Gridded North American Monthly Snow Depth and Snow Water Equivalent for GCM Evaluation

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ABSTRACT *Evaluation of snow cover in GCMs has been hampered by a lack of reliable gridded estimates of snow water equivalent (SWE) at continental scales. In order to address this gap, a snow depth analysis scheme developed by Brasnett (1999) and employed operationally at the Canadian Meteorological Centre (CMC), was applied to generate a 0.3° latitude/longitude grid of monthly mean snow depth and corresponding estimated water equivalent for North America to evaluate GCM snow cover simulations for the Atmospheric Model Intercomparison Project II (AMIP II) for the period 1979–96. Approximately 8000 snow depth observations per day were obtained from U.S. cooperative stations and Canadian climate stations for input to the analysis. The first-guess field used a simple snow accumulation, aging and melt model driven by 6-hourly values of air temperature and precipitation from the European Centre for Medium-range Weather Forecasting (ECMWF) ERA-15 Reanalysis with extensions from the Tropical Ocean Global Atmosphere (TOGA) operational data archive. The gridded snow depth and estimated SWE results agree well with available independent in situ and satellite data over mid-latitudinal regions of the continent, and the snow depth climatology exhibited several improvements over Foster and Davy (1988). The monthly snow depth and estimated SWE climatologies are available for downloading from the Canadian Cryospheric Information Network (http://www.ccin.ca).*

RÉSUMÉ *[Traduit par la rédaction] L'absence d'estimations sur quadrillage fiables de l'équivalent en eau de la neige (ÉEN) à des échelles continentales a ralenti l'évaluation de la couverture de neige dans les MCG. Pour combler cette lacune, on a utilisé une méthode d'analyse de l'épaisseur de neige, élaborée par Brasnett (1999) et en usage fonctionnel au Centre météorologique canadien (CMC), afin de générer une grille à mailles de 0,3° en latitude/longitude de l'épaisseur de neige moyenne mensuelle et de l'équivalent en eau correspondant pour l'Amérique du Nord, en vue de l'évaluation des simulations de la couverture de neige des MCG dans le cadre du Projet de comparaison des modèles de l'atmosphère (AMIP II) pour la période de 1979 à 1996. Environ 8 000 observations par jour de l'épaisseur de neige ont été fournies par des stations collectives aux États-Unis et des stations climatologiques au Canada; elles ont servi de données d'entrée pour l'analyse. Le champ d'essai a utilisé un simple modèle de la hauteur, du vieillissement et de la fonte de la neige alimenté au moyen de valeurs de la température de l'air et des précipitations prélevées par pas de 6 heures et provenant de la réanalyse (ERA-15) du Centre européen pour les prévisions météorologiques à moyen terme (CEPMMT), avec des extensions tirées de l'archive des données opérationnelles du Programme d'étude des océans tropicaux et de l'atmosphère du globe (TOGA). Les résultats sur quadrillage de l'épaisseur de neige et les résultats estimés de l'ÉEN concordent bien avec les données in situ indépendantes et les données satellitaires obtenues dans les secteurs tempérés du continent. En outre, la climatologie de l'épaisseur de neige fait voir plusieurs améliorations par rapport à celle de Foster et Davy (1988). Les climatologies mensuelles de l'épaisseur de neige et les climatologies estimées de l'ÉEN peuvent être téléchargées à partir du site Web du Canadian Cryospheric Information Network (http://www.ccin.ca).*

1 Introduction

A realistic representation of seasonal and spatial variation in snow cover in climate models is important for snow-cover climate feedbacks (albedo-temperature, snow-cloud), soil moisture, runoff and ground temperatures (Cohen and Rind, 1991; Marshall et al., 1994; Lynch-Stieglitz, 1994). Previous evaluations of snow cover in General Circulation Models (GCMs) (Foster et al., 1996; Walland and Simmonds, 1996; Frei and Robinson, 1998) demonstrated that, to first order, GCMs cap-

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ture the seasonal cycle of Northern Hemisphere (NH) snow cover extent. However, they tended to underestimate fall and winter snow extent (especially over North America) and overestimated spring snow extent (especially over Eurasia) (Frei and Robinson, 1998). A recent assessment of the representation of snow in land surface schemes (Slater et al., 2001) revealed considerable differences in snow simulation results between models, particularly in the timing of spring melt.

Evaluation of GCM simulations of snow cover over the NH has been hampered by a lack of reliable data, particularly snow water equivalent (SWE). Existing snow depth climatologies (e.g., Schutz and Bregman, 1988; Foster and Davy, 1988) are based extensively on data collected prior to the Atmospheric Model Intercomparison Project II (AMIP II) period, and may not be representative of snow conditions during the 1980s and 1990s when NH snow cover extent decreased approximately 10% (Groisman et al., 1994). Reliable estimates of SWE have been obtained from passive microwave data over prairie land cover regions (Goodison and Walker, 1994; Derksen et al., 2002a). However, research is still ongoing to develop reliable satellite algorithms for extracting SWE over forested terrain, and the resolution of Scanning Multichannel Microwave Radiometer (SMMR) and Special Sensor Microwave Imager (SSMI/I) passive microwave data (25 km) is insufficient for mapping SWE in mountainous areas.

The aim of this project was to apply the snow depth analysis scheme developed by Brasnett (1999) and employed operationally at the Canadian Meteorological Centre (CMC), to develop monthly snow depth and SWE information for North America to evaluate GCM snow cover simulations during the AMIP II period. This is a contribution to AMIP II diagnostic subproject 28, “Snow Cover in General Circulation Models” (Robinson et al., 2000). The CMC operational snow depth analysis receives a limited number of snow depth observations from synoptic stations in real time. However, in historical analysis mode, an extensive database of daily snow depth observations from U.S. and Canadian cooperative stations can be used which provides ~8000 observations per day to the analysis. The analysis was only applied to North America as there were insufficient observations from Eurasia during the full AMIP II period.

2 Data sources

a *In situ Snow Depth*

1 CANADA

Regular daily ruler observations of the depth of snow on the ground have been made at most Canadian synoptic stations since the 1950s. The daily observing program was extended to climatological (cooperative) stations in the early 1980s, approximately quadrupling the number of stations in the network to over 2000. The observing network is concentrated over southern populated regions of Canada, and is biased to low elevations (Fig. 1). The number of stations reporting daily snow depth declined ~15% during the latter half of the 1990s in response to budget reductions and automation. Data rescue of previously undigitized Canadian snow depth data and reconstruction of missing values were carried out by Brown and Braaten (1998). The reconstructed values are required in the first two years of the AMIP II period (1979 and 1980) to maintain the spatial distribution of the observing network. Snow depth values from Canadian snow course reports were also incorporated into the snow depth analysis. These reports are less frequent (weekly, bi-weekly or monthly) but are an important source of information in mountainous areas of southern British Columbia and Alberta, and over northern Québec.

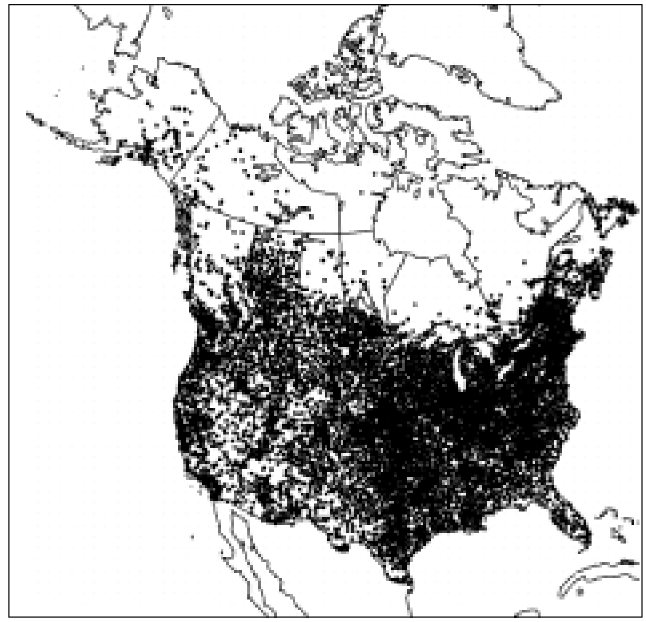

Fig. 1 The daily snow depth network available for the analysis. Station density decreases rapidly north of about 55ºN.

2 UNITED STATES

Daily snow depth observations were taken from the National Climatic Data Center (NCDC) TD-3200 Cooperative Summary of the Day database for the 1979–97 period. These include manual ruler measurements of snow depth from over 7500 cooperative stations across the contiguous U.S. and Alaska. Doesken and Judson (1997) provide a description of the U.S. snow depth observing program. The available network of stations (Fig. 1) gives excellent spatial coverage over the contiguous U.S. The data exhibit a low elevation bias in mountainous regions such as the western cordillera, and data coverage is sparse in Alaska. The NCDC perform basic quality control of the data (outliers, internal consistency, areal consistency) and values failing their checks were excluded from the analysis.

A total of over three million observations were available from the U.S. and Canada for each year of the analysis, with ~85% of these being zero reports. Zero reports are important, however, for the accurate location of the snow/no-snow boundary. Overall, a total of ~400,000 non-zero daily snow depth observations were available each year for inclusion in the analysis. Further quality control of the observations was carried out as part of the analysis process.

b *ECMWF Air Temperature and Precipitation*

First-guess snow depth fields for the snow depth analysis were generated using a simple snowpack model driven with 6-hourly 2-m air temperature and total precipitation values from the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-15 Reanalysis described by Gibson et al. (1997). Details of the snowpack model are provided later. Additional data for the 1994–97 period not covered by ERA-15 were obtained from the ECMWF/World Climate Research Program Tropical Ocean and Global Atmospheric Program (ECMWF/WCRP TOGA) operational data archive. The data were provided on a 1.125° latitude/longitude grid and were interpolated to a 1/3° grid for running the analysis. The ERA-15 data have been subjected to extensive evaluation (e.g., Hanna and Valdes, 2001; Cullather and Bromwich, 2000; Serreze and Hurst, 2000) and the general consensus is that they provide a realistic representation of temperature and winter precipitation over the study area. A number of the islands in the Sverdrup Basin (Canadian high Arctic) were not resolved on the ECMWF 1.125° grid which meant that 2-m air temperatures remained below freezing throughout June, July and August at these points, causing snow depth to increase monotonically over the period of the historical analysis. These points need to be rejected from any data summaries in this area and have been set as ocean points in the land/sea mask accompanying the dataset.

3 Analysis methodology

The daily snow depth objective analysis is based on statistical interpolation (Daley, 1991), and follows the approach described in Brasnett (1999) modified to accept snow depth observations once every 24 hours instead of every 6 hours. The basic steps in this process are: (1) derivation of a "first-guess" or background field based on a simple snowpack model driven with 6-hourly values of ECMWF precipitation and 2-m air temperature; (2) quality control of observations; and (3) statistical interpolation of the observations to grid points every 24-hours. The important details of each of these steps are outlined below:

a *Background Field (Snowpack Model)*

The background field was generated with a simple snowpack model using 6-hourly ECMWF forecast precipitation and analysed 2-m air temperatures as input. A previous run of the historical snow depth analysis (Brown et al., 2001) showed that the simple snow aging scheme used in Brasnett (1999) did not provide an adequate simulation of the spatial variation in mean snow density over North America. It was therefore decided to include additional terms in the model that had an impact on snow density, snow accumulation and melt. These included mixed precipitation type, rain melt, a more detailed treatment of snow aging, a variable melt factor, and estimates of blowing snow and canopy sublimation loss.

It would have been preferable to apply a physically based snowpack model such as SNTHERM (Jordan, 1991) to simulate the background snow depth field. However, the required additional input data (incoming short and longwave radiation, relative humidity, wind speed) were not available, and a multi-layer physical model would have exceeded available computing resources. In the configuration described here, the analysis took approximately one day to run a full year snow depth analysis over North America on a Silicon Graphics Inc. (SGI) Origin 2000 server. Another reason for keeping the snow modelling approach simple was that any improvements could readily be incorporated into the CMC operational snow depth analysis, which is subject to strict computer resource and time constraints.

The analysis was run on a global 1/3° latitude-longitude grid. The ECMWF fields were interpolated to this grid, and a saturated lapse rate of 0.006 K m^{-1} applied to adjust air temperature to the mean elevation of each grid square. The snowpack model was run at an hourly time step using linearly interpolated air temperature and hourly precipitation rate computed as one-sixth the 6-hourly forecast total precipitation. The fraction of precipitation occurring as snow was assumed to be 1.0 for air temperatures of 0°C or less, and to decrease linearly to zero at an air temperature of +2°C. This particular relationship was found to give improved simulation of snow cover at several locations in Canada compared to a fixed 0°C threshold. New snowfall density (ρ_{sfall}) was assumed to be a function of air temperature (T_{air}) following Hedstrom and Pomeroy (1998):

$$\rho_{sfall} = 67.9 + 51.3 \exp(T_{air}/2.6) \qquad T_{air} \le 0°\mathrm{C}, \tag{1}$$

$$\rho_{sfall} = 119.2 + 20.0\, T_{air} \qquad T_{air} > 0°\mathrm{C}. \tag{2}$$

Snowmelt was computed using a variable degree-day melt factor (γ) determined as a function of snowpack density (ρ_s) and vegetation cover (open/forested) following Kuusisto (1984):

$$\gamma = 0.0104\, \rho_s - 0.70 \quad (1.4 < \gamma < 3.5) \text{ forested areas (mm d}^{-1}\text{)} \tag{3}$$

$$\gamma = 0.0196\, \rho_s - 2.39 \quad (1.5 < \gamma < 5.5) \text{ open areas (mm d}^{-1}\text{)}. \tag{4}$$

This relationship takes into account the higher melt rates in the second half of the snow year when average snow albedo is lower. The open area expression (Eq. (4) was used in all regions except the "*taiga*" snow-climate classification zone of Sturm et al. (1995) where Eq. (3) was used. The hourly melt was computed using a standard temperature index approach

$$(T_{air} - T_{melt})\, \gamma/24, \tag{5}$$

where T_{melt} is the threshold air temperature for snowmelt. Evaluation of Eq. (5) at several sites across Canada yielded an optimum value for T_{melt} of –1.0°C. There is considerable variability in published values of T_{melt} and a value less than 0°C is not physically unrealistic since radiation melt can take place when air temperatures are below freezing. Kuusisto (1984) obtained T_{melt} values of –1.3°C and –1.2°C for open and forested sites in Finland.

Melt from rain falling on the snowpack was computed from the standard expression

$$(C_w R_{fall} T_w) / (L_f \rho_{ice}) \text{ (mm s}^{-1}\text{)}, \qquad (6)$$

where C_w is the heat capacity of water, R_{fall} is the rainfall rate in mm s^{-1}, T_w is the rainfall temperature (assumed equal to air temperature), L_f is the latent heat of fusion, and ρ_{ice} is the density of ice.

The increase in snow density at each time step was computed using empirically-based snow aging expressions for cold and melting snow. The snow melt threshold air temperature of –1.0°C was used to discriminate between the two regimes. The aging expressions were calibrated and evaluated using observed snow depth and density data for a range of observing sites in the snow-climate classes identified by Sturm et al. (1995). The Sturm et al. (1995) classification separated seasonal snow cover into six classes (*tundra, taiga, alpine, maritime, prairie* and *ephemeral*) based on unique ensembles of textural, stratigraphic and climate characteristics. It was reasoned that this provided a convenient, physically based approach for treating the snow aging process over the North American continent. The snow classification was obtained on a global 0.5° latitude/longitude grid from the National Snow and Ice Data Center (NSIDC), and interpolated (nearest neighbour method) to a 1/3° latitude-longitude grid for input to the snow analysis.

The hourly settling rate for cold snow ($\Delta\rho_s$) was estimated from Anderson (1976) by

$$\Delta\rho_s = 1000\, C1 \exp(-0.08\,(T_{melt} - T_{snow}))\ 0.6\ \text{SWE} \exp(-C2\,\rho_s) \text{ kg m}^{-3}\text{ h}^{-1}, \qquad (7)$$

where $C1$ is the snow settling rate at T_{melt}, T_{snow} is snow temperature (°C), SWE is snow water equivalent (cm), $C2$ is an empirically derived constant, and ρ_s is the snow density (g cm^{-3}) from the previous time step. SWE was multiplied by a factor of 0.6 to approximate applying Eq. (7) over a series of discrete snow layers as intended by Anderson. Kojima (1967) reported a range of values for $C1$ (0.026 to 0.069) and a value for $C2$ of 21. Anderson (1976) obtained the best results for Sleepers River Research Watershed, Vermont with $C2$=21 and $C1$=0.1. In this study, air temperature was substituted for snow temperature in Eq. (7). Snow temperatures are normally warmer than air temperatures for much of the snow cover season, so this approximation would underestimate aging rates if Anderson's suggested values for $C2$ and $C1$ were used. Comparison of the snowpack model with snow depth observations at several sites across Canada revealed that using values of $C1$ = 0.02 and $C2$ = 21 in Eq. (7) gave reasonable estimates of snow depth evolution. The major exception was in deep, cold snowpacks (e.g., *taiga* region) where the computed aging rate was too high. In these areas, greatly improved results were obtained with $C2$ = 28. The evaluation process also revealed that the initial densification rate of snow on the ground was too slow using Eq. (7). In order to improve this, a minimum mean snowpack density was used to initialize the snowpack. These values were derived from an analysis of observed density information from the Canadian Snow Data CD-ROM (MSC, 2000) for each of the Sturm et al. (1995) snow classes (Table 1).

TABLE 1. Minimum mean snow pack density values obtained from analysis of Canadian snow course observations classified by Sturm et al. (1995) snow-climate classes. A mean minimum value of 200 kg m^{-3} was assumed for snow on land ice (the "ice" class).

Sturm et al. (1995) Classification	Minimum mean snowpack density (ρ_{min}) kg m^{-3}
Tundra	200
Taiga	160
Maritime	160
Ephemeral	180
Prairie	140
Alpine	120
Ice	200 (assumed)

The calculation of the density increase for "warm" snow ($T_{air} > T_{melt}$) at each hourly time step was computed following Tabler et al. (1990) by

$$\rho = \rho^* - 200 \exp(\log((\rho^* - \rho_{-1})/200) - 2.778e^{-6}\ 3600), \qquad (8)$$

where ρ_{-1} is the density from the previous hour, and ρ^* is the maximum possible value of snow density which is a function of snow depth h_s (cm) following

$$\rho^* = 700 - (20470 / h_s)(1 - \exp(-h_s / 67.3)). \qquad (9)$$

This particular formulation was found to give good agreement with observed spring season snow aging at Goose Bay, Canada and Col de Porte, France. Both these sites were used in the recent international snow model intercomparison project (SnowMIP) (Essery et al., 1999).

The loss of snow mass from blowing snow sublimation is important in exposed landcover regions of North America such as prairie and tundra (Pomeroy and Gray, 1995; Pomeroy et al., 1997). Pomeroy and Gray (1995) estimated sublimation loss over prairie environments to be 15 – 41% of annual snowfall. Sublimation of snow from conifer canopy was determined to be of a similar magnitude: Pomeroy and Gray (1995) estimated that approximately one-third of the total snow falling on spruce and pine was lost through canopy sublimation. Pomeroy and Gray (1995) provided simplified climatological equations for computing sublimation loss from

TABLE 2. SWE and snow depth evaluation results for the simple snowpack model at three locations in different snow-climate zones. Statistics were computed using annual mean values.

	Estevan (1965–1993) *Prairie*	Goose Bay (1969–1983) *Taiga (forested)*	Resolute Bay (1966–1993) *Tundra*
Observed Mean SWE	34.5	155.0	48.8
Simulated Mean SWE	39.6	147.8	58.9
Correlation	0.74	0.91	0.65
Slope	0.99	1.11	1.15
	Estevan (1963–1995) *Prairie*	Goose Bay (1969–1983) *Taiga (open)*	Resolute Bay (1966–1993) *Tundra*
Observed Mean Snow Depth	10.5	57.8	20.3
Simulated Mean Snow Depth	11.6	50.5	19.7
Correlation	0.84	0.94	0.85
Slope	0.95	0.94	0.94

blowing snow as a function of wind speed, relative humidity and temperature. Unfortunately, ECMWF wind speed and humidity data were not available for this study. In their absence, a mean sublimation loss of 20% of precipitation was applied to the Sturm snow-climate classes with the greatest potential for blowing snow sublimation (*tundra, prairie)* and a similar 20% loss applied to the *taiga* class to approximate canopy sublimation.

Evaluation of simulated SWE with snow course observations at three different snow-climate zones (Table 2) revealed reasonable performance, although there was noticeable interannual variability in results at the *prairie* and *tundra* sites. The simulated snow depths showed somewhat better agreement with snow depth observations, and the model was able to simulate closely the seasonal evolution of snow depth at all three sites (Fig. 2). For these comparisons, the model was run with 6-hourly values of air temperature at 0000, 0600, 1200 and 1800 UTC, and 6-hourly totals of precipitation corrected for wetting loss and under-catch following Goodison et al. (1998).

b *Quality Control*

The quality control checks fall into two main categories: those relying solely on grid point data and those using nearby observations. Two tests of the former type are applied. In the first, an observation of non-zero snow depth is rejected if there was no snow in the previous analysis (valid 24 hours earlier) and the ECMWF temperature fields show temperatures consistently above 3°C during the intervening period. In the second test, an estimate of the snow depth using only ECMWF fields (and no observations) is updated daily and used for quality control. If an observation is less than this estimate and the difference exceeds 40 cm, the observation is rejected. Finally, the most powerful quality control check is one that brings to bear information from nearby observations. Using the statistical interpolation methodology described in the next section, a snow depth is computed at a station location using the background and neighbouring observations and compared with the reported depth at the station. The rejection criterion for this test depends on the number of neighbouring stations and their proximity to the report being tested, but allowed discrepancies never exceed 20 cm. Typically, this test rejected about 1% of the observations.

c *Statistical Interpolation*

The method of statistical interpolation, also called optimum interpolation (OI), is used to blend information from the observations and the background (or first guess). As described in Brasnett (1999), the corrections to the first guess due to the observations are computed at each grid point, g, using the weighted sum

$$\hat{f}_g = \Sigma w_i f_i \tag{10}$$

where f_i are the differences between the observations and the first guess at station locations (the result of a horizontal bilinear interpolation of the first guess) and w_i are the optimum weights. The sum in Eq. (10) is over all stations within 600 km of g. The computation of the weights relies on knowledge of the error characteristics of the first guess and the observations. Errors in the first guess may come from errors in the previous analysis, errors in the ECMWF precipitation and temperature inputs, and errors in the assumptions described above to account for the evolution of the snowpack. These errors can be expected to be spatially correlated. Accordingly, a model for the first guess error has been chosen such that the spatial correlation depends only on horizontal and vertical separation with an e-folding distance of 120 km in the horizontal and 800 m in the vertical. Observation error includes observer error, instrument error, and errors of representativeness (from point measurements of snow depth that may contain information on small scales not resolved by the analysis scheme). These errors are assumed to be uncorrelated from one station to the next. The observations are ascribed a standard deviation of 2.4 cm, while the first guess standard deviation is taken to be 3.2 cm.

4 Results and evaluation

Monthly means and climatological means of snow depth and estimated SWE during the 1979/80 to 1996/97 snow seasons were computed from the daily output from the historical analysis. The monthly timescale was selected to correspond with previously published snow climatologies and GCM output. This particular statistic and time average may not be optimal for some applications such as evaluation of satellite algorithms, but higher temporal resolution information can be retrieved from the archived daily grids if desired. The main features and improvements of the new dataset are revealed through the evaluation process, which involved comparisons with existing climatologies, satellite data, and in situ observations. In the following discussion, the historical analysis is referred to as “CMC”.

a *Snow Depth and Snow Cover Extent*

Comparison of the CMC monthly snow depth climatology with the Foster and Davy (1988) snow depth climatology

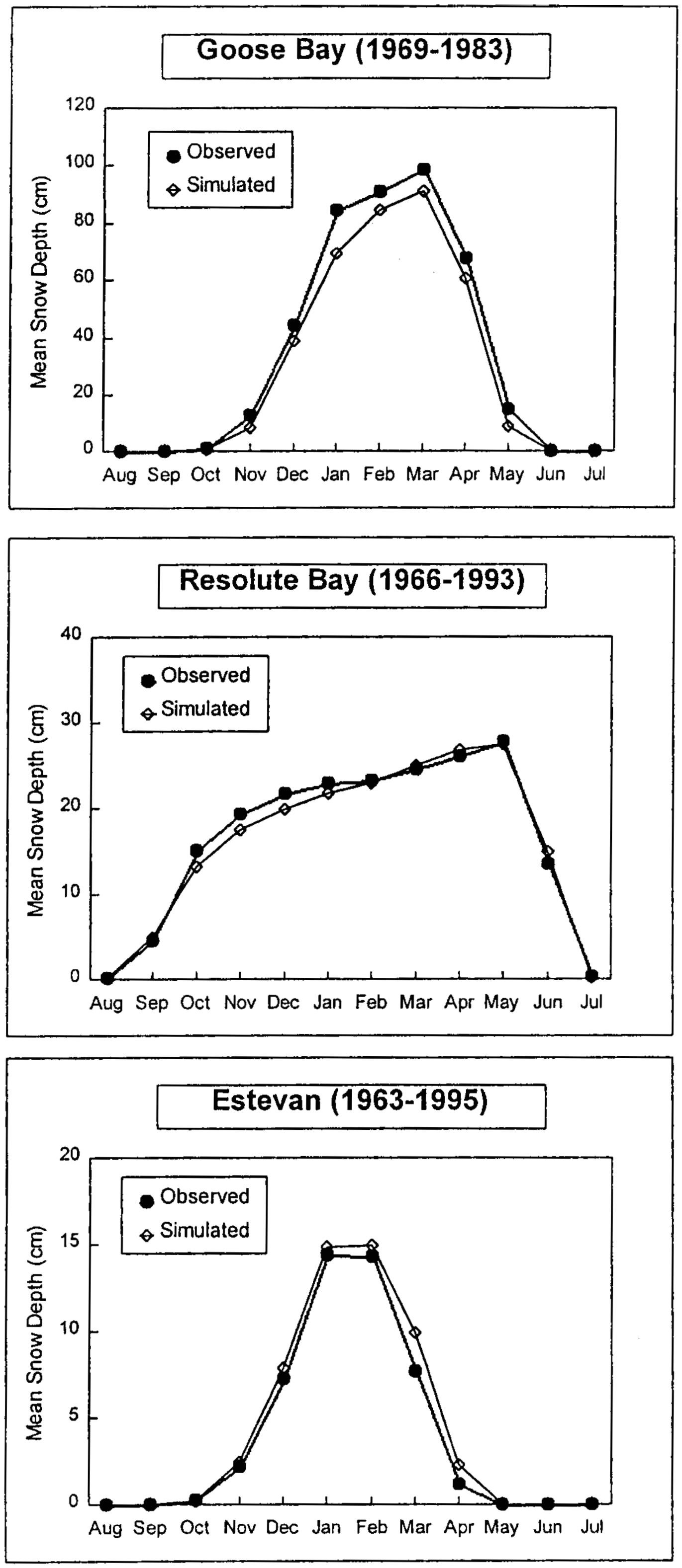

Fig. 2 Comparison of observed and simulated mean monthly snow depths for three sites in different snow-climate zones: Goose Bay (*taiga* – open site), Resolute Bay (*tundra*), Estevan (*prairie*).

(henceforth termed "USAF") revealed similar overall patterns in maximum winter accumulation (Fig. 3b) with maxima located over the western cordillera and Québec-Labrador. In the CMC climatology, however, the maxima were more spatially constrained, particularly over the western cordillera. This difference is related to the higher resolution of the CMC

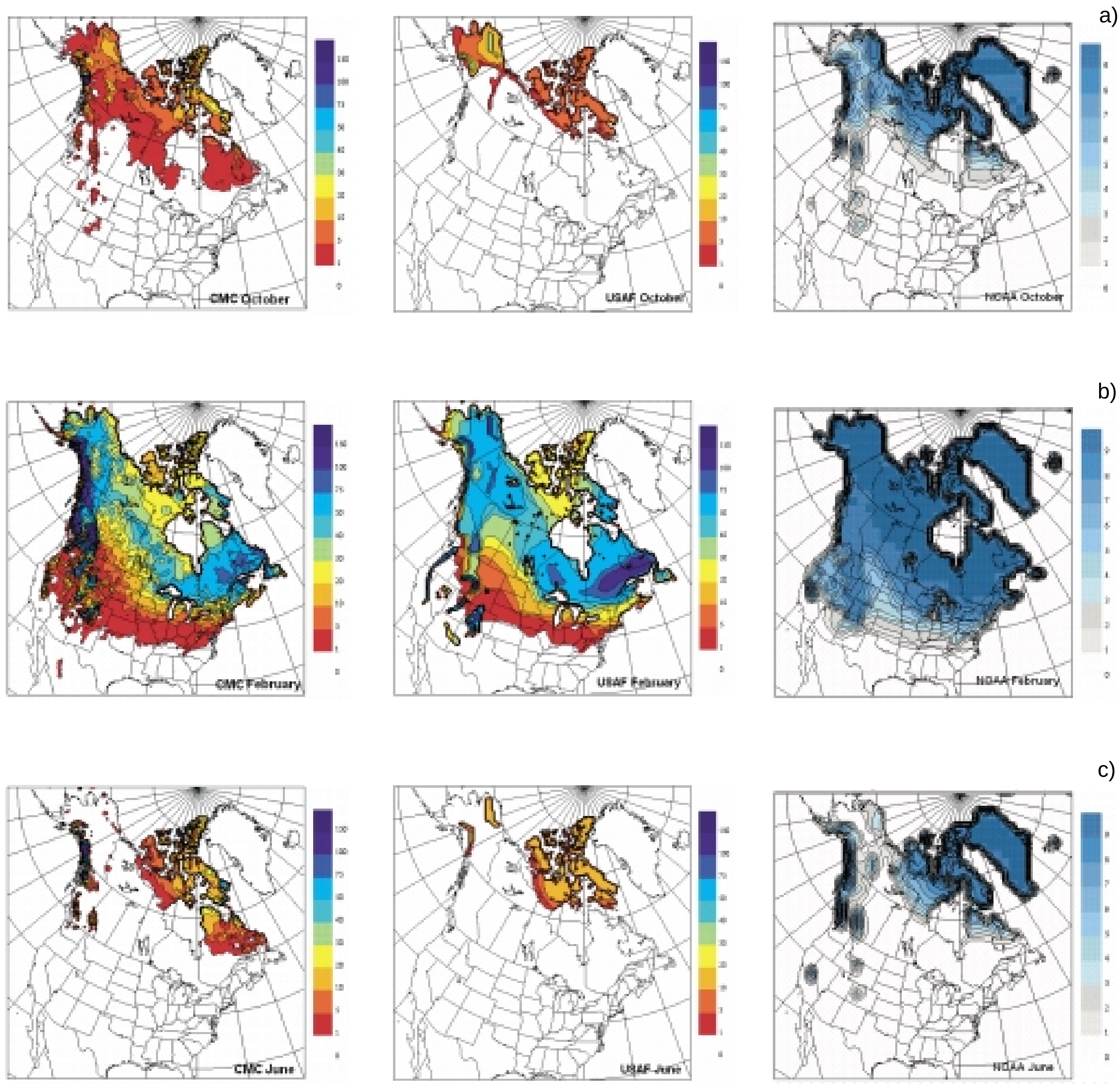

Fig. 3 Comparison of mean snow depths (cm) for the CMC (1979–97) and Foster and Davy (1988) snow depth climatologies for (a) October, (b) February and (c) June. Mean snow cover in percent for the 1979–97 period from the NOAA weekly dataset is shown on the right.

snow depth analysis (~37 km versus ~48 km for the USAF climatology), and to the more detailed treatment of snow in data sparse regions: the USAF climatology used manually interpolated values of mean monthly precipitation from World Meteorological Organization (WMO) Climatic Atlases, with a constant assumed snow density of 300 kg m^{-3} to determine the snow depth in data sparse mountainous regions.

Comparison of the two snow depth climatologies with mean snow cover extent computed from the National Oceanic and Atmospheric Administration (NOAA) weekly satellite-derived dataset (Fig. 3) revealed that the CMC climatology provided a more realistic representation of snow cover extent in October (Fig. 3a) and June (Fig. 3c), as well as improved representation of winter snow cover over the south-western United States (Fig. 3b). The CMC analysis underestimated June snow cover over northern Alaska and the Yukon (Fig. 3c) due to overly fast melt of snow in the spring. This problem is partly related to an open location bias in the observed snow depth data included in the analysis, and to the fact that the snow model assumed that all areas were open with the exception of the boreal forest zone.

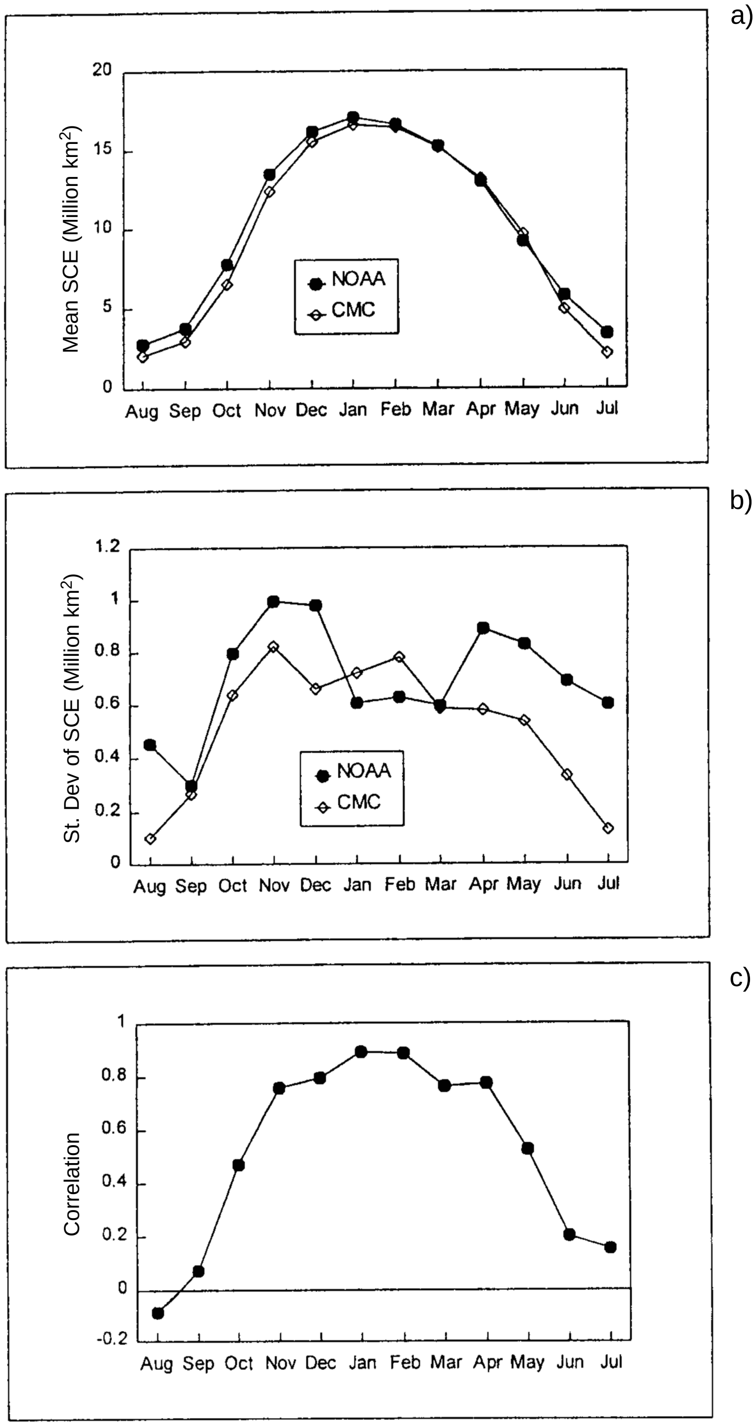

Fig. 4 Comparison of mean (a) and standard deviation (b) in North American monthly SCE for the period 1979/80 to 1996/97. The detrended correlation between the monthly SCE series is shown in (c).

Comparison of mean snow cover extent (SCE) for all months (Fig. 4a) revealed that the CMC snow depth climatology provided a close approximation of observed seasonal variation in mean SCE. SCE was estimated from snow depth using the following empirically derived expression based on comparison of Canadian in situ snow depth observations with corresponding SCE from the NOAA dataset:

$$\text{SCE}\ (\%) = 100 - (15 - h_s)^{1.7} \quad h_s < 15\text{ cm}, \qquad (11)$$

where h_s is snow depth in cm. An investigation of the ability of the snow depth analysis to replicate the observed interannual variability in continental snow cover extent during the 18-year AMIP II period revealed excellent agreement during the November to April period (mean correlation = 0.82) when the snow line was located over observation-rich mid-latitudes (Fig. 4c). However, the two datasets exhibited no significant correlations during the May to October period. There are several factors contributing to this. First, in May the snowline has moved into data sparse regions of northern Canada, and the analysis relies, more-or-less exclusively, on simulated snow depth using ECMWF air temperature and precipitation inputs. Second, comparison of the CMC analysis with the NOAA product for individual years showed that the CMC analysis had melted off nearly all the continental snow cover by the end of June, and there was virtually no snow left to generate any interannual variability during the rest of the summer months. This is clearly seen in the rapid drop-off in the standard deviation of SCE in Fig. 4b. However, it should also be noted that NOAA satellite data are known to be less accurate in areas with persistent cloud cover and/or heavy forest (Scialdone and Robock, 1987; Wiesnet et al., 1987, Robinson et al., 1993). Thus, greater uncertainties in the NOAA product are also likely to contribute to the May–October results. Although, the NOAA dataset has been used extensively to document recent trends in continental SCE (e.g., the recent third Intergovernmental Panel on Climate Change (IPCC) assessment), it has not been subject to detailed verification over high latitude regions of the NH.

b *Snow Water Equivalent*

SWE was estimated from the analysed snow depth and the corresponding snow density simulated by the snowpack model. In areas with a high density of snow depth observations, the resulting SWE estimates are largely controlled by the observed spatial and temporal variability in snow depth. In data sparse areas, however, the SWE estimates will be entirely simulated from the snowpack model driven with the ECMWF temperature and precipitation fields.

Evaluation of the estimated SWE results was carried out using a compilation of monthly and bi-monthly snow course SWE observations for Canada during the 1960–95 period (MSC, 2000), and snowpack telemetry (SNOTEL) snow pillow measurements of SWE over the western U.S. during the period 1980–98 (Serreze et al., 1999). The Canadian snow course data are mainly located over river basins in southern Canada, and tend to be concentrated in the second half of the water year as most of the snow course data collection programs are designed to monitor peak SWE prior to melt (Brown et al., 2000). The SNOTEL data are available on a daily basis but only values from the first and fifteenth of the month were used to approximate the observing frequency of the Canadian data. A total of approximately 1,800 snow course and SNOTEL sites were available (spatial distribution shown in Fig. 5). Strictly speaking, the Canadian data are not totally independent as the snow course depth observations were used in the analysis. However, these observations were usually only available on a semi-monthly basis, and would not have exerted a strong impact on the snow depth analysis in areas with regular daily snow depth observations.

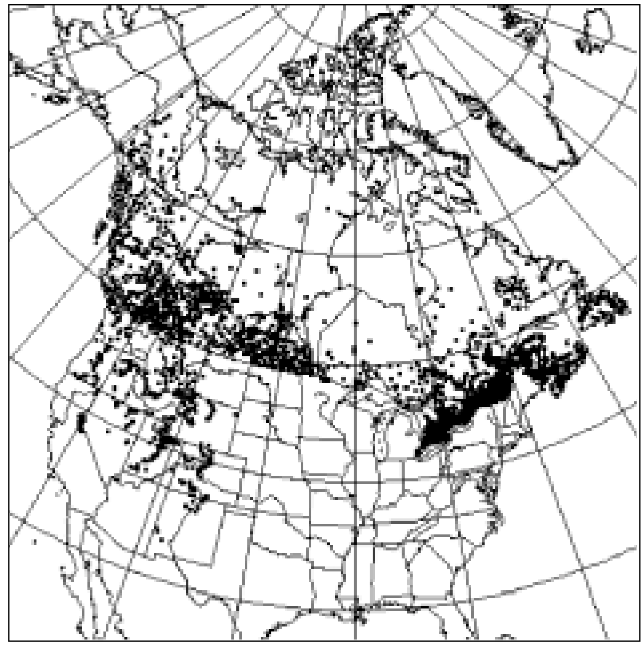

Fig. 5 The spatial distribution of Canadian snow course and U.S. SNOTEL snowpillow sites available for SWE evaluation. The plot shows the distribution of observations for March 1982.

In order to facilitate the SWE comparison, the snow course data were interpolated to a 1° latitude/longitude grid using an inverse-distance weighting scheme that excluded observations outside ±400 m of the mean gridpoint elevation. The gridding was carried out for 15 snow seasons during the period 1980/81 – 1994/95. The resulting gridded data had extensive gaps. Incorporating SWE estimates from Braaten (unpublished manuscript) in the gridding process eliminated some of these gaps. These estimates were derived from blending snow depth observations with observed density information, and as such, are less “independent” than the snow course SWE observations. Two levels of comparison were carried out: (1) comparison of SWE and density climatologies along a north-south and east-west transect through the continent, and (2) comparison of interannual variability in regionally-averaged SWE for two areas with a good spatial distribution of snow course observations over relatively homogeneous terrain. The two regions selected were southern Saskatchewan (49°–52°N, 100°–105°W), and southern Ontario (42°–45°N, 75°–85°W).

The climatology comparison was carried out during the month of March when the spatial distribution of snow course observations across Canada was at a maximum. Two transects were taken through the CMC and snow course climatology grids along 53ºN and up 120ºW. It would have been preferable to have the north-south transect closer to the centre of the continent, but there were insufficient snow course observations to allow this. The results of the east-west transect (Fig. 6) show a good level of agreement, particularly for SWE. The CMC results exhibit greater spatial variability than the snow course observations, as the grid resolution is much

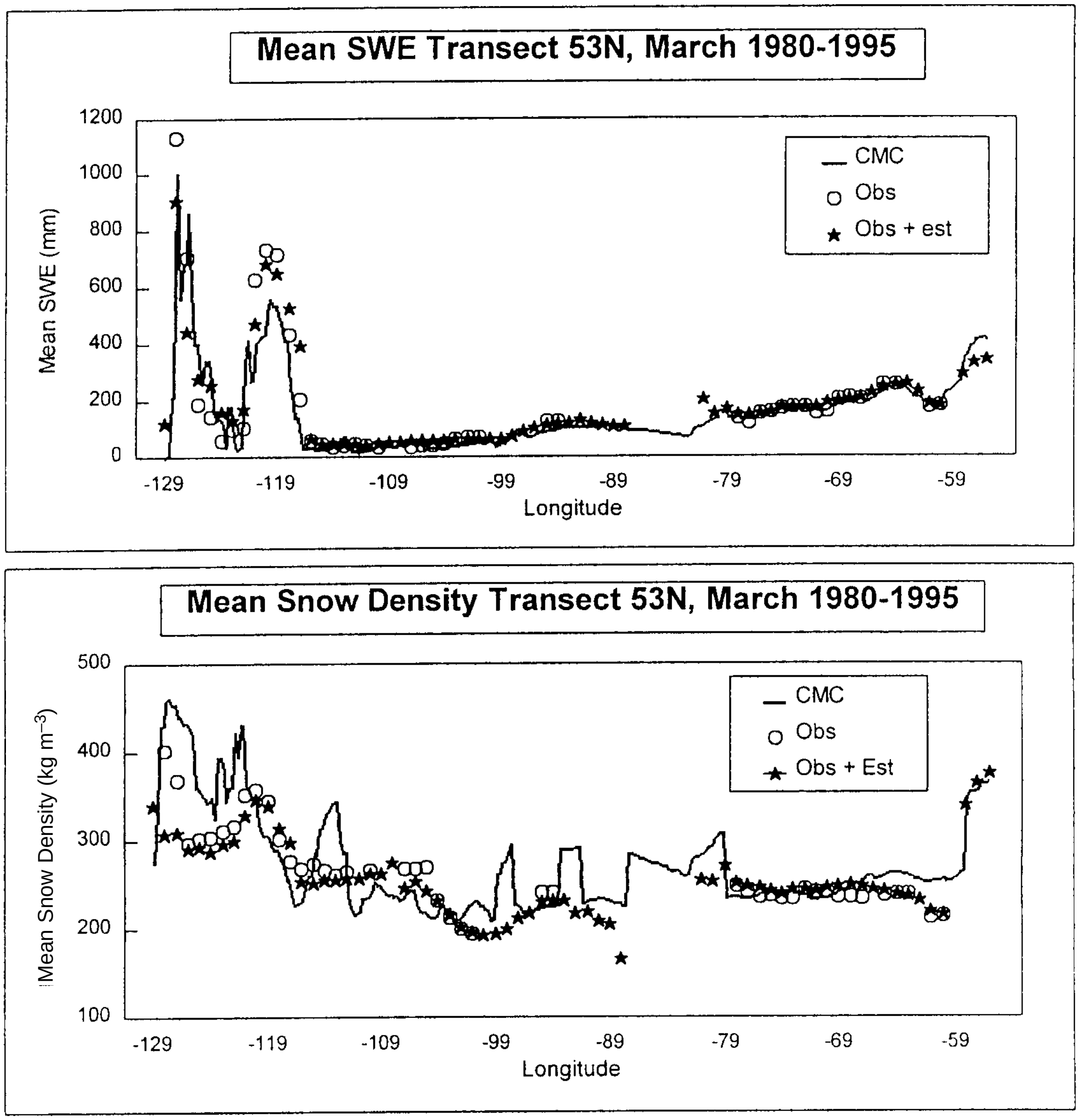

Fig. 6 Comparison of east-west transect of mean SWE along latitude 53ºN. Open circles are gridded SWE values from snow course observations; stars are gridded SWE results incorporating SWE estimates from Braaten (1997).

finer (0.25° versus 1°). Density agrees well with the available observations over most of southern Canada with the exception of the western cordillera where density is overestimated by 30–50%. The depth-dependent aging scheme used in the CMC snow model does not appear to work well in this high snow accumulation region – processes such as depth hoar formation are not considered in the model, and this may account for some of the observed bias. SWE is simulated well along the north-south transect (Fig. 7) with the exception of values over Oregon (42°–45°N) which are much lower than the observed SNOTEL data. This is likely related to the location of the SNOTEL sites in higher elevation areas, and underscores the difficulty of comparing datasets with differing spatial resolutions. The density simulation for the north-south transect does not agree at all well with the observations; densities are too high over the boreal forest zone, and do not exhibit the observed step to higher values north of the treeline. Improving the representation of snow aging would most likely require a more detailed snowpack model and additional input parameters to those used in this analysis.

Comparison of regionally averaged SWE time series for southern Saskatchewan and southern Ontario for the month of February for the 1980–95 period (Fig. 8) revealed that the two series were closely correlated during the 16-year period. Brown (2000) also found that snow depth data provided good estimates of interannual variability in SWE over the Canadian prairie region using fixed seasonal values for snow density. Brown (2000) found no evidence for secular variability or trends in snow density over southern Canada for the 1964–93 period which suggests that detailed simulation of snow density may not be necessary to reconstruct SWE successfully from snow depth data. However, use of a climatological

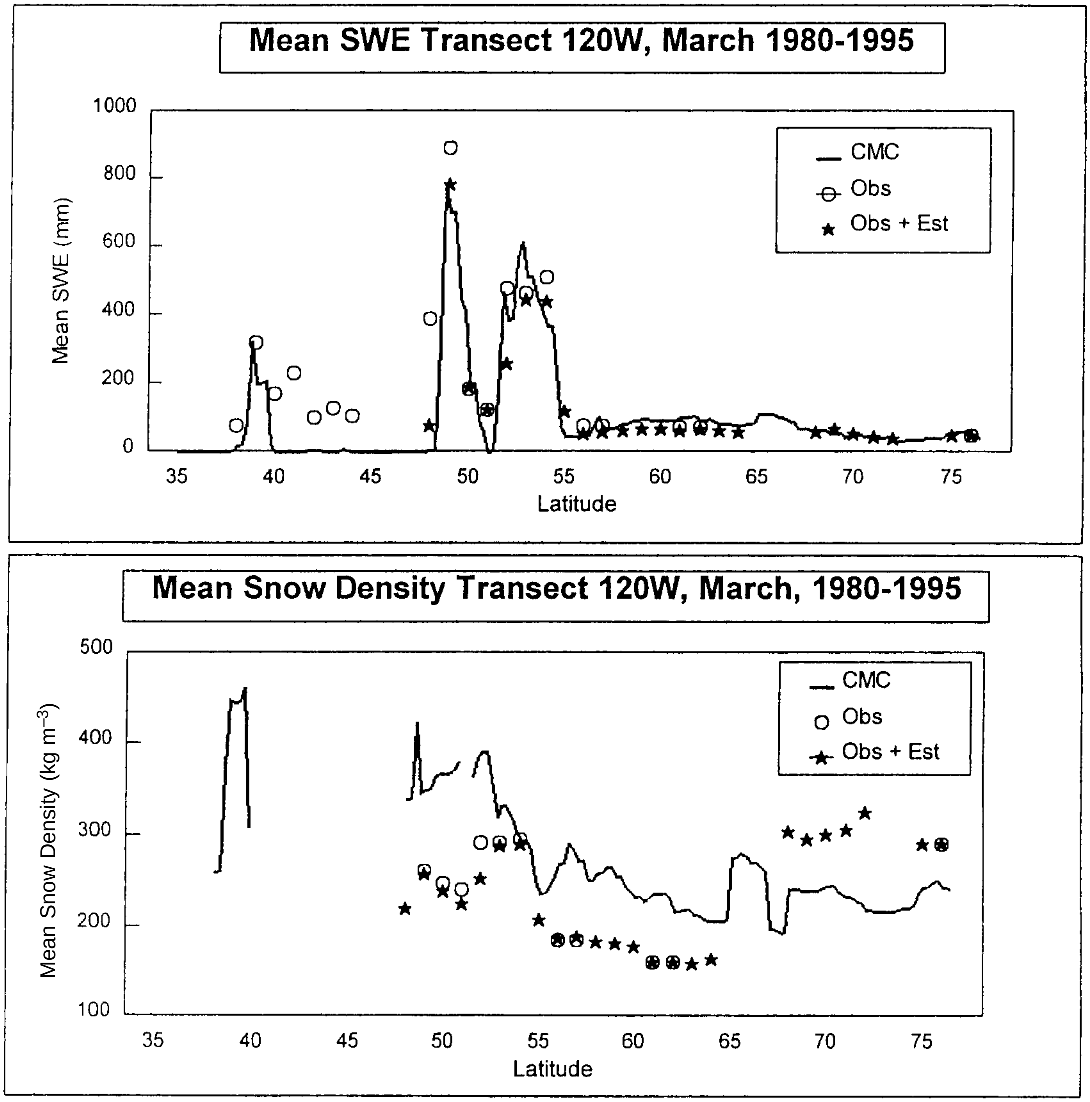

Fig. 7 Comparison of north-south transect of mean SWE along longitude 120ºW. Open circles are gridded SWE values from Canadian snow course observations and U.S. SNOTEL data; stars are gridded SWE results incorporating SWE estimates from Braaten (1997).

average snow density in estimating SWE would not take account of events such as winter thaws and rain-on-snow that would have a major impact on snowpack density.

A comparison of the SWE estimates with those obtained from a previous run of the analysis with a simpler snowpack model (not shown) revealed marked improvements in the simulation of SWE over mountain regions, and the elimination of a positive winter SWE bias for the southern Saskatchewan evaluation region described above. However, the enhanced snowpack model did not perform as well along the Labrador coast where SWE was overestimated.

5 Conclusions

An extensive network of daily snow depth observations (~8000 stations) exists over much of the contiguous United States and southern Canada. These data were used in a snow depth analysis scheme following Brasnett (1999), to develop monthly mean fields of snow depth and estimated SWE for North America for the AMIP II period (1979–96). In observation rich areas, SWE information was derived by applying an estimate of snow density (computed as a function of age, snow depth and air temperature) to the objectively analysed snow depths. In data sparse areas, snow depth and SWE were estimated from an empirically based snowpack model driven with 6-hourly temperature and precipitation fields from ECMWF.

Evaluation of the historical analysis with independent in situ and satellite data revealed that the gridded dataset was able to capture successfully the important features of the North American snow cover climate such as continental-scale variation in SCE and SWE. The snow depth climatology revealed a number of improvements over the Foster and Davy (1988) product, namely an improved representation of the snow line in June and October, and a more realistic spatial

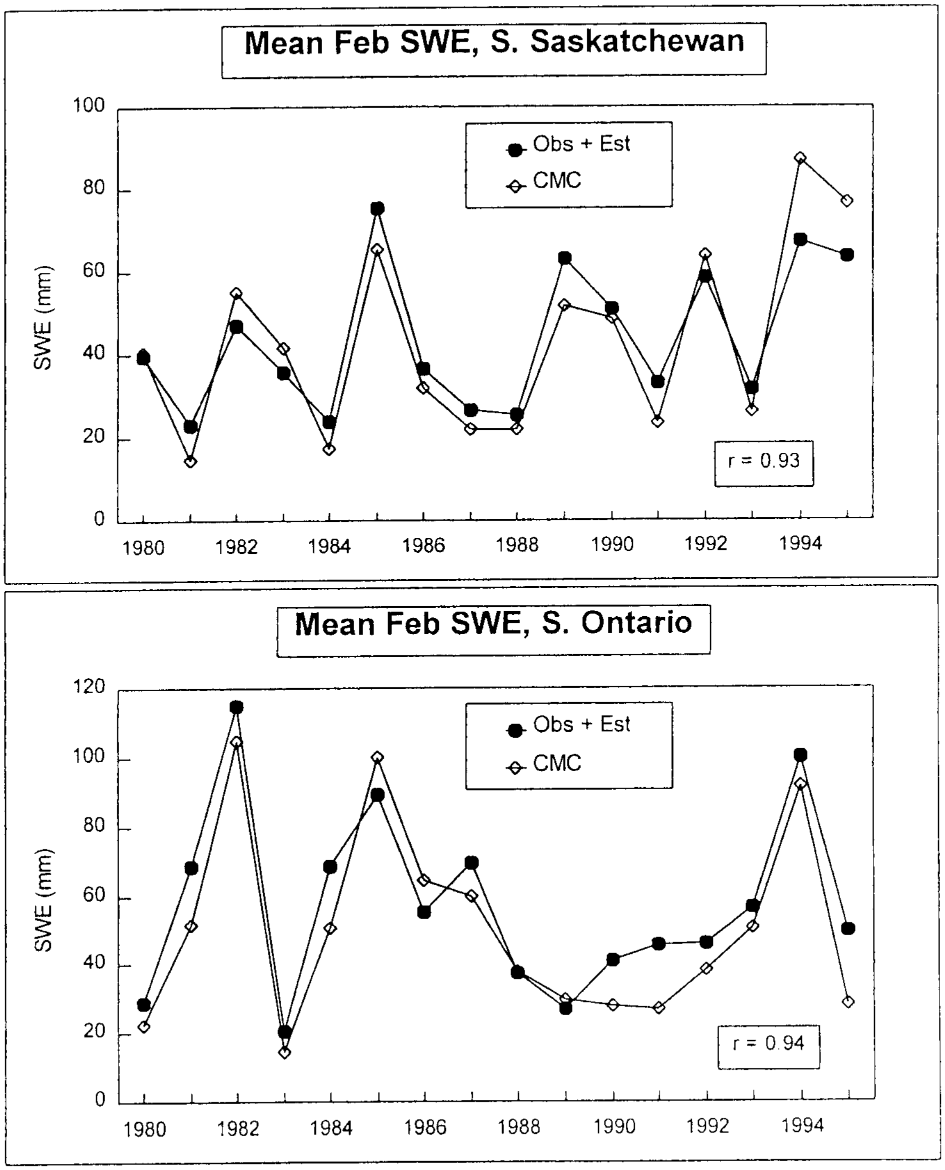

Fig. 8 Comparison of interannual variability in areally-averaged SWE for two regions of southern Canada with dense snow course networks. Observed SWE values and SWE estimates from Braaten (1997) were used in the comparison to obtain a continuous time series for the period.

distribution of snow over the western cordillera. The dataset successfully captured interannual variability in SCE and SWE during the November–April period, but was less successful in the May–October period when the snowline was located over data sparse regions. Overly rapid melt of snow in the spring contributed to this problem at high latitudes. The gridded snow depth and SWE dataset represents an important new source of information for the evaluation of climate and hydrological models, satellite algorithm development, and climatological applications. For example, the CMC SWE dataset is currently being used to investigate discrepancies in SWE retrievals between SMMR and SSM/I over the Canadian prairies (Derksen et al., 2002b).

Cross-validation of the snow depth and SWE results, for example, through systematic withholding of observations, would have been desirable for characterizing the error in the analysis methodology. This was not practical with the server available for this project, but will be investigated at a future date. There is also some interest at CMC in developing an objectively analysed SWE product that incorporates SWE observations. The background field for this product would use simulated SWE values from the Interaction-Surface-Biosphere-Atmosphere (ISBA) surface scheme run in the Canadian regional weather forecast model (Bélair et al., 2002).

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