Using a Temperature-Based Model of Snow Accumulation and Melt to Assess the Long-Term Hydrologic Behaviour of Forested Headwater Basins in South-Central Ontario

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ABSTRACT

The Dorset Environmental Science Centre (DESC) has monitored the hydrochemistry of small forested basins on the Precambrian Shield of south-central Ontario since the mid-1970s. The resulting data set can be used to examine factors controlling the hydrochemical behaviour of headwater basins in this landscape, as well as their response to such anthropogenic disturbances as forest management and atmospheric deposition of contaminants. These analyses would benefit from accurate estimates of snow accumulation and melt, since snowmelt is the dominant hydrologic event in this region. The temperature-based WINTER snowmelt model (Scheider et al. 1983, Proc. Eastern Snow Conf. 28: 157-168) performed well in a limited test against one year of snow survey data from one of the DESC basins and melt estimates from an energy balance-based snowmelt model. The WINTER model was calibrated in the present study using five years of snow survey data from two sites in the DESC study area. Nash-Sutcliffe model efficiencies across all surveys were 53% and 79% for the southern and northern snow survey sites, respectively. The model was tested against in-situ data from nearby Meteorological Service of Canada climate stations, and provided realistic simulations of observed patterns of snow accumulation and melt. The value of predicted time series of snowpack development and ablation in interpreting the long-term hydrological record for the DESC basins is discussed.

Keywords: snow accumulation, snowmelt, rain-on-snow, temperature-index model, south-central Ontario

INTRODUCTION

Energy balance models provide more accurate estimates of snowmelt in a variety of environments (Price 1988, Kane et al. 1997, Garen and Marks 2005). Advances in energy balance modelling of snowmelt include greater consideration of such factors as snowpack internal energy and the role of grain metamorphism in temporal changes in snow albedo (e.g. Strasser et al. 2002, Garen and Marks 2005). However, snowmelt estimates are often required at sites where the data needed to run such models are not available (Rango and Martinec 1996). Air temperatures are generally the most ubiquitous meteorological data (Male and Gray 1981), thus encouraging use of temperature-based snowmelt models (but see Walter et al. 2005). Such models often perform poorly in open areas where air temperature may not be strongly correlated to incoming solar radiation, which can be the dominant source of energy to the snowpack (e.g. Bruland et al. 2001). However, they perform better in forests (USACE 1956), where there may be a stronger association between air and canopy temperatures (Male and Gray 1981). The latter in turn control longwave

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energy fluxes to the snow surface which comprise a greater portion of total energy inputs to the snowpack than in the open areas (Bengtsson 1976, Price 1988, Ward and Trimble 2003).

The Dorset Environmental Science Centre (DESC) has monitored the hydrology and hydrochemistry of a number of small forested basins on the Precambrian Shield of south-central Ontario since the mid-1970s. The streamflow and stream chemistry data can be used to examine the factors controlling the hydrochemical behaviour of this landscape, as well as basin response to natural and anthropogenic disturbance. Snowmelt is the dominant hydrologic event in this region, and such analyses would benefit from accurate estimates of snow accumulation and melt. Scheider et al. (1983a) compared the ability of an energy-balance model (MOEHYDR, Logan 1977) and a temperature-based model (WINTER) to simulate snow accumulation and melt in the DESC region. The WINTER model was driven by local air temperature and precipitation data, while the MOEHYDR model used local data on air temperature, wind velocity, dew point temperature, vapour pressure and potential evapotranspiration. Net radiation data were not available for the DESC region and were obtained from Ottawa ~260 km to the east. Scheider et al. (1983a) concluded that the WINTER model performed better than the MOEHYDR model at simulating snow survey results from the northern-most of the DESC basins (Harp Lake) in the 1981/82 Winter and Spring. However, snow survey data used to calibrate the WINTER model were also used to test it, and the extent to which this contributed to its superior performance is unclear. Scheider et al. (1983a) caution that this calibration means that the WINTER model may not be as widely applicable as the MOEHYDR model.

This paper evaluates the WINTER snowmelt model through calibration using a larger data set than was available to Scheider et al. (1983a), and testing the model against independent snow accumulation data from the DESC region. It also demonstrates the value of the predicted time series of snow accumulation and melt for interpreting the long-term hydrological record for the DESC basins.

STUDY AREA

The DESC basins are located in the District of Muskoka or Haliburton County, ON within 50 km of the DESC (45°13’N, 78°56’W, Figure 1). They occupy the southern portion of the Boreal ecozone, which has a humid continental climate with long cool summers (Köppen class Dfb). Average July and January temperatures are 18.7 and -11.1°C, respectively, and mean annual temperatures range from 3.6 (1993/94) to 7.6 (2001/02) °C . Annual precipitation at the DOR2 precipitation station for the period 1976 – 2001 ranged from 803 mm (1987) to 1278 mm (1980), with about 30% falling as snow (Eimers and Dillon 2002). Annual runoff for the DESC region is ~400 – 500 mm (Scheider et al. 1983b). Most of this runoff (49 – 77% on average) is delivered during Spring snowmelt, which generally occurs from mid-March to early May (McDonnell and Taylor 1987).

Bedrock in the DESC region is largely granitized biotite and hornblende gneiss, with some amphibolite and schist (Watmough and Dillon 2003). Surficial geology ranges from exposed bedrock, thin till (< 1m thick) interrupted by rock ridges, to plains with continuous till cover 1 - > 10 m thick (Devito et al. 1999). Surficial deposits in low lying areas are often mantled by peat (Watmough and Dillon 2003). Dominant upland soil types are acidic brunisols and podzols (Jeffries and Snyder 1983). Gleysols occur in moderately-to-poorly drained valleys in the southeastern part of the study area underlain by marble bedrock, while organic soils are common throughout the region in poorly drained zones (Jeffries and Snyder 1983). The DESC basins are in the Great Lakes-St Lawrence forest region. The area underwent selective logging in the 1800s and early 1900s, and the basins are now vegetated by secondary growth forests (Dillon et al. 1991). Detailed information on forest cover is given in Dillon et al. (1991) and Watmough and Dillon (2003). In general, well-drained soils have deciduous or mixed forests, while poorly drained soils have mixed or coniferous forest (Dillon et al. 1991).

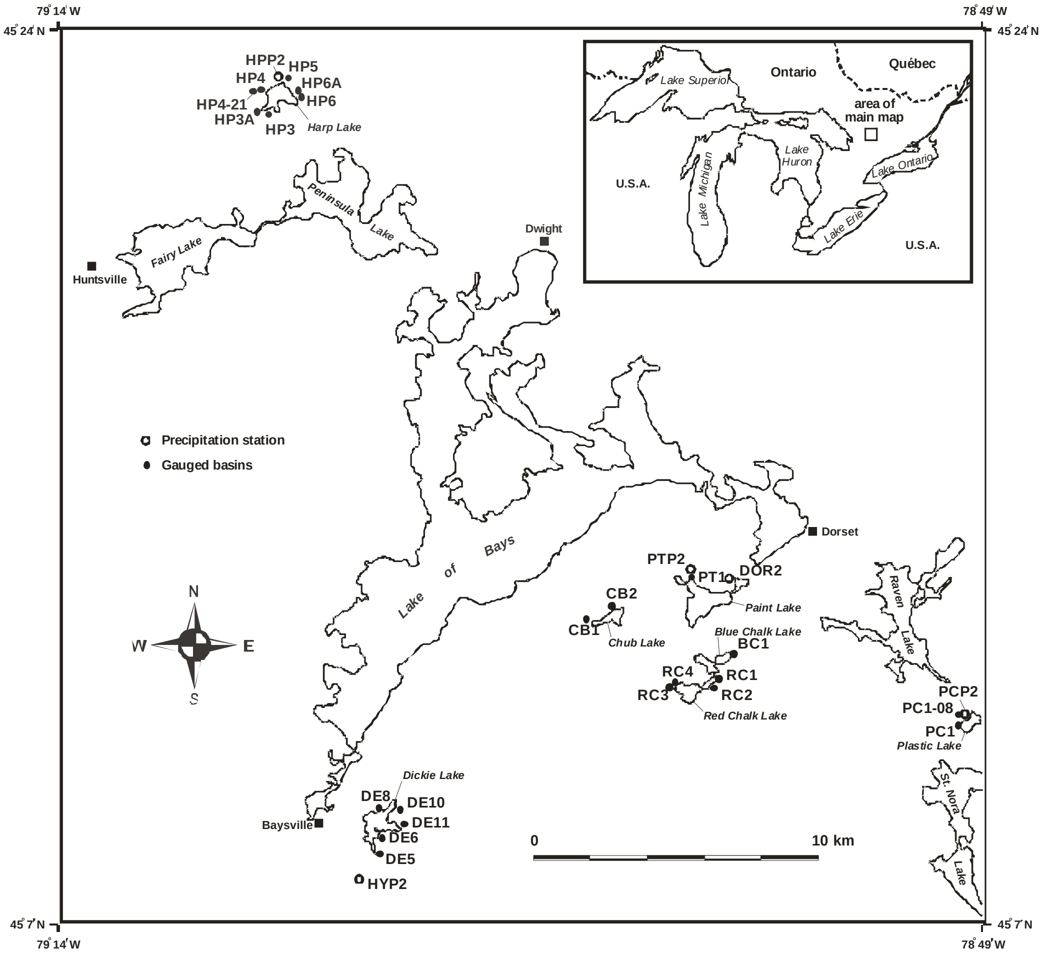

Figure 1: Location of the DESC study basins and precipitation stations.

METHODS

Snow course data

Snow water equivalent (*SWE*) and snow-covered area (*SCA*) data from the HPP2 and PCP2 precipitation stations were used (Figure 1). DESC staff made measurements using a Meteorological Service of Canada (MSC) snow tube at 6-point snow courses at each station for the Winters of 1987/88 to 1991/92 on a roughly weekly basis until the beginning of melt, when measurements were made more frequently (Findeis et al. 1993). The MSC snow tube overestimates true *SWE* by 6% on average (Goodison 1978). Average peak *SWE* at HPP2 ranged from 154 ± 18 mm in 1990/91 to 215 ± 20 mm in 1987/88, and from 118 ± 24 mm in 1989/90 to 181 ± 18 mm in 1987/88 at PCP2. Peak *SWE* values were consistently larger at HPP2 relative to PCP2, which reflects regional snowfall patterns (see below) and greater snow interception by the largely-coniferous forest cover at PCP2 compared to hardwood forest cover at HPP2.

Estimated point SWE

The WINTER model was used to simulate point-scale accumulation and melt (*Xmelt*) based on mean daily air temperature *T* and daily precipitation *P* (Figure 2). All precipitation below a threshold temperature (*B*) is assumed to be snow which accumulates in the snowpack, while *P* above this threshold is assumed to fall as rain. Melt for dry and rain conditions is estimated using separate relations (Scheider et al. 1983a). Air temperature and precipitation at HPP2 and PCP2 were used when available; otherwise, data from the MSC station at Dorset were used for PCP2, while data from MSC stations at Huntsville and Dwight were used for HPP2. The WINTER model was calibrated against measured *SWE* at HPP2 and PCP2 using the Nash and Sutcliffe (1970) model efficiency (*E!*) as the objective function. Model efficiencies of 1 indicate a perfect fit between observed and simulated values, while *E!*s less than 0 indicate that the model provides a poorer simulation of the observations than simply using the observed mean. Calibration was done using all 5 winter seasons of snow survey data, since they appeared to capture the range of snow accumulation and melt conditions found in the DESC landscape. Observed *SWE* was regressed on the optimized predicted *SWE*, and the regression slope was used to adjust the optimized predicted *SWE*s in order to fit a 1:1 relationship with observed *SWE*. Adjusted *SWE*s were used to estimate point-scale melt on day *i*:

$$Melt_i = SWE_{i-1} - SWE_i, \text{ when } SWE_{i-1} > SWE_i \quad [1]$$

$$Melt_i = 0, \text{ when } SWE_{i-1} \leq SWE_i \quad [2]$$

This approach assumes that sublimation losses from the snowpack are negligible and that reductions in *SWE* are due to loss of meltwater from the snowpack.

Testing of WINTER model

Predicted *SWE* at HPP2 was compared with snow-on-ground data from the Dwight MSC station (22 years of data) for 1976 – 2002. Snow-on-ground data from Huntsville were frequently incomplete and were not used. Predicted *SWE* at PCP2 was compared with snow-on-ground data from the Dorset (15 years of data) and West Guilford (7 years of data) MSC stations. West Guilford is 20 km southeast of PCP2. Snow surveys at HPP2 and PCP2 showed an increase in mean snowpack density with time for each Winter (Figure 3a). Density increases up to ~ 120 days after November 1 (March 1 of the following year) were largely due to compaction with increasing snow depth, while increases after this data were related to snow ripening and melt (Figure 3b). Best-fit relationships between density and time since November 1 were used to transform snow depths to *SWE*, with the HPP2 relationship applied to Dwight snow-on-ground data and the PCP2 relationship applied to Dorset and West Guilford snow-on-ground data. Nash-Sutcliffe *E!*s were determined for simulations of *SWE* at each station for each Winter with snow-on-ground data. Predicted and observed peak *SWE*, as well as the dates of the start of continuous snowcover (≥ 7 consecutive days of snow-on-ground), occurrence of peak *SWE* and the end of continuous snowcover (≥ 7 consecutive days of no snow-on-ground) were also determined.

Estimated snow-covered area

SCA for all snow survey years was plotted against normalized cumulative melt (*NCM*) estimated from the snow course data. A 5-parameter sigmoid was used to obtain best-fit relationships between *SCA* and *NCM* at each station (Figure 4) which were used to estimate the temporal change in *SCA* based on *NCM* from estimated point-scale daily melt (eqs. 1 and 2). The latter were then adjusted by the *SCA* to estimate basin-scale daily melt. Basin-scale melt was combined with estimated daily rainfall to obtain daily input to a given DESC basin.

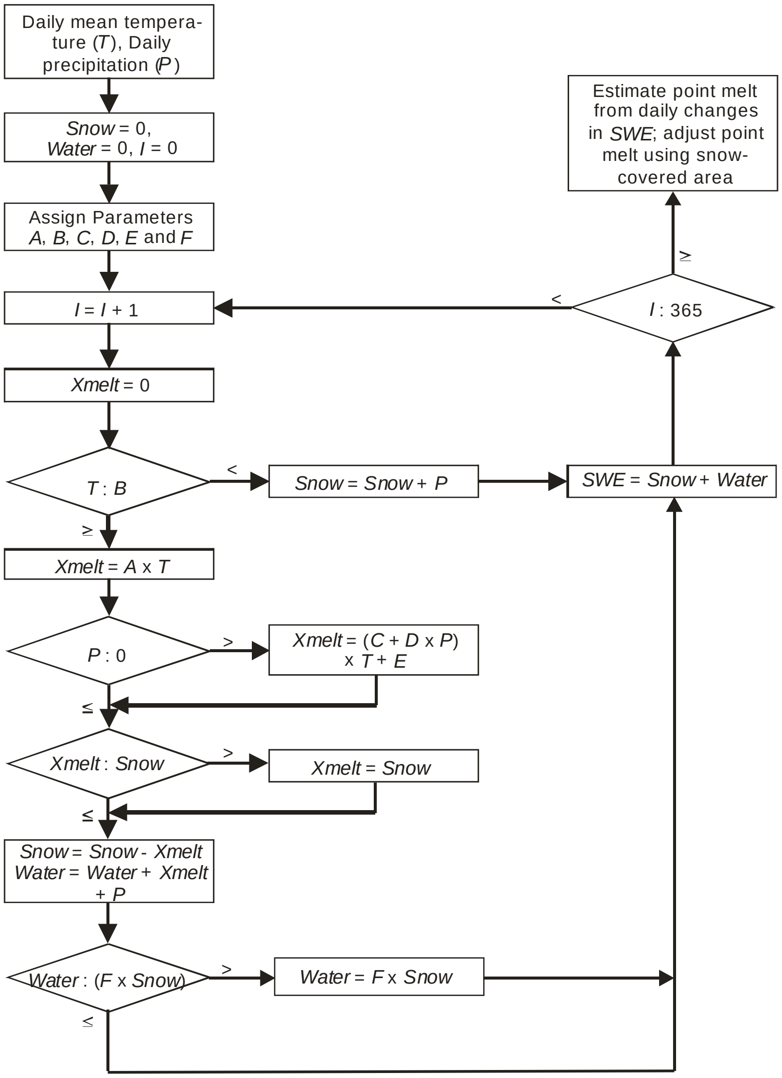

Figure 2: Flowchart of the WINTER model (adapted from Scheider et al. 1983a).

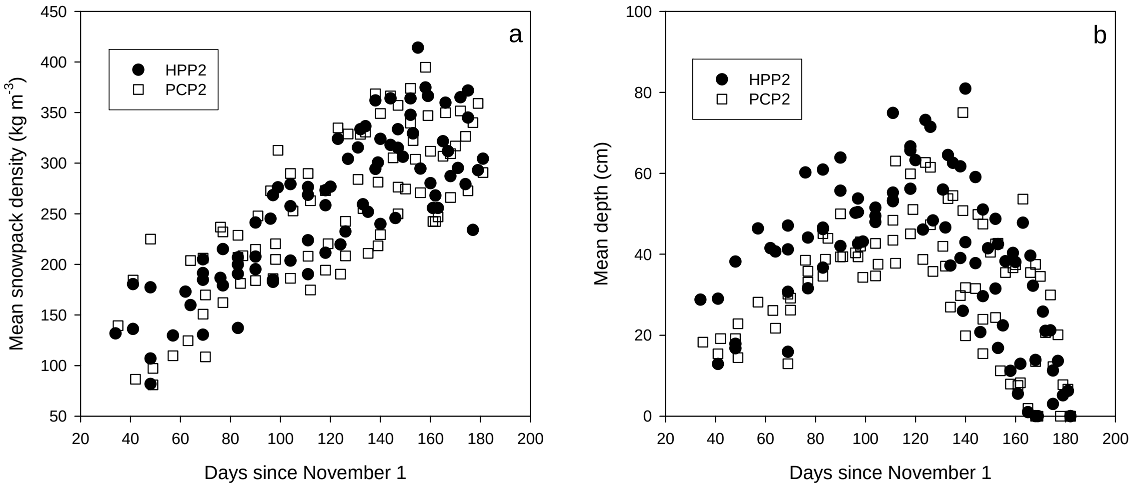

Figure 3: Mean snowpack density (a) and mean snowpack depth (b) measured at the HPP2 and PCP2 snow courses vs. days since November 1, Winters of 1987/88 to 1991/92.

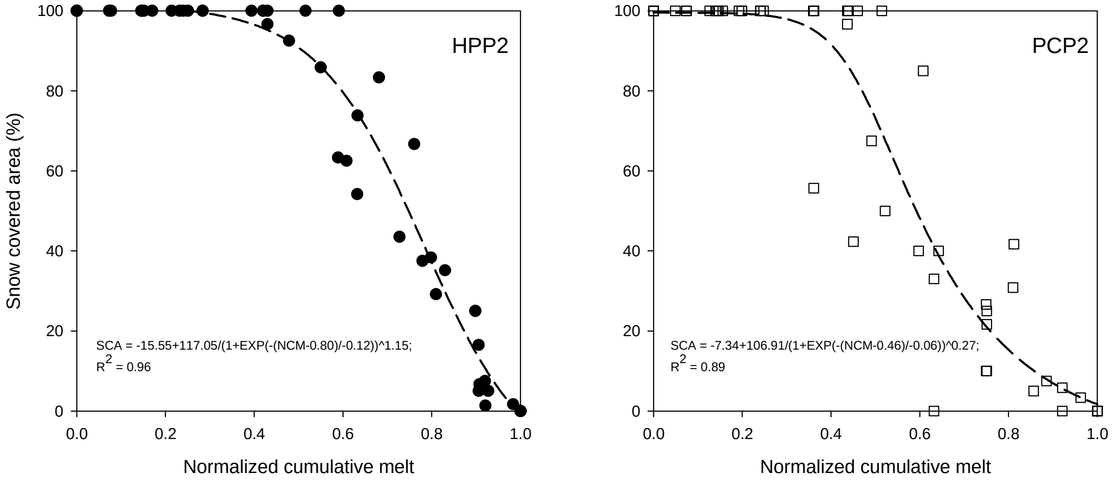

Figure 4: Snow-covered area (*SCA*) vs. normalized cumulative melt (*NCM*) measured at the HPP2 and PCP2 snow courses, Winters of 1987/88 to 1991/92. Five-parameter sigmoid best-fit relationships are indicated.

RESULTS AND DISCUSSION

Calibration of WINTER model

The calibrated WINTER model gave better fits to HPP2 snow survey data than for PCP2 (Table 1); nevertheless, the model had an *E!* > 0.5 at the latter site. Scheider et al. (1983a) obtained calibrated model parameters using snow survey data from Harp Lake in the 1981/82 Winter of A = 1.3, B = 1.5, C = 3.5, D = 0.012, E = 1.2 and F = 0.03. This study found smaller precipitation partitioning temperatures (parameter B), larger melt factors when $P > 0$ (parameters D and E) and much smaller optimum snowpack liquid water fractions (parameter F). Figure 5 presents best (a) and worst (b) fits between observed (mean ± 1 SD) and predicted *SWE* from the calibrated model. There was generally good agreement between observed and predicted *SWE* (Figure 5c, d);

however, the WINTER model tended to overestimate *SWE* at both HPP2 and PCP2. Slope coefficients of best-fit lines were used to adjust predicted *SWE* to remove this systematic overestimation, and adjusted predictions of *SWE* for snow survey days are shown in Figure 5a, b.

Table 1 Calibrated parameters for WINTER model and model efficiencies (E!) for all five years with snow survey data.

Parameter	HPP2	PCP2
A	1.9	1.2
B	0.8	1.2
C	3.5	3.3
D	0.03	0.019
E	1.6	2.3
F	0.00005	0.00001
E!	0.79	0.53

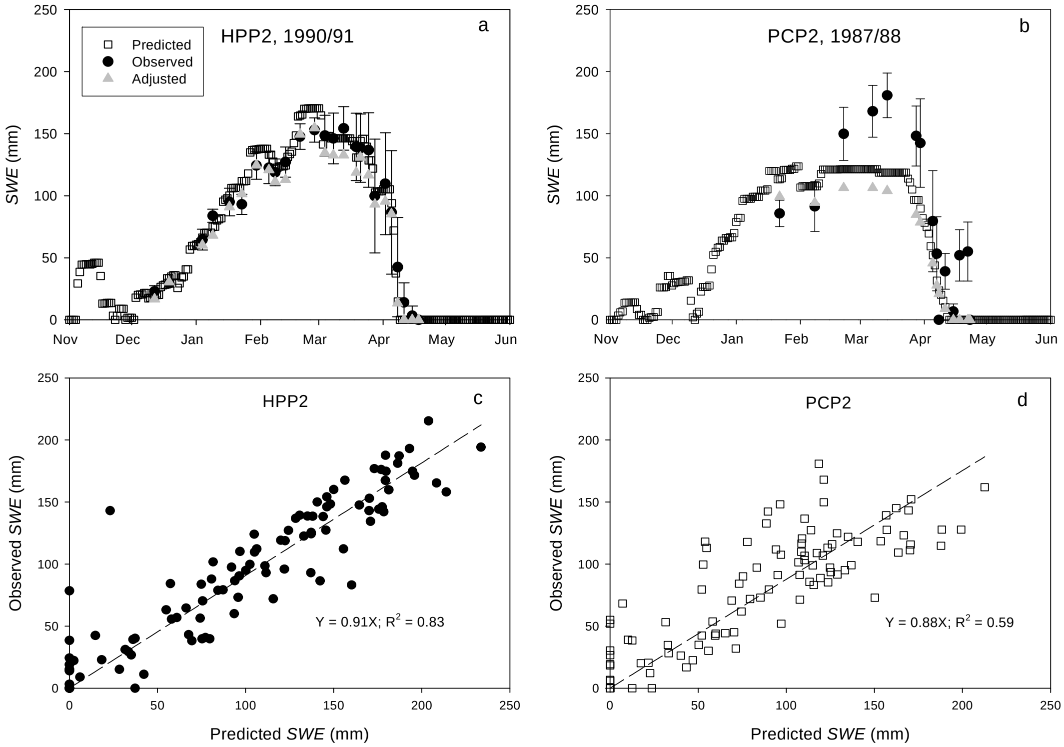

Figure 5: Examples of observed (± 1 SD), predicted and adjusted SWE values (a, b); observed vs. predicted SWE for the HPP2 (c) and PCP2 (d) sites, Winters of 1987/88 to 1991/92.

Testing of WINTER model

Calibrated parameters from HPP2 were used to predict *SWE* at Dwight, while PCP2 parameters were used to predict *SWE* at Dorset and West Guilford. Examples of good and poor simulations of *SWE* are given in Figure 6 (a, b), while Figure 6c summarizes *E!* values for the MSC stations. In all three cases, the WINTER model had *E!*s ≥ 0.5 for at least 50% of simulations. An *E!* threshold of ~0.5 has been used elsewhere in hydrology (e.g. Muleta and Nicklow 2005) to determine whether model simulations are behavioural (i.e. they provide an acceptable representation of the hydrologic system under study, Beven and Freer 2001). The poorest fits in general were for the

Dorset MSC station, while much better fits were found at Dwight and West Guilford. These *E!* values partly reflect the unknown error in the *SWE* values for each MSC station determined using observed snow-on-ground and assumed snowpack average density (Figure 3). Model predictions of the start of continuous snowcover were generally within ± 5 days of the observed date (Figure 7a), while most predictions of peak *SWE* date were within ± 10 days of the observed date (Figure 7b). The model generally predicted peak *SWE* to occur slightly later than was observed at the MSC stations, consistent with the tendency of *SWE* to peak earlier in open sites (e.g. MSC stations) than in forest areas (e.g. Buttle et al. 2005). Model predictions of peak *SWE* (Figure 7c) were within ± 50 mm of estimated *SWE* for most years at Dorset and West Guilford, and tended to underpredict peak *SWE* at Dwight and overpredict at Dorset. As noted above, these results partly reflect the unknown error in *SWE* values for the MSC stations. Nevertheless, predicted peak *SWE* values agree with snowfall patterns in the region. HPP2 lies between the Huntsville and Dwight MSC stations and the 1971 – 2000 climate normal snowfall for these stations is 285.6 mm and 332.1 mm, respectively. This suggests that HPP2 receives less snowfall than Dwight and accords with underprediction of peak *SWE* at Dwight. Similarly, PCP2 lies between the Dorset and West Guilford MSC stations and 1971 – 2000 climate normal snowfall for these stations is 277 mm and 384.9 mm, respectively. PCP2 likely receives more snowfall than Dorset, which agrees with the model's overprediction of peak *SWE* at Dorset. Most predictions of the date of loss of continuous snowcover (Figure 7d) were within ± 10 days of observations. The model predicted later dates of loss of snowcover at the Dorset MSC station, which agrees with lower peak *SWE* at Dorset and relatively faster melt at an open site compared to a forested location such as PCP2. Thus, the WINTER model provides realistic simulations of observed patterns of snow accumulation and loss in the DESC region.

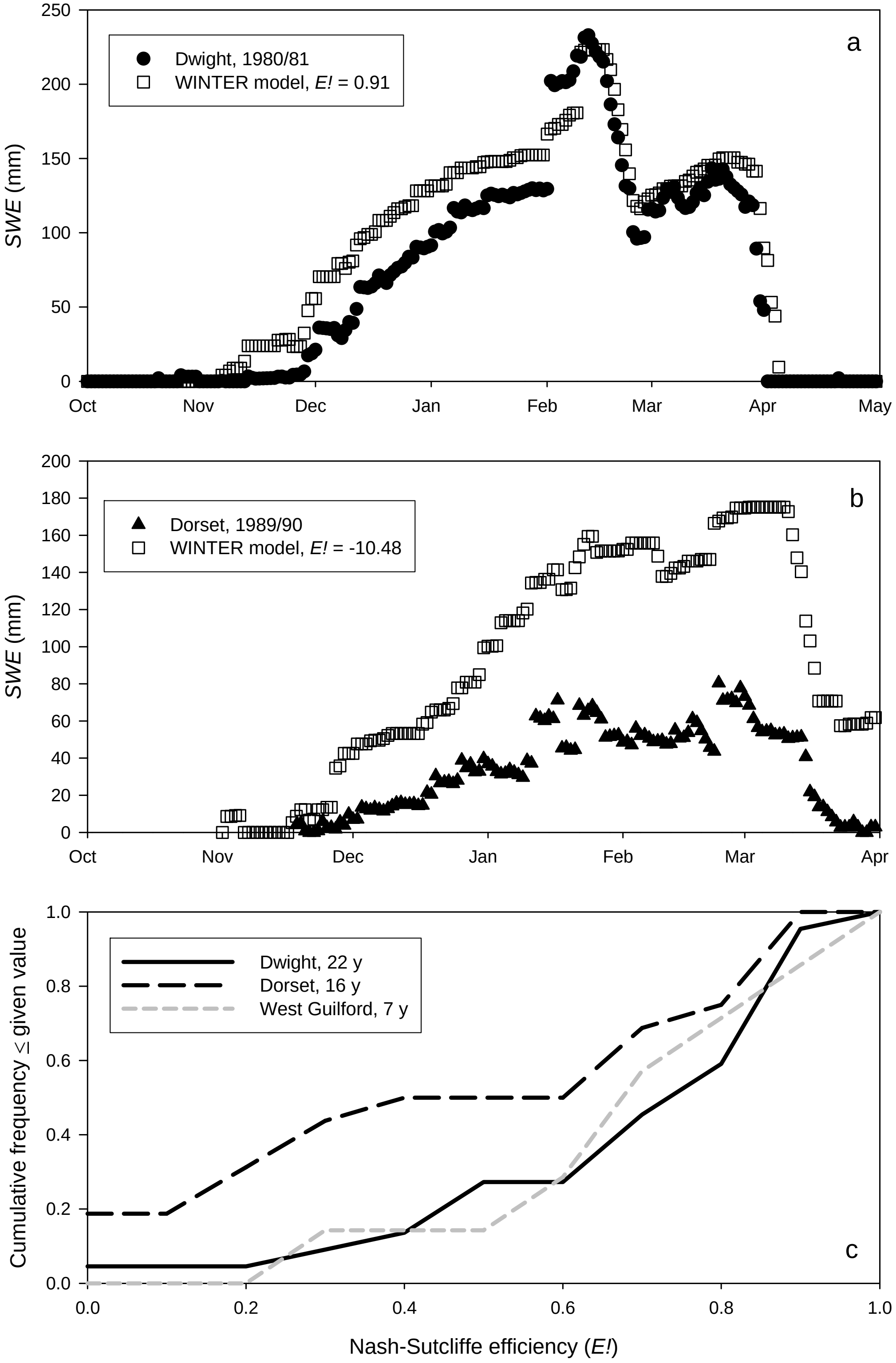

Figure 6: Examples of good (a) and poor (b) simulations of estimated SWE for MSC stations in the DESC region; cumulative frequency distributions of model efficiencies (E!) obtained in tests of WINTER model predictions of SWE estimated for MSC stations in the DESC region.

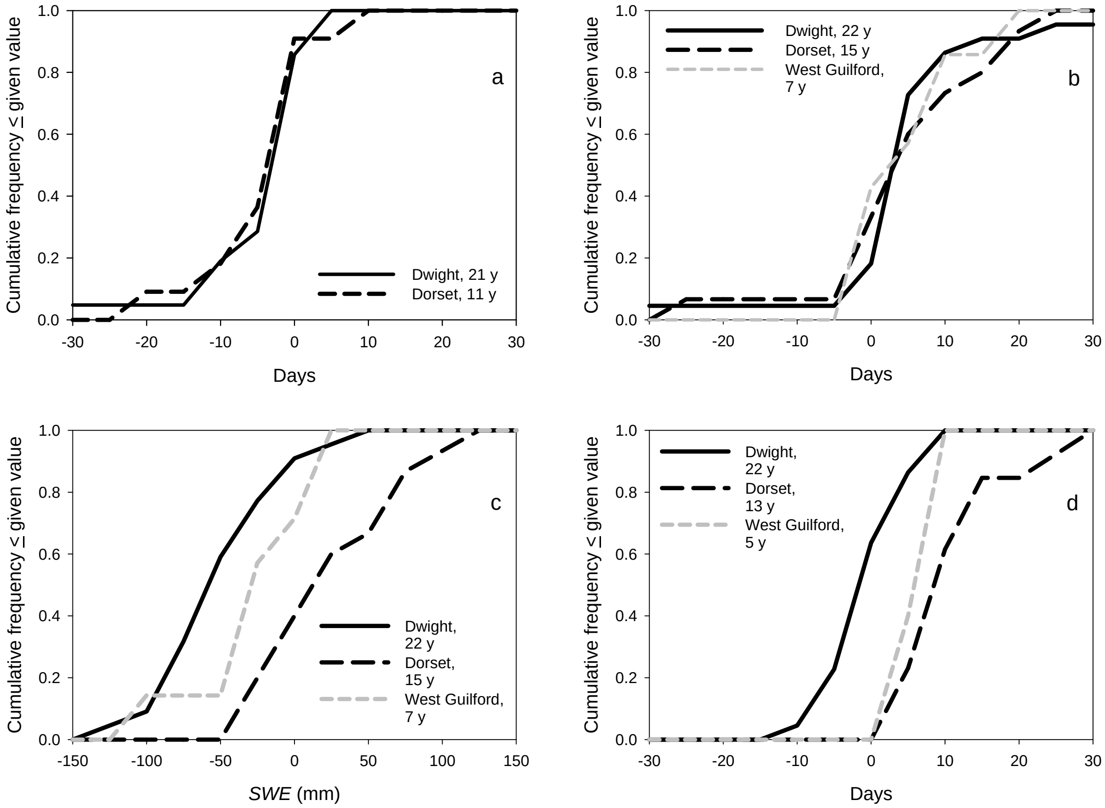

Figure 7: Cumulative frequency distributions of the difference between predicted and observed (a) date of the start of continuous snowcover, (b) date of peak *SWE*, (c) peak *SWE* and (d) date of loss of continuous snowcover for MSC stations in the DESC region.

Applications of *SWE* and melt predictions from the WINTER model

Daily estimates of *SWE*, rainfall and snowmelt from the WINTER model have been obtained using temperature and precipitation records from each of the DESC precipitation stations, and have been linked to streamflow time series from proximal DESC basins. This helps account for spatial variations in snow accumulation, snowmelt and rainfall across the DESC region when comparing streamflow behaviour between basins. The ability of snowcover to insulate soils and maintain above-freezing temperatures during Winter (Hardy et al. 2001, Monson 2006) has a strong control on mineralization and nitrification of organic N to NO_3-N (Schimel et al. 2004). Thus, the WINTER model's predictions of snowpack development and ablation enable interannual variations in the relative availability of NO_3-N for subsequent transport in Winter and Spring to be estimated.

Rainfall, snowmelt and rain-on-snow inputs to DESC basins can also be distinguished (Figure 8) and model predictions assist in explaining hydrograph peaks with no observed precipitation input, as shown by the snowmelt-induced peak event in early April 1992. The BC1 hydrograph was separated into quickflow and delayed flow components using the method of Hewlett and Hibbert (1967), and Figure 8 illustrates the importance of rain-on-snow inputs for quickflow production and peak streamflow events.

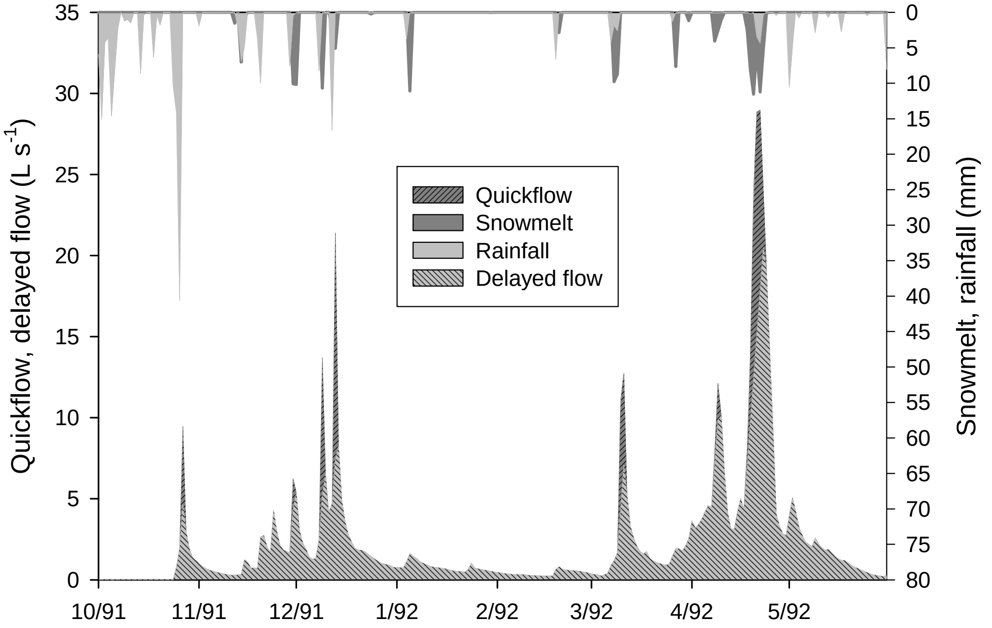

Figure 8: Quickflow and delayed flow components of streamflow, and estimated rainfall and snowmelt for the BC1 basin, 1991/92.

Rain-on-snow events promote Winter NO_3-N export from forested basins in the DESC region (Eimers et al. 2007), and climate warming in northern latitudes has been predicted to result in an increased fraction of winter precipitation falling as rain (IPCC 2007). By enabling streamflow events and their associated hydrochemical behaviour to be assigned to rainfall, snowmelt and rain-on-snow inputs for the entire period of record, the WINTER model will assist us in understanding how water and nutrient fluxes may differ with input type. This information will contribute to efforts to assess how the hydrochemical behaviour of streams in the DESC landscape will respond to a changing climate.

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