is ca. \(16^\circ\text{C}\) (Hydrological Atlas of Canada, 1978).

### 3. Instrumentation and methods

The data were collected from early April to mid September in 1994 and from April to June in 1995. Since freeze-up begins in early to mid October, the 1994 field season does not encompass the full period of open water conditions.

The water balance at a point in a boreal forest basin underlain by discontinuous permafrost is defined by:

\[
M + P - E - R - \Delta S = b
\]  

(1)

where \(M\) is snowmelt as snow water equivalent (SWE), \(P\) the precipitation, \(E\) the evapotranspiration, \(R\) the runoff and \(\Delta S\) is the change in storage (both ice and water). The calculation of \(\Delta S\), as opposed to its determination as the residual in the water balance (\(\Delta S_R\)), enables an estimate of \(b\). \(b\) is the divergence from 0 and represents the net cumulative error in measurement of all water balance components, given that some errors may be compensating (Dooge, 1975).

Basin runoff integrates spatially variable physical processes controlling changes in the magnitude of water balance components. As landscape heterogeneity increases, so too does the difficulty and importance of measuring these spatially variable water balance components accurately. Accurate estimates of \(E\) and \(\Delta S\) are particularly difficult given the mosaic of wetlands, uplands and areas of depression storage. Spatial heterogeneity of \(E\), \(\Delta S\), and \(M\) were addressed within a semi-distributed framework by associating these variables with vegetation units. Vegetation characteristics were derived from an analysis of canopy structure using Landsat Thematic Mapper (TM) data (Metcalfe and Buttle, 1998) and from a digital Forest Resources Inventory (FRI) image (Table 1). The latter image was produced by digitizing Manitoba Natural Resources Forest Management (MNR) maps into an ARC/INFO vector coverage and gridding the data at a 30 m pixel resolution to produce the FRI image (Knapp and Tuinhoff, 1998). The MNR maps were produced at a scale of 1:15,840 using 1988 aerial photography. The TM image was used to classify vegetation as wetland (W), wetland with willow (WWW), open forested (OMF) (primarily black spruce with some mixed forest), and closed forested

<table>
<thead>
<tr>
<th>Landcover characteristics</th>
<th>Total basin</th>
<th>Sub-basin 1</th>
<th>Sub-basin 2</th>
<th>Organic layer thickness (m)</th>
<th>Thaw depth sites</th>
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<tr>
<td><strong>Thematic Mapper (TM) classification</strong></td>
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<td>2700</td>
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<tr>
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<td>67 500</td>
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<td>82 800</td>
<td>442 800</td>
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<td>94 500</td>
<td>603 900</td>
<td>53.0</td>
<td>1, 8, 9</td>
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<td>191 700</td>
<td>1 138 500</td>
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<td><strong>Forest resources inventory (FRI)</strong></td>
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<td>Black spruce</td>
<td>638 105</td>
<td>105 301</td>
<td>159 301</td>
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<td>Black spruce /jack pine</td>
<td>65 701</td>
<td>6 701</td>
<td>65 701</td>
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<tr>
<td>Black spruce /jack pine/trembling aspen</td>
<td>236 702</td>
<td>236 702</td>
<td>20.8</td>
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<td></td>
</tr>
<tr>
<td>Trembling aspen/black spruce</td>
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<td>9000</td>
<td>9000</td>
<td>0.8</td>
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<td>Treed muskeg</td>
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<td>70 201</td>
<td>612 005</td>
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<td>Flooded land</td>
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<td>14 400</td>
<td>55 800</td>
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<td>0.00</td>
<td></td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td>2 386 820</td>
<td>189 902</td>
<td>1 138 509</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
(MF) (black spruce and mixed forest), ‘open’ and ‘closed’ being qualitative measures of a site’s relative canopy density. Areas disturbed by road construction (as identified in the FRI) were re-classified as OMF and MF.

Sharp-crested 90° v-notch weirs and water level recorders were installed at the basin outflow and outflows of two sub-basins (SB1 and SB2) to estimate R. Mean water levels were recorded on Campbell Scientific CR10 data loggers at 15 min intervals. Water levels were rated against flow measurements using a stopwatch and container of known volume, or current meters, depending on discharge. Manual water level measurements were also taken from stage boards beside the stilling wells.

Spatial distribution of SWE was determined prior to spring melt by a 62-point snow survey in 1994 and a 64-point snow survey in 1995 using a stratified sample based on the TM vegetation classification (Metcalf and Buttle, 1998). Snowmelt was measured daily at six sites of varying canopy density, and daily regression relationships between a site’s daily snowmelt and its canopy density (derived from the TM image) were estimated. Canopy density of each pixel was then substituted into the regression equation to provide a spatially distributed estimate of daily snowmelt (Metcalf and Buttle, 1998). M was calculated by totalling mean daily snowmelt for each basin.

Precipitation (P) was measured at the Fen tower and three locations in the basin (Fig. 1) using tipping bucket rain gauges connected to Campbell Scientific CR10 data loggers and recorded as 15 min totals. Daily precipitation for each gauge was weighted using an inverse-distance-squared interpolator with a 30 m grid cell resolution to obtain mean daily precipitation for each basin. This algorithm was selected because it is an exact interpolator that preserves the original point values (Lam, 1983), and because of the absence of any physical controls on precipitation over such a small area that would require more sophisticated methods.

Daily E measured at the Fen and Old Black Spruce (OBS) tower flux sites was scaled-up to the wetland (W, WWW) and upland (OMF, MF) areas, respectively, for calculation of weighted mean daily E for the basin and each sub-basin. The TM classification was used for the extrapolation since the FRI only classified the more contiguous flooded land in the central valley bottom as wetlands. The bulk transfer approach (Heron and Woo, 1978; Price, 1988) was used to estimate E at the Fen from the initiation of springmelt to when discontinuous snow cover (areal albedo = 0.37) prevented further measurement (Metcalf and Buttle, 1998). Estimates of E were then obtained by the Bowen Ratio energy balance approach (Lafleur et al. 1997) and the aerodynamic method. The latter was used to estimate sensible heat flux (\( Q_H \)) and calculate \( E \) indirectly by the energy balance approach when freezing air temperatures prevented use of wet-bulb psychrometers. The assumption of equality of transfer coefficients used in the bulk transfer and aerodynamic approaches could result in a systematic over-prediction of \( E \) upwards of 50% (Male and Granger, 1979); however, their use during a brief period when \( E \) was small would minimize the potential error. Relative importance of the error would be greatest in springmelt water balance calculations but would be comparable between years. \( E \) was measured directly at the OBS tower from eddy correlation (EC) methods but contained significant data gaps that were bridged using the Priestley–Taylor approach (Priestley and Taylor, 1972). On days with few missing values (~20% of the daily total number of values) the latent heat flux (\( Q_E \)) and equilibrium evaporation (EQE) were derived from linear interpolation. The coefficient of evaporability (\( \alpha \)) was then determined from daily \( Q_E \) and EQE totals:

\[
\alpha = \frac{Q_E}{EQE}
\]
basin ($r^2 = 0.99$, s.e. = 0.5°C). This enabled the calculation of EQE (W m$^{-2}$):

$$\text{EQE} = \frac{s}{s + \gamma} (Q^* - Q_G) \quad (3)$$

where $s$ is the slope of the saturation vapour pressure vs. temperature curve (Pa °C$^{-1}$), $\gamma$ the psychrometric constant (Pa °C$^{-1}$), $Q^*$ the net radiation (W m$^{-2}$), and $Q_G$ the ground heat flux (W m$^{-2})$. $Q_G$ was assumed to be 5% of $Q^*$ (Jarvis et al., 1997). Daily $E$ was determined by substituting EQE in Eq. (2), solving for $Q_E$, and dividing by latent heat of vaporization (2.48 MJ kg$^{-1}$ at 10°C).

Soil characteristics for calculation of $\Delta S$ were obtained from field observations, other BOREAS field investigations and the literature. The typical soil profile in the basin consisted of an organic peat layer (mean porosity [$\phi$] = 0.813 ± 0.087 s.d., $n = 7$) overlying a less-permeable glacio-lacustrine silty clay layer ($\phi$ = 0.384 ± 0.077 s.d., $n = 12$). On steeper slopes these layers were sometimes separated by an open-worked, botryoidal layer ($\phi$ = 0.536 ± 0.052 s.d., $n = 15$) caused by flocculation during freeze–thaw cycles (Pitty, 1979). A similar soil profile was observed at the OBS site (Nijssen et al., 1997). Difficulty in estimating spatial extent or depth of the botryoidal soil layer suggested a simplified profile consisting of a peat layer and a silty clay layer with $\phi$ of 0.46. Organic layer thickness for the vegetation units recorded during soil sampling and well installation are shown in Table 1.

Since perched wetlands in the basin are separate from the regional groundwater system, $\Delta S$ represents the storage change in the saturated, surface-water reservoir $\Delta S_w$ (small peat-filled depressions and larger wetlands), and the storage change in the unsaturated soil–water reservoir $\Delta S_s$ (upland slopes). Water level recorder 4 (WLR4) in a small wetland (Fig. 1) was used to estimate $\Delta S_w$:

$$\Delta S_w = \theta_{sy} \Delta d A_{w+www} \quad (4)$$

where $\theta_{sy}$ is specific yield, $\Delta d$ is change in water level, and $A_{w+www}$ is the fraction of the basin occupied by wetland and wetland-with-willow. WLR4 became inoperable on DOY 227 in 1994 and a manual reading on DOY 254 was used to estimate $\Delta S_w$ for the spring-to-autumn balance. $\theta_{sy}$ is defined as:

$$\theta_{sy} = \phi - \theta_c \quad (5)$$

where $\phi$ = 0.813 and $\theta_c$ is volumetric water content at field capacity (Boelter, 1968). We used $\theta_c$ of 0.6 in Eq. (5) (Nijssen et al., 1997). This is slightly less than values reported for partly decomposed/decomposed peat (e.g. Boelter, 1975; Dooge, 1975; Roulet and Woo, 1986) but coincides with a measured $\phi$ that was also slightly less than reported values. Observed $\phi$ was, however, within the range of reported values (0.807–0.952; Walmsley, 1977). The $\theta_{sy}$ value of 0.21 is consistent with values reported for partially decomposed peat (Dooge, 1975).

Vertical profiles of volumetric water content ($\theta$) were measured in nine access tubes along two upland slopes using a neutron probe (Fig. 1). Measurements were taken at a vertical interval of 0.2 m, and access tube depths were variable with a maximum depth of 1 m. Maximum count ratios (count at any depth/background count) for the two soil layers were assumed to represent saturated conditions ($\theta_{sat}$), where

Maximum Count Ratio $\equiv \theta_{sat} \equiv \phi \quad (6)$

All count ratios were expressed relative to these conditions in order to estimate volumetric soil water content at time $t$ ($\theta_t$). Changes in $\theta$ at a given depth were recorded as differences from initial base-line values ($\theta_b$) following Price and Hendrie (1983). $\Delta S_s$ at time $t$ was calculated as:

$$\Delta S_s = (\theta_t - \theta_b) z_t A_{OMF+MF} \quad (7)$$

where $z_t$ is depth to the frost table at time $t$, and $A_{OMF+MF}$ is fraction of the basin consisting of uplands. The frost table acted as the lower boundary for water balance calculations and deep seepage was assumed to be zero. Frost table depth was determined by inserting a metal rod into the ground. Frost table measurements were sporadic in 1994 but were taken on a 2–3 day interval in 1995 at 10 sites extending up the centre of the basin (Fig. 1). Measurements were taken at each site at a central location and 2 m away in each cardinal direction to calculate mean thaw depth (Fig. 2). Sampling sites were classified based on their vegetation unit (Table 1) and daily mean frost table depth was calculated for each unit (Fig. 3). Linear regressions between the 0°C isotherm depth at the forest tower and mean frost table depth for each vegetation unit were used to estimate isotherm movement in 1994 using soil temperature profiles (Fig. 4). The 0°C isotherm depth was interpolated by
Fig. 2. 1995 mean thaw depth measurements with vertical bars showing range of measurement.
Fig. 3. Mean thaw depths for each vegetation unit.

Fig. 4. Relationships between vegetation unit thaw depth and the 0°C isotherm depth at the forest tower.
kriging using mean daily soil temperatures recorded with copper–constantan thermocouples (±0.3°C) at 0.01, 0.05, 0.1, 0.25, 0.5, 0.75 and 1 m below the surface (Fig. 5). When the 0°C isotherm passed below the depth of measurement, its position was estimated by extrapolating the logarithmic temperature gradient calculated on a 5-day interval using only those measurement points below the reach of the diurnal temperature wave (≥ 0.25 m). Isotherm depths for intermediate days were estimated using linear interpolation. Frost table measurements were also used to calculate total water storage capacity of the unsaturated active layer at time \( t (SC_t) \):

\[
SC_t = z_t \phi
\]  

(8)

A digital terrain model (DTM) at a grid cell resolution of 30 m was used to delineate basin divides. Spot heights for the DTM were established using a digital video plotter (DVP) and 1:15,840 digital stereo photographs scanned at 1000 dpi (dots per inch). The DTM was produced through kriging.

4. Results


Water balance computations extended 20 days longer in 1995 owing to a delayed runoff peak and a precipitation event on the recessional limb of the springmelt hydrograph (Fig. 6). However, a comparison of spring water balances showed significantly greater runoff from the basin in 1994 relative to 1995 (Fig. 7 and Table 2). Runoff in 1994 was 86 mm compared to 25 mm in 1995. The difference in runoff can be explained by differences in SWE, pre-melt soil moisture deficits, surface storage capacities, and rainfall input. Rainfall timing relative to thaw conditions of the soil is particularly important.

Initial SWE was smaller in 1995 relative to 1994, while the 1995 pre-melt soil moisture deficit was greater. Soil water contents in the fall of 1994 (DOY 265) in the peat and silty clay horizons were 20 and 3% below \( \theta_{fc} \), respectively. In the spring of 1995 (DOY 102) soil water in the peat was at \( \theta_{fc} \) while the silty clay layer was 16% below \( \theta_{fc} \). The changes in \( \theta \) during the winter suggested movement of water from the silty clay layer towards the freezing front as it developed in the peat, and infiltration of rainfall in the fall of 1994 (DOY 265 to freeze-up, DOY 294) and snowmelt. Precipitation during this period at Thompson was 14.2 mm.

Soil water contents in the fall of 1993 prior to the 1994 snowmelt were assumed to be high, although data were not available. Rainfall recorded in Thompson in the fall of 1993 was 112 mm, approximately twice that in 1994. This suggested \( \theta \) was closer to field capacity at the start of the 1994 snowmelt period and
Fig. 6. Water balance components for (a) 1994 and (b) 1995.
Fig. 7. Cumulative water balance components and total soil storage capacity for (a) 1994 and (b) 1995.
Table 2
Water balance calculations in mm of water

<table>
<thead>
<tr>
<th>Snowmelt</th>
<th>Precipitation</th>
<th>Runoff</th>
<th>% of input</th>
<th>Evaporation</th>
<th>% of input</th>
<th>Depth thaw</th>
<th>Storage capacity</th>
<th>Storage ( \Delta S )</th>
<th>Inputs–outputs</th>
<th>( b )</th>
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<tbody>
<tr>
<td>( M )</td>
<td>( P )</td>
<td>( R )</td>
<td>( E )</td>
<td>( z )</td>
<td>( SC )</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(1)</td>
<td>(2)</td>
<td>(3)</td>
<td>(4)</td>
<td>(5)</td>
<td>(6)</td>
<td>(7)</td>
<td>(8)</td>
<td>(9)</td>
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<td>(12)</td>
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1994

**Spring**
DOY 102 to DOY 147

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<th>Precipitation</th>
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<th>% of input</th>
<th>Evaporation</th>
<th>% of input</th>
<th>Depth thaw</th>
<th>Storage capacity</th>
<th>Surface storage ( \Delta S_w )</th>
<th>Soil storage ( \Delta S_r )</th>
<th>( \Delta S_s )</th>
<th>Inputs–outputs</th>
<th>( b )</th>
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<tr>
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<tr>
<td>SB 2</td>
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<td>61</td>
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<td>11</td>
<td>39</td>
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**Summer**
DOY 148 to DOY 254

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<th>Evaporation</th>
<th>% of input</th>
<th>Depth thaw</th>
<th>Storage capacity</th>
<th>Surface storage ( \Delta S_w )</th>
<th>Soil storage ( \Delta S_r )</th>
<th>( \Delta S_s )</th>
<th>Inputs–outputs</th>
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<tr>
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<td>−44</td>
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**Spring-to-autumn**
DOY 102 to DOY 254

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<th>% of input</th>
<th>Evaporation</th>
<th>% of input</th>
<th>Depth thaw</th>
<th>Storage capacity</th>
<th>Surface storage ( \Delta S_w )</th>
<th>Soil storage ( \Delta S_r )</th>
<th>( \Delta S_s )</th>
<th>Inputs–outputs</th>
<th>( b )</th>
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<td>−38</td>
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1995

**Spring**
DOY 101 to DOY 167

<table>
<thead>
<tr>
<th>Basin</th>
<th>Snowmelt</th>
<th>Precipitation</th>
<th>Runoff</th>
<th>% of input</th>
<th>Evaporation</th>
<th>% of input</th>
<th>Depth thaw</th>
<th>Storage capacity</th>
<th>Surface storage ( \Delta S_w )</th>
<th>Soil storage ( \Delta S_r )</th>
<th>( \Delta S_s )</th>
<th>Inputs–outputs</th>
<th>( b )</th>
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<td>145</td>
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</tbody>
</table>
that surface storage capacity was reduced. In contrast, there was less snowmelt in 1995, a greater proportion of which was required to satisfy the soil moisture deficit of the thawed layer before being transmitted down slope to storage areas that also had a comparatively larger storage capacity. Total potential storage capacity only approached cumulative input for an extended period beginning 5 June (DOY 156) in 1994, but surpassed cumulative input as early as 7 May (DOY 127) in 1995 (Fig. 8). Water loss via $E$ during snowmelt was comparable in 1994 and 1995 (18 and 19 mm, respectively), but was 36 mm greater during the snow-free period in 1995 relative to 1994. Hence, much of the initial meltwater in 1995 contributed to the soil moisture deficit and storage and more was lost to $E$.

Water balances following springmelt in both 1994 and 1995 indicate that significant inputs were unaccounted for (Table 2). This is demonstrated by calculating storage as the residual of the water balance equation ($S_R$) in Table 2 (i.e. inputs–outputs, see shaded area in Fig. 8). The error term ($b$) in Eq. (1) is simply the difference between $S$ and $S_R$. $R$ may have been underestimated by $\sim 10\%$ due to brief bypassing of the SB2 weir during the spring hydrograph peak; however, visual observations suggested that $b$ was dominated by surface depression storage in areas not classified as wetlands in either vegetation classification, and therefore not included in the calculation of $\Delta S$. This included ephemeral depression storage in non-wetland areas and storage in smaller wetlands below the resolution of the imagery used for classification. This added storage was either eventually lost via $E$ (thus contributing to the error in estimation of $E$ for later periods), subsequently drained to the outflow as thaw depth and interconnectivity of storage areas increased, or remained in storage.

4.2. Hydrologic conditions influencing springmelt water balance dynamics

The magnitude and rate of water transmission down slope depends on soil conditions. The persistent wetness of the underlying silty clay and the contrasting porosities at the peat/silty clay interface promote formation of a saturated soil layer and surficial ponding. This can result in the formation of an ice-rich layer during downward progression of the freezing front, aided by migration of water towards the freezing front from the underlying soil (Stein et al., 1994). Ice layer formation could begin in late autumn from water present at the interface during initial freezing or from late rainfall after the start of freeze-up but before snowpack formation. Conditions supporting ice layer formation and/or thickening would be enhanced after freezing since the wetness of the clay, and subsequent formation of ice within the soil matrix, would decrease empty pore space. Conversely, desiccation of the overlying peat could increase air-filled pore space. The empty pore space of the frozen peat layer will vary throughout the winter as ice is added by refreezing of infiltrating meltwater and lost by sublimation into the overlying snowpack. Freeze-thaw cycles during springmelt with sufficiently low air temperatures may also contribute to ice lens thickening. Stein et al. (1994) found that minimum air temperatures of $-5^\circ C$ were insufficient for infiltrating meltwater to contribute to thickening of concrete frost during springmelt; however, we hypothesised that minimum diurnal temperatures between $-10$ and $-15^\circ C$ and extended daytime periods of freezing temperatures, such as those in 1994, would provide favourable conditions for ice layer thickening.

In both years the thawing front slowed from 8–24 May (DOY 128–144) at a depth of $\sim 0.15–0.25$ m (location of the peat/silty clay interface, Fig. 9). The change in thaw rate at the interface was likely caused by the presence of an ice-rich layer. In 1994, 49 mm of rain fell within this period, 6 mm in the two days after the basin became snow-free. Heat contributions from rainfall during this period were 6.4 times greater than that required to account for the observed decrease in thaw depth, assuming the heat capacity of a mineral soil at $\theta_{fc}$, rain temperature equal to air temperature, and a frozen soil temperature of $-1^\circ C$. Energy inputs were also not accounted for by increases in soil temperature (Fig. 5). Hence, the lack of thermal response to these energy contributions shows little infiltration occurred in the mineral soil. Transmission of this water to the basin outflow depended on the holding capacity of the intervening storage areas. Rapid streamflow response to inputs in 1994 (Fig. 6) supports the hypothesis that storage areas were close to capacity. In contrast, only 19 mm of rain fell during this period in 1995, with the first rainfall (only 2 mm) occurring 8 days after the basin became snow-free.
Fig. 8. Cumulative water balance inputs, outputs, storage ($S_R$), and soil storage capacity for (a) 1994 and (b) 1995.
Fig. 9. Comparison of (a) degree days, climatological variables measured at the Fen tower in (b) 1994, (c) 1995 and (d) thaw depth at the forest tower measured in 1995 and estimated in 1994.
The first significant rainfall (DOY 154–156) occurred after the thawing front had penetrated the silty clay layer. Rainwater infiltration was indicated by a distinct increase in soil temperature (Figs. 5 and 9) and was supported by a 9% increase in $\theta$. Infiltration of water to lower soil layers with significantly smaller hydraulic conductivities also results in slower movement of water down slope. Thus, larger inputs into a soil layer of high hydraulic conductivity at a time of minimum storage capacity, combined with smaller $E$, generated more surface runoff in 1994 relative to 1995.

The effectiveness of infiltrating rainwater in warming the soil, penetrating ice layers, and thickening the thaw layer is a function of climatological conditions and soil thermal characteristics. Greater active layer thickening in 1995 (Fig. 9), given less latent heat input from rainfall, could have also been enhanced by excessive drying of the peat in the summer of 1994. This would result in reduced thermal conductivity and an insulation effect that reduced ground cooling during winter (Brown, 1977). This would have been exacerbated by the milder winter that followed ($-2631^\circ$ days in the 1994/1995 winter compared to $-3363^\circ$ days in the 1993/1994 winter). Excessive drying would also result in higher infiltration capacities in 1995 by lowering ice content within the silty clay matrix.

The greatest influence of interannual differences in active layer development on potential storage capacity occurs in the upland portions of the basin, since the higher thermal conductivity of wet areas precludes permafrost formation at this latitude. Thus, ice lenses in larger wetlands quickly became discontinuous during springmelt (DOY 140–143, 1995) and thawed completely by the end of melt in both years. Conversely the uplands, carpeted by a dry insulating moss layer, maintained permafrost at depth.


The runoff pattern from spring to early autumn is characteristic of other nival peatland basins (e.g. Bay, 1969; Woo et al., 1983): (1) snowmelt runoff dominates the annual hydrograph; (2) $E$ becomes an increasingly important water balance component as summer progresses; (3) storage capacity is increased by $E$ and advance of the frost table such that only large rainfalls produce a detectable increase in runoff; and (4) extended recession limbs follow steep discharge peaks generated by large summer rainfalls, indicating that peatland basins are effective short-term storage areas that subsequently release water during protracted periods.

Snow accounted for 33% of water input to the basin and contributed to a spring runoff that supplied 65% of the study period water loss during 43 days (28% of the period). Although 49% of total precipitation ($M + P$) fell during the summer, only 3 rainfalls (DOY 150–9 mm, DOY 165–39 mm, and DOY 186–53 mm) produced a significant hydrograph response. The quick response and long recession typical of the saturated areas is shown by WLR4 (Fig. 6). Other rainfalls contributed only to $S$ and $E$. $E$ was the dominant water-loss component. The pattern of $E$ followed that described for a boreal forest basin in southern Manitoba (Amiro and Wuschke, 1987), increasing to a maximum at summer solstice when $Q^*$ was greatest, water was generally available, and vegetation was fully developed. Reduced $E$ during springmelt resulted in partitioning of excess inputs between $S$ and $R$, with the degree of partitioning dependent on antecedent soil moisture and surface storage deficits. During summer months $E$ and $P$ were of similar magnitude and runoff was sustained by slow depletion of $S$. $E$ gradually decreased in late summer/early autumn as $Q^*$ was limited and $S$ was replenished by excess $P$.

There were significant differences between spring-to-autumn water balances of the basin and its subbasins (Table 2 and Fig. 7). These are attributable to variations in $R$, since the other water balance components associated with vegetation classes do not vary significantly between sub-basins (Table 1). Runoff ratios were greatest in SB1 and were similar between spring and summer periods (0.49 and 0.43, respectively). Cumulative $R$ was greater than cumulative $E$ until 9 August (DOY 221), after which rainfall inputs to the basin decreased. The runoff ratio was much less in SB2 (0.14) and $E$ dominated outputs. Variations in $R$ are a function of differences in basin physiography and its control on surface depression storage, particularly those stores not accounted for in calculation of $S_w$ (see above).

There was good closure of the summer water
balance in SB1; however, $\Delta S_R$ remained greater than $\Delta S$ in the basin and SB2. This is also partly the result of differences in physiography and temporary storage of water in surface depressions; however, possible water loss via $E$ fluxes greater than those measured at the Fen or OBS towers is more significant. Most of the error would originate in forested areas of the basin. Although a significant proportion of forested areas have open canopies with thick sphagnum ground cover, Price et al. (1995) found that interception and subsequent evaporation from a moss layer in a boreal forest black spruce stand was similar to canopy interception. Many smaller summer rainfalls in the basin failed to infiltrate through the moss layer. Ephemeral depression storage areas and smaller wetlands were located in the hummocky terrain of the open black spruce where water was also close to the surface in micro-scale hollows. These surface storages explain the treed muskeg classification of these areas in the FRI. Lafleur (1992) found similar open-forested areas at the tree line transition in northern Manitoba increased within-canopy ventilation and permitted greater runoff ratios, and less error in scaling-up of $E$ from the OBS tower to upland portions of the basin. This suggests that transit times within SB2 are longer than for SB1, thus providing greater opportunity for evaporative loss. SB1 contained the largest proportion of treed area and the least treed muskeg in the FRI classification (Table 1), suggesting better drainage, greater runoff ratios, and less error in scaling-up of $E$ from the OBS tower to upland portions of the basin. Field verification confirmed that part of the area classified as OMF in SB1 was open-canopy mature black spruce stands on well-drained slopes. This more closely resembled conditions at the OBS tower in contrast to the wetter open black spruce sites, and supports the suggestion of less error in scaling-up of $E$ for this sub-basin.

4.4. Sources of error in water balance terms

Roulet and Woo (1986) assumed a water balance error for a small sub-arctic basin of 15% derived from estimated errors for the individual components. We assume an error of 15–20% based on the estimated accuracy of $Q_E$ measurements (McCaughey et al., 1998). This is comparable to the range of total error proposed by Winter (1981) using uncertainties in estimating individual water balance components for lakes. Other potential sources of error associated with methodologies used here include:
4.4.1. $\Delta S_W$

The porosity of peat can decrease considerably with depth due to greater rates of decomposition (Boelter, 1965; Dooge, 1975). Therefore, $\theta_{soil}$ will exhibit a similar decrease (Goode et al., 1977). However, the constant $\theta_{soil}$ used in this study across the entire range of water table fluctuation is reasonable in that: (1) water table fluctuations were limited to the upper peat layer; and (2) the mean peat $\phi$ was calculated from samples covering this depth range. Spatial variations in water table position caused by microtopographic differences and the hypsometric position of WLR4 within the basin contribute to potential error in $\Delta S_W$ (Goode et al., 1977). At 259 m a.s.l., WLR4 is 3 m below the weighted mean elevation of the basin. A better estimate of average water table changes would be obtained from records for several wetlands throughout the basin (e.g. Bavina, 1975), and would improve our understanding of storage dynamics in treed muskeg areas. Neglecting drainage of the peat layer immediately above the water table to below $\theta_{fc}$ in 1994 also contributed to potential error in $\Delta S_W$. However, greatest uncertainty in the estimation of $\Delta S_W$ results from the inability to identify smaller surface storage areas using the TM imagery and, to a lesser extent, underestimating the surface area of smaller wetlands that expand during springmelt inundation.

4.4.2. $E$

Evaporation decreases noticeably as the water table level drops below ground surface (Dooge, 1975). Thus, there is potential error in extrapolating $E$ from tower flux sites to the basin if water table levels in the respective landscape units are significantly different from those at the towers.

4.4.3. $z$ and estimation of $\Delta S_S$

There was considerable variation in thaw depth at each site in 1995, primarily the result of spatial variability in microtopography and moisture conditions. Active layer development will be less under hummocks than in hollows because the reduced thermal conductivity of dry peat impedes warming of the underlying soil (Brown, 1977). Therefore, some error in $\Delta S_S$ will occur when mean thaw depth is used as the lower boundary of the soil storage zone. The error will be smallest within the silty clay layer because of its small porosity.

4.4.4. $\Delta S_S$

Measurement of $\theta$ using the neutron probe assumes that contributions to the maximum count rate (Eq. (6)) from the PVC access tubes and organics is constant (i.e. the “background” level at saturation) and that change in the count rate is attributed only to variations in $\theta$. However, loss of intracellular water upon drying of peat ($\theta < \theta_{ic}$) would result in the underestimation of $\theta$ and contribute to error in $\Delta S_S$. Significant errors in the methodology were not observed (i.e. $\theta$ estimates that were negative or greater than $\phi$). Another potential error includes loss of neutrons at the soil surface during dry conditions, which would lead to underestimation of $\theta$ in the peat. Nevertheless, the water balance is relatively insensitive to error from the calculation of $\theta$ for peat, since a 20% measurement error in $\theta$ results in only a $\pm 6$ mm change in the estimate of $\Delta S_S$ for the spring-to-autumn water balance.

5. Discussion

Measurement of all water balance components, particularly $S_W$ and $S_S$, has provided insight into water balance and runoff dynamics of the boreal forest that may have been undetected had storage been calculated as the water balance residual (i.e. $\Delta S_R$). Visual observations showed that during springmelt there was considerable surface storage in areas not identified in the calculation of $\Delta S_W$. During this period, $\Delta S_R$ provided a better estimate of storage contributions and insight to the basin’s total storage capacity, especially since water loss to $E$ was small at this time. However, $\Delta S_R$ overestimated spring-to-autumn storage contributions, evidenced by the decrease and/or absence of surface water in these same areas at the end of summer and the water deficit calculated for $\Delta S$. The greatest overestimation occurred in areas with considerable depression storage (e.g. SB2). Divergence of $\Delta S$ and $\Delta S_R$ would also be expected given the underestimation of $\Delta S$ caused by the under-representation of smaller saturated areas. Evaporation from these areas would accentuate our underestimation of true $E$. Hence,
this semi-distributed approach suggests that small surface storages, and their contribution to $E$, are a key component in boreal forest water balance dynamics.

Scaling-up from point measurements of spatially variable, complex processes to meso- and macro-scale models often involves relating such measurements to landscape features that can be lumped spatially and treated as homogeneous units (Amiro and Wuschke, 1987). A priori selection of homogeneous landscape features such as contiguous units of closed canopy forest (old black spruce, old jack pine, young jack pine) and larger fens, as was done in the BOREAS NSA, can overlook important areas such as open black spruce forest or treed muskeg. FRI data showed that treed muskeg (39%) and black spruce (23%) dominated landscape communities within an area encompassing all BOREAS NSA tower flux sites and most of the modelling sub-area (1710 km², see Fig. 1), with flooded land (6%) as the next major surface type. Extrapolating $E$ estimates from the closed-canopy OBS tower flux site to the more-common open-canopy black spruce forests would produce considerable errors in regional water balance calculations, since the latter forests tend to have relatively larger $E$ fluxes due to their surface roughness, canopy ventilation, and proximity of water to the ground surface.

Surficial water in the treed muskeg landscape was stored in three major landscape units.

1. **Ephemeral storage in shallow depressions:** these areas are inundated during springmelt and are usually open water surfaces. They are not connected to the drainage system because of shallow active layer development. All water is lost via $E$ at potential rates during spring and early summer. The water holding capacity of these areas at the beginning of springmelt exhibits little annual variation. Possible fluctuations are caused by the presence of standing water at freeze-up, reducing the volume of the depression.

2. **Perennial wetlands:** these reach maximum capacity during springmelt when the water table is often above the peat surface. Larger wetlands are continually connected to the drainage system, whereas some smaller wetlands are linked only at high water levels while others occupy disconnected clay-lined depressions. They drain slowly, and $E$ dominates water losses during summer. The water level falls beneath the peat surface following springmelt and, although replenished by summer rain, rises above the surface for short periods only during the largest rainfalls (see WLR4, Fig. 6). $E$ proceeds at potential rates when the water level is near or above the surface and near-potential rates at other times. The water holding capacity at the start of springmelt exhibits annual variation, with greater fluctuation associated with decreasing wetland size.

3. **Peat filled hummocky terrain:** this is the dominant sphagnum landscape in the open black spruce areas. Water storage occurs in hollows, perched on the silty clay layer. The hollows may be connected to the drainage network only during high water levels. The water holding capacity of these microstorages at the start of springmelt may vary annually.

Our inability to recognize ephemeral storages, microstorages, and smaller perennial wetlands is reflected in the error term of the water balance calculations ($b$), the magnitude of which indicates their significance. The interannual variation in $b$ is controlled primarily by antecedent water levels in the microstorages and smaller wetlands, since the ephemeral storage capacities remain largely unchanged. These storages behaved predictably during springmelt of 1994 and 1995 in that $b$ was larger in 1995 when water levels in surface stores were lower than in 1994.

These three forms of storage contribute to the heterogeneity of the treed muskeg landscape at the micro-scale; however, treed muskeg is an identifiable community in the landscape at the meso-scale with different physical processes from those of other surface types. The challenge is to collect point data that would best characterize this heterogeneous community and establish a methodology for aggregating these areas in the landscape to permit up-scaling of measurements for meso- and macro-scale water balance estimates. Although some of the heterogeneity was detected using our classification of the TM data, ephemeral storage areas, microstorages, and many small wetlands were not identified at the 30 m grid cell resolution. IGBP (1998) identified similar
problems in representing the functionality of wetland landscapes while balancing sensor utility with the feasibility of data acquisition. For the purpose of scaling-up values of $E$ for water balance modelling in the BOREAS NSA, it seems reasonable to aggregate these communities using FRIs treed muskeg classification and applying $E$ estimates obtained from the Fen tower flux site.

Maintaining the physiographic integrity of these areas within data structures of distributed hydrologic models for the boreal forest is important, since the antecedent storage capacity largely determines snowmelt and rainfall contributions to runoff. Therefore, any DTM used within a distributed hydrologic model must contain the most detail logistically possible to represent sufficiently the surface storage components within these units. Rodhe and Seibert (1996) used the topographical index from TOPMODEL (Beven and Kirkby, 1979) with a digital 1:50,000 topographic map to predict the wetness pattern (and inferred mire location) for four catchments in central Sweden. They instead found ground slope provided a better estimate of mire occurrence. IGBP (1998) suggest that a vertical discrimination of better than 1 m is required to represent the functionality of wetland landscapes. The production of a “depressionless, hydrologically correct” DTM would not be physically realistic within the northern boreal forest environment since storage of water in surface depressions, often unconnected to the drainage network, is a fundamental component of water balance dynamics.

Runoff generation is also highly dependent on thaw depth, given its control on storage capacity and drainage network connectivity. Since thaw depth is influenced by variations in topography and soil moisture, soil wetness indices derived from a topographical index (e.g. Beven and Kirkby, 1979) might provide a more physically realistic method for scaling-up thaw depth measurements than the semi-distributed method based on vegetation cover used here. Better methods for determining the spatial variation in peat thickness would also be useful.

5.1. Implications for boreal forest water balance dynamics under a warming climate

The 1994 study period was warmer and significantly drier than the long-term climate normals (Lafleur et al., 1997). Mean daily temperatures from April to September averaged 2°C above normal. Total precipitation was 158 mm below normal with August and September rainfall averaging 45 mm below normal. Simulation studies generally predict warmer, drier climate in continental interiors of higher latitudes (45–65°N) with increasing CO$_2$ (Sellers et al., 1995); hence, hydrologic conditions could be similar to those observed in the summer of 1994 and the spring of 1995. Open black spruce areas would be most sensitive to climate warming. Reduced water input and increased evapotranspiration would decrease surface and soil water storages and would impact on biogeochemical cycles. It would also result in reduced streamflow response to spring runoff and summer and fall rainstorms. Decreases in the extent or thickness of ice layers at the peat/silty clay interface or ice volume within the soil would facilitate faster progression of the thawing front, also diminishing spring runoff response. The importance of these freeze–thaw transitions on regional and global climate, hydrology, and biogeochemistry is receiving increased attention (Running et al., 1999).

6. Conclusions

This study revealed runoff processes similar to those in other nival regime, peatland basins in both permafrost (Woo et al., 1983) and non-permafrost (Bay, 1969) environments. Springmelt dominates annual runoff while evaporation dominates basin outputs during summer months, resulting in a steadily decreasing storage component that is only replenished by large rainfalls. The magnitude of spring runoff is controlled by the amount of SWE, occurrence and timing of rain events, thaw depth, and antecedent soil moisture conditions on upland slopes and, more importantly, in surface storage areas. The latter are predominantly located in the poorly drained, open black spruce areas which also contain the greatest SWE and a canopy structure that enhances evaporation. Depletion of surface storage areas in a dry year, combined with less winter snowfall and the delay of rainfall inputs until after the thawing front has passed the peat/silty clay interface, can dramatically decrease the subsequent spring runoff peak. The predominance
of open black spruce areas in the northern boreal forest emphasizes the need to understand better the functionality of these areas, including the identification of smaller surface storages and their incorporation into hydrologic models.

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