Implementation of a Large-Scale Variable Velocity River Flow Routing Algorithm in the Canadian Regional Climate Model (CRCM)

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ABSTRACT Implementation and validation of a flow routing scheme for the North American domain of the Canadian Regional Climate Model (CRCM) is described. A variable velocity flow routing algorithm is used to transport runoff from the land surface to the continental edges and provide freshwater flux forcing for the oceans. The flow routing scheme uses Manning’s equation to estimate flow velocities for river channels whose cross-sections are assumed to be rectangular. Discretization of major North American river basins and their flow directions are obtained at the polar stereographic resolution of the CRCM using 5-minute global river flow direction data as a template. In the absence of observation-based gridded estimates of runoff, model runoff estimates from a global simulation of the Variable Infiltration Capacity (VIC) hydrological model (forced with observation-based meteorological data) are used to validate the flow routing scheme. Model results show that the inclusion of flow routing improves the comparison with observation-based streamflow estimates when compared to the unrouted runoff. Monthly comparison of simulated streamflow with observation-based estimates, and basin-wide averaged flow velocities, suggests that the flow routing scheme performs satisfactorily.

RÉSUMÉ L’article décrit la mise en œuvre et la validation d’un schéma de routage du ruissellement pour le domaine nord-américain du Modèle régional canadien du climat (MRCC). Un algorithme de routage du ruissellement à vitesse variable est utilisé pour simuler le transport du ruissellement à la surface de la terre jusqu’à la marge continentale et fournir le flux d’eau douce pour le forçage des océans. Le schéma de routage du ruissellement utilise l’équation de Manning pour évaluer la vitesse d’écoulement dans un chenal fluvial dont la coupe transversale est présumée être rectangulaire. Les principaux bassins hydrographiques nord-américains ont été discrétisés et la direction d’écoulement de leurs rivières a été déterminée selon la résolution stéréographique polaire du MRCC, en utilisant comme modèle des données sur la direction d’écoulement globale des rivières à une résolution de 5 minutes. En l’absence d’estimations maillées du ruissellement, on a eu recours à des estimations de ruissellement établies à partir d’une simulation globale par le modèle hydrologique à capacité d’infiltration variable (VIC) (forcée à l’aide de données d’observations météorologiques) pour la validation du schéma de routage du ruissellement. Les résultats de la modélisation montrent que l’incorporation du routage du ruissellement améliore la concordance avec les estimations de l’écoulement fluvial fondées sur des observations, comparativement au ruissellement sans routage. Une comparaison du ruissellement mensuel simulé et du ruissellement estimé à partir d’observations ainsi que la moyenne des vitesses d’écoulement à l’échelle du bassin laissent croire que le schéma de routage du ruissellement fonctionne de façon satisfaisante.

1 Introduction

Land surface runoff from river basins, which flows through river networks to the continental edges, acts as an important forcing for the ocean. With lower density than ocean saltwater, freshwater from rivers floats at the surface and freezes at a higher temperature. The change in ocean salinity, due to freshwater flux from rivers, affects the thermohaline circulation and also sea-ice formation at high latitudes. Both thermohaline circulation and sea-ice formation have an important impact on climate because they influence the energy, water, and momentum fluxes at the ocean-atmosphere interface. Freshwater flux from rivers can be included in a climate model by implementing a routing algorithm that simulates the river flow in channel streams and transfers runoff from land to the continental edges. Runoff used as input to the flow

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routing scheme is provided by the land surface module in the climate model which partitions precipitation into soil moisture storage, evapotranspiration and runoff.

In addition to providing freshwater forcing to the ocean at continental edges, incorporation of river routing in climate models is important for at least two other reasons. First, streamflow is a spatial integrator of hydrological processes which is useful for assessing land surface schemes and climate model simulations at large spatial scales (Wood et al., 1998; Arora et al., 2000; Arora, 2001). Since streamflow is routinely measured it can be used to evaluate the performance of atmospheric models on a climatological basis by comparing simulated and observed streamflow at river mouths or at gauging stations within the basins. Second, the ability to simulate streamflow reliably within atmospheric models has water management implications. The implementation of flow routing in climate models provides an opportunity to study the impact of climate change on simulated streamflow and the hydrology of major river basins (Arora and Boer, 2001).

The coupling of atmospheric and hydrologic models also provides opportunities for flash flood studies and assessment of model and radar estimates of precipitation. Atmospheric and hydrologic models have been coupled in a number of investigations to simulate flash floods (Lin et al., 2002; Westrick and Mass, 2001; Droegemeier et al., 2000; Yates et al., 2000; Warner et al., 2000). In these short-duration (usually 2–7 days) and high-resolution (~5–10 km) studies, runoff from an atmospheric model is used as input to a hydrological model over small regions. These investigations have demonstrated the feasibility of using coupled atmospheric-hydrologic models for flash flood forecasts. Benoit et al. (2000) use coupled atmospheric-hydrologic models to validate precipitation and evaporation fields from their atmospheric model. They demonstrate that using streamflow, which provides a spatially and temporally integrated response of large river basins to atmospheric forcing, is a better approach to assessing model precipitation over a region than comparisons with observations at discrete stations. Yates et al. (2000) use comparisons between simulated flood discharge estimates and observations to assess the adequacy of rainfall estimates obtained from radar, a dynamic model and an automated algorithmic system.

This paper describes the implementation of a river flow routing scheme in the Canadian Regional Climate Model (CRCM) and provides an assessment of digital river networks and simulated streamflow obtained, at the CRCM resolution, for major North American river basins in an offline mode. The implementation of flow routing in the CRCM will yield a fully coupled atmospheric, hydrological, and ocean modeling system at regional scales. This coupling will also replace the use of climatological freshwater flux data to provide freshwater forcing to the ocean grid cells at the continental edges. The inclusion of river flow routing also yields a climate modelling framework in which, like other climate variables, streamflow from major river basins is simulated on a regular basis. Section 2 provides a brief description of the routing scheme. The routing scheme implemented in the CRCM is the variable velocity flow routing scheme of Arora and Boer (1999) (hereinafter referred to as AB99), which explicitly relates river flow velocity to slope, discharge, and river cross-section simultaneously, using Manning’s equation. Most of the other existing approaches either assume a uniform constant flow velocity or transfer coefficient (Miller et al., 1994; Coe, 1998; Oki, 1997; Coe et al., 1998; Oki et al., 1996; Olivera et al., 1998; Vorosmarty and Moore, 1991) or use simple formulae that use time-independent flow velocities parametrized as a function of the topographic gradient (Miller et al., 1994; Coe, 1997; Marengo et al., 1994; Costa and Foley, 1997; Sausen et al., 1994; Hagemann and Dömenil, 1998). Section 3 describes the methodology adopted for obtaining model parameters and modifications made to the river flow model to deal with the regional polar stereographic domain of the CRCM, which covers most of North America. Section 4 presents results from an offline validation of the flow routing scheme. The routing scheme is driven with runoff data from the Variable Infiltration Capacity (VIC) hydrologic model. Simulated streamflow obtained from the routing model is compared with observation-based streamflow for the eight major North American rivers. In Section 5 the results of the study are summarized and conclusions are drawn.

2 The flow routing scheme

The variable velocity flow routing algorithm of AB99 is used to perform flow routing at the standard 45-km resolution of the CRCM and is described here briefly (routing scheme details are provided in the Appendix). This scheme has been successfully implemented in the coupled general circulation model (GCM) developed at the Canadian Centre for Climate Modelling and Analysis (CCCma). The implementation of the routing scheme in the CRCM is different from that in the CCCma GCM. In the CRCM, the routing scheme is implemented at a finer resolution (~45 km compared to ~375 km in the GCM) and on a polar stereographic projection (compared to a latitude-longitude projection in the GCM). In addition the Great Lakes, which are represented by only two grid cells in the GCM, are treated in a detailed manner in this study.

Figure 1 shows the schematic of the flow routing algorithm. The flow routing scheme consists of surface and groundwater reservoirs, which obtain daily estimates of surface runoff and drainage inputs, respectively, simulated by the land surface scheme (see Fig. 1). The cross-section of the river channel is assumed to be rectangular, and the river width is obtained using a geomorphological relationship between width and mean annual discharge (as discussed in Section 3). Manning’s equation is used to estimate surface river channel flow velocity and hence the travel time between the grid cells, while the residence time in the groundwater reservoir is related to the dominant soil type in a CRCM grid cell following Arora et al. (1999).
3 Estimating flow routing parameters for the CRCM grid

The CRCM is a limited-area regional climate model. It uses the semi-implicit, semi-Lagrangian scheme to solve the fully elastic non-hydrostatic Euler equations (Caya and Laprise, 1999). The CRCM uses the complete subgrid-scale physical parametrization package of the second-generation CCCma atmospheric GCM (McFarlane et al., 1992). At its boundaries, the model may be driven by data from a GCM or from atmospheric analysis. The model’s horizontal grid uses a polar stereographic projection with a typical 45-km grid mesh and 30 vertical levels.

The river flow routing parameters are estimated for the CRCM computational domain, which covers most of North America in its operational configuration (see Fig. 2). Estimation of routing parameters essentially consists of describing river basins, and obtaining their digital river flow networks, at the model’s resolution and projection. Once river flow directions have been obtained, other model parameters such as distance to the downstream cell, channel slope and channel width are estimated for every grid cell in a relatively straightforward manner.

a Discretization of Major River Basins

The discretization of the major North American river basins and their river flow directions at the CRCM resolution were obtained using global 5-minute basin discretization and the flow direction dataset of Graham et al. (1999; hereinafter referred to as G99). Although flow routing could have been performed at 5′ resolution using G99’s river flow directions, it was deemed desirable to have the resolution and projection of the routing model identical to the climate model. This avoids inconsistencies due to the differences in model resolutions and in particular the interpolation of the CRCM runoff to a different routing model resolution, and the interpolation of simulated streamflow back to the CRCM resolution. River basin discretizations and flow directions were scaled up from the G99 5′ dataset to the CRCM resolution. A CRCM grid cell is assigned to a river basin as follows: (1) the 5′ cells, corresponding to different river basins based on the G99 dataset, lying within each CRCM cells are counted; (2) the river basin with the maximum number of 5′ cells is assigned to a CRCM cell; and (3) the resulting areas of the river basins at the CRCM resolution are compared with the original areas at 5′ resolution. The G99 dataset identifies 55 major river basins of the world, and 10 of these lie within the CRCM domain in this study (see Fig. 2). Only 8 of these 10 major North American river basins lie completely within the CRCM domain. The CRCM’s drainage areas of major river basins are compared with the drainage areas of G99 at 5′ resolution in Table 1. The CRCM drainage areas for the Colorado and Rio Grande rivers do not compare well with the G99 values since parts of these two river basins lie outside the CRCM domain.

b Obtaining Digital River Flow Networks at the CRCM Resolution

To obtain river flow directions at the CRCM resolution, we use G99’s 5′ flow directions as a template and a methodology, shown in Fig. 3, that requires minimal manual intervention (Lucas-Picher et al., 2002). The original 5′ flow directions were obtained by using mean elevations of the grid cells and assuming that the water drains in the direction characterized by the steepest slope. At spatial scales comparable to the CRCM grid cells, the mean elevations of grid cells are not representative of their river flow directions. This is because rivers tend to flow in localized areas of low elevation, and the mean elevations of the grid cells are not representative of river bed profiles. Interpolation of 5′ flow directions at the CRCM resolution is also difficult since the projection of the CRCM is polar-stereographic, while the 5′ flow directions...
are projected on a latitude-longitude grid. Therefore river flow directions are assigned to CRCM grid cells as follows: (1) the northern, eastern, southern and western corners of the CRCM grid cell are identified; (2) these corners are used to outline a latitude-longitude grid box that encloses the CRCM grid cell; (3) a uniformly distributed unit amount of runoff is generated within this new grid box and allowed to find its way out of the box following the 5’ flow directions; (4) the direction, in which the maximum amount of water is drained, is identified; and (5) by identifying the CRCM grid cell that receives this outflow, a river flow direction is assigned. The methodology used here for discretizing river basins and obtaining flow directions is an extension of the one used by AB99.

Tests are also performed to ensure that water does not flow between basins, and that flow directions are not assigned such that runoff keeps revolving within a basin. The river flow direction algorithm performs checks to ensure that all grid cells eventually drain to ocean cells. As an example, Fig. 4 shows the flow directions for the Mackenzie River basin at the CRCM resolution.

The adequacy of digital flow networks obtained in this manner is assessed by visually comparing the digital river network plots with actual river networks from atlases and global flow networks from other observation-based datasets. To visualize the digital river network obtained at the CRCM resolution, a river “order” graph is plotted. The grid cell where a stream originates is assigned an order of 1. The union of two streams of order $n$ creates a stream of order $n+1$. The union of two streams of order $n$ and $m$ is assigned the order

<table>
<thead>
<tr>
<th>River Basin</th>
<th>Graham et al. (1999)</th>
<th>CRCM</th>
<th>Percentage Difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mississippi</td>
<td>3,218,720</td>
<td>3,228,415</td>
<td>0.30</td>
</tr>
<tr>
<td>Mackenzie</td>
<td>1,735,635</td>
<td>1,748,230</td>
<td>0.73</td>
</tr>
<tr>
<td>Nelson</td>
<td>1,303,641</td>
<td>1,305,674</td>
<td>0.16</td>
</tr>
<tr>
<td>Columbia</td>
<td>1,106,969</td>
<td>1,106,439</td>
<td>–0.05</td>
</tr>
<tr>
<td>St. Lawrence</td>
<td>1,090,564</td>
<td>1,087,952</td>
<td>–0.24</td>
</tr>
<tr>
<td>Yukon</td>
<td>884,867</td>
<td>882,128</td>
<td>–0.31</td>
</tr>
<tr>
<td>Rio Grande*</td>
<td>856,547</td>
<td>149,144</td>
<td>–82.59</td>
</tr>
<tr>
<td>Colorado*</td>
<td>770,829</td>
<td>616,827</td>
<td>–19.98</td>
</tr>
<tr>
<td>Churchill</td>
<td>296,190</td>
<td>300,800</td>
<td>1.56</td>
</tr>
<tr>
<td>Fraser</td>
<td>262,854</td>
<td>260,419</td>
<td>–0.93</td>
</tr>
</tbody>
</table>

*The CRCM drainage areas for the Colorado and Rio Grande do not compare well with Graham et al. (1999) values since these two river basins do not lie completely within the CRCM domain.
Max \((n, m)\). Figure 5 compares the digital flow network obtained at the CRCM resolution with the flow network based on the Simulated Topological Network (STN-30p) dataset of Vorosmarty et al. (2000). Note that the projection of the CRCM flow network is polar stereographic, while the STN-30p flow network is projected onto a latitude-longitude grid. Nevertheless, Fig. 5 shows that the river flow directions thus obtained maintain the flow networks for major North American rivers reasonably well despite the rather coarse CRCM resolution.

\[ L_f = 1.024 - 0.077 \ln \left( \frac{W_f}{W_c} \right) \]

where \(L_f\) and \(L_c\) are river lengths at the finer and coarser resolutions respectively, and \(W_f\) and \(W_c\) are the finer and coarser horizontal grid spacings, approximately equal to 10 km (for the 5° G99 dataset) and 45 km for the CRCM.

River bed slope \(s\), also used in Eq. (13), is estimated using mean elevations of the upstream and downstream grid cells and the distance between these cells at the CRCM resolution. As mentioned in Section 3b, mean elevations of grid cells at the CRCM resolution are not expected to represent the longitudinal profile of the river bed. Use of mean elevation to estimate river slope, thus leads to about 20% of grid cells having negative slope values. Negative or zero slopes physically impede the flow and, following AB99, these values were therefore replaced by a minimum threshold value of 0.001. Rather than selecting a minimum flow velocity as in other models (e.g., Miller et al., 1994), we chose a minimum slope so that flow velocity is still modelled as a function of amount of runoff. Figure 6 shows the spatial distribution of the river slope parameter over the CRCM domain. The western part of the North American continent with significant topography, in particular the Rocky Mountains, is characterized by higher values of river slopes, and similarly in the east for the Appalachian Mountains. The Mississippi River basin plains and the river basins draining into Hudson Bay are characterized by relatively flat topography and therefore low values of river slope.

**d River Width**

An effective river width is also required in Eq. (13) and this is obtained using a geomorphological relationship between annual mean discharge \((Q_m, \text{m}^3 \text{s}^{-1})\) passing through a river section and river width \((W, \text{m})\). Following AB99, Eq. (2) is
used to estimate river width, $W$, at every point along the digital river network for the eight major North American river basins:

$$W = \max \left( 10, ZQ_m^{0.5} \right)$$

$$Z = (10^{-4}Q_{m,\text{mouth}} + 6.0)$$

(2)

where $Q_m$ is the annual mean discharge passing through a river section, and the subscript $\text{mouth}$ represents $Q_m$ at the river mouth. For grid cells that do not belong to any major river basin, and for basin cells whose river mouth does not lie within the CRCM domain, a slightly different relationship is used:
Geomorphological relationships between width and discharge, which describe the morphology of a river channel, are discussed in geomorphological literature (Leopold et al., 1964; Richards, 1985). A minimum river width is prescribed to be 10 m. Estimates of annual mean discharge passing through each grid cell are obtained using the digital flow network and global runoff dataset from Cogley (1998). Sensitivity of flow velocity to river width has shown that an error of 100% in the prescribed river width results in a corresponding error of about 25% in flow velocity (AB99). In addition, an error in river width does not change the amount of runoff, just the flow velocity and thus the time it takes for runoff to reach the river mouth. Table 2 shows the river widths at the mouths of major North American rivers estimated from Eq. (3), with two observation-based values from AB99.

**Groundwater Residence Time**

The groundwater reservoir is designed to receive drainage input from the land surface scheme. The outflow of groundwater from the groundwater reservoir into the surface river channel reservoir is designed to simulate the slow seepage of groundwater into the surface channel. Following Arora et al. (1999) and AB99, the residence time of water in the groundwater reservoir is assigned based on the dominant soil type in a CRCM grid cell; groundwater delay factors of 10, 30 and 60 days are associated with coarse (sand), medium (silt), and fine (clay) textured soils, respectively. Arora et al. (1999) estimated the groundwater delay factors from the receding sections of hydrographs from sub-catchments within the Amazon and Mississippi basins. Their analysis indicates that simulated streamflow is fairly sensitive to the groundwater delay factor, with a higher value producing larger attenuation of inflow runoff. Grid cells characterized by ice (such as the glacial cells in Greenland and northern Canada) or rock are not assigned a groundwater delay factor since no drainage to the groundwater occurs from these cells. The major soil types in CRCM grid cells are estimated using Wilson and Henderson-Sellers’ (1985) soil dataset. Figure 7 shows the major soil types within the grid cells at the 45-km grid mesh of CRCM for the North American domain.

\[
W = (0.00013Q_m + 6.0)Q_m^{0.5}.
\]  

(3)

Table 2. River width at the mouth of the major rivers and basin-wide averaged mean annual flow velocity for each basin.

<table>
<thead>
<tr>
<th>River basins</th>
<th>Model Width (m)</th>
<th>Observation-based estimate*</th>
<th>Flow velocity (m s(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mackenzie</td>
<td>1608.6</td>
<td>1600</td>
<td>0.92</td>
</tr>
<tr>
<td>Mississippi</td>
<td>1065.8</td>
<td>1250</td>
<td>1.15</td>
</tr>
<tr>
<td>St. Lawrence</td>
<td>835.8</td>
<td>75</td>
<td>0.75</td>
</tr>
<tr>
<td>Yukon</td>
<td>689.6</td>
<td></td>
<td>1.65</td>
</tr>
<tr>
<td>Columbia</td>
<td>684.9</td>
<td></td>
<td>0.94</td>
</tr>
<tr>
<td>Fraser</td>
<td>478.4</td>
<td></td>
<td>1.73</td>
</tr>
<tr>
<td>Nelson</td>
<td>389.1</td>
<td></td>
<td>1.05</td>
</tr>
<tr>
<td>Churchill</td>
<td>194.9</td>
<td></td>
<td>1.22</td>
</tr>
</tbody>
</table>

*from Arora and Boer (1999).
Treatment of the Great Lakes

The methodology described so far in Section 3 explains the manner in which river flow is routed over land. In the St. Lawrence River basin, however, flow is also routed through the five Great Lakes (lakes Superior, Michigan, Huron, Erie and Ontario). Like any other reservoir, these lakes have a damping effect on the flow. Large reservoirs damp the outflow via their storage such that the outflow hydrograph is a strongly attenuated form of the inflow hydrograph. Detailed hydrological modelling of flow routing via the Great Lakes would require information about the lake bathymetry, the differences in water levels, and characteristics of natural and man-made hydraulic structures over which the water flows when it is transferred between lakes. Since the primary objective of coupling the CRCM to a flow routing model is to transfer runoff from land grid cells to ocean cells and not to perform detailed hydrological modelling for any particular river basin, we parametrize the damping effect of lakes in a relatively simple manner. We consider each lake grid cell, at the CRCM resolution, as a reservoir on its own, whose discharge is a linear function of its storage. The storage and discharge from lake cells are thus modelled similar to those of the groundwater reservoir. Figure 8 shows the flow directions on the CRCM grid for the St. Lawrence River basin and adjacent areas, and how the flow is routed via the lake cells. Flow directions are prescribed not only for the land cells but also for the Great Lakes cells. A lake cell may receive inflow from a land cell draining into the lake, another lake cell, or both. Prescription of flow directions over the Great Lakes cells (based on G99’s 5’ flow direction dataset) ensures that flow continuity is maintained and that the damping effect of the lakes is taken into account. A residence time of 550 days is prescribed for each lake cell which results in a realistic attenuation of outflow hydrograph.

4 Validation of the routing scheme

Testing and validation of the flow routing scheme requires gridded estimates of observed daily runoff. In the absence of
observed data, model-simulated runoff can be used to test the routing scheme. CRCM-simulated runoff could have been used but these runoff values would depend on CRCM precipitation and the way in which this precipitation is processed into evapotranspiration and runoff by the CRCM land surface module. It would be difficult to distinguish between errors in CRCM runoff and errors in flow routing. Here we use runoff fields from the VIC hydrological model (Lohmann et al., 1998) driven with observation-based meteorological data. Nijssen et al. (2001a) performed a 14-year (1980–1993) global simulation with the VIC model at a 2° latitude-longitude resolution. In this simulation, the VIC model was forced with observation-based meteorological data that are primarily based on the National Centers for Environmental Prediction (NCEP) reanalysis. We use gridded estimates of daily runoff from this simulation projected on the CRCM’s polar-stereographic 45-km resolution to test the routing scheme. The VIC model parameters have been calibrated for major river basins (including the Mackenzie and Mississippi) to yield reliable runoff estimates, although the final calibration does not completely eliminate biases in annual runoff (Nijssen et al., 2001b). In the absence of observed estimates of gridded daily runoff, model runoff estimates from a calibrated hydrological model driven with observation-based meteorological data appear to be the best alternative. Nevertheless, as with any other model estimates, these runoff data are likely to contain errors which make it difficult to assess the routing scheme adequately.

The VIC model runoff data provide estimates of total runoff but do not provide separation of total runoff into surface runoff and drainage. The routing scheme used in this study is designed to account for surface runoff and drainage estimates from the land surface scheme separately. Since most land surface schemes simulate a much higher fraction of drainage than surface runoff, all runoff is assumed to be drainage. Arora and Boer (2002), for example, estimate that on a global average basis, the Canadian Land Surface Scheme (Verseghy, 2000) used in the CCCma’s third-generation atmospheric GCM allots about 80% of the total runoff into drainage.

Eleven years (1980–1990) of daily runoff data from the VIC model are used as input to the routing scheme in an offline mode. In order to “spin-up” the model surface and groundwater reservoirs, these data are input to the routing scheme twice, and the results are analysed only from the second 11 years. We first compare annual mean runoff from the VIC model, for the eight major North American river basins that lie completely within the CRCM domain, with observation-based estimates. Clearly, if annual mean runoff simulated by the VIC model does not compare well with observations, then we cannot expect realistic streamflow simulations from the routing model. Two estimates of observation-based mean annual basin-wide averaged runoff are used. First, based on basin-discretization at the CRCM resolution, we obtain an estimate of observed annual mean runoff using the 1° global runoff dataset of Cogley (1998). Second, we use observed streamflow data from the United States Geological Survey (USGS) dataset (USGS, 2002) and the Global River Discharge (RivDis) dataset of Vorosmarty et al. (1998) for the American rivers and from the Canadian National Water Data Archive (HYDAT) dataset for Canadian rivers (HYDAT, 1999) for the gauging station closest to the river mouth. Location of the gauging stations, their upstream drainage areas, and the years for which streamflow data are used are shown in Table 3. Comparisons of annual mean basin-averaged runoff from the VIC model with these two observation-based estimates are presented in Fig. 9. The two observation-based estimates are not in agreement for all river basins. Cogley’s (1998) annual mean runoff is larger than the runoff based on observed streamflow data for the Churchill, Fraser, St. Lawrence and Yukon river basins. Compared to the annual runoff estimates based on observed streamflow data, the VIC model runoff compares well for the Fraser, Mackenzie, Mississippi and Nelson river basins. The VIC model runoff is fairly large for the Churchill and Yukon river basins, and small for the Columbia river basin. The VIC model runoff is higher for the Yukon and lower for the Columbia river basins because of the problems associated with precipitation forcing (Nijssen et al., 2001b). We will therefore not perform additional analysis for the Churchill, Columbia and Yukon river basins and will focus our attention on the remaining river basins. We also do not consider the Nelson River because its streamflow seasonality is likely influenced by Lake Winnipeg which is not represented in the CRCM domain considered. Simulated monthly streamflows are thus analysed in detail for the Fraser, Mackenzie, Mississippi and St. Lawrence river basins.
Figure 10 compares the mean monthly-simulated streamflow at the mouth of these four river basins with observed streamflow data. The “unrouted” runoff from the VIC model, as if it were to appear instantaneously at the river mouth, is also shown. The primary objective of the routing scheme is to obtain realistic estimates of freshwater flux from the land surface to the ocean, and therefore we only analyse streamflow at the river mouths and do not perform streamflow validation for sub-catchments within the major basins. The routing of runoff induces a lag in streamflow and attenuates the maximum monthly discharge. The lag in routed runoff is a measure of the basin-wide averaged time which the runoff takes to reach the mouth of the river. The attenuation in maximum monthly discharge is caused by the storage effect of surface and groundwater reservoirs. The seasonality of simulated streamflow for the Fraser, Mackenzie and Mississippi rivers compares reasonably well with observations. The correlation coefficient of unrouted and routed streamflow with observed streamflows is given in Table 4: it shows that the inclusion of routing improves streamflow simulations in general.

Simulated streamflow for the St. Lawrence River is significantly attenuated due to routing of runoff via the Great Lakes and the large delay prescribed for the Great Lakes’ cells.

We further evaluate the streamflow simulations in terms of the error variances for the streamflow values. Following AB99, the error ($e$) or the difference between model-simulat-

**Table 3.** Details about gauging stations for the major North American river basins from which observation-based streamflow data are obtained. CRCM drainage areas are also shown.

<table>
<thead>
<tr>
<th>Basin name</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Gauging station</th>
<th>CRCM</th>
<th>Years</th>
<th>Station name</th>
<th>Data source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Churchill</td>
<td>58°07'N</td>
<td>94°37'W</td>
<td>287000</td>
<td>300800</td>
<td>1980–1990</td>
<td>Red Head Rapids</td>
<td>HYDAT</td>
</tr>
<tr>
<td>Columbia</td>
<td>45°36'N</td>
<td>121°11'W</td>
<td>613830</td>
<td>1106439</td>
<td>1980–1988</td>
<td>The Dalles</td>
<td>RivDis</td>
</tr>
<tr>
<td>Fraser</td>
<td>49°13'N</td>
<td>122°49'W</td>
<td>232000</td>
<td>260419</td>
<td>1983–1990</td>
<td>Port Mann</td>
<td>HYDAT</td>
</tr>
<tr>
<td>Mackenzie</td>
<td>67°27'N</td>
<td>133°44'W</td>
<td>168000</td>
<td>1748230</td>
<td>1980–1990</td>
<td>Arctic Red River</td>
<td>HYDAT</td>
</tr>
<tr>
<td>Mississippi</td>
<td>32°19'N</td>
<td>90°54'W</td>
<td>2964254</td>
<td>3228415</td>
<td>1980–1983</td>
<td>Vicksburg</td>
<td>RivDis</td>
</tr>
<tr>
<td>St. Lawrence</td>
<td>45°00'N</td>
<td>74°47'W</td>
<td>774000</td>
<td>1087952</td>
<td>1987–1990</td>
<td>Cornwall</td>
<td>HYDAT</td>
</tr>
<tr>
<td>Yukon</td>
<td>61°56'N</td>
<td>162°53'W</td>
<td>831392</td>
<td>882128</td>
<td>1980–1988</td>
<td>Pilot Station</td>
<td>USGS</td>
</tr>
</tbody>
</table>

Figure 9 Comparison of mean annual runoff simulated by the VIC hydrological model with observation-based estimates from Cogley (1998) and runoff values based on observed streamflow data, for the eight North American river basins that lie completely within the CRCM domain.
ed (routed) \( f_m \) and the observed \( f_o \) climatological annual cycles of streamflow is written as,

\[
e = f_m - f_o = (\overline{f}_m - \overline{f}_o) + (f'_m - f'_o) = \overline{e} + e
\]

(4)

where \( \overline{f} \) and \( f' \) are the annual mean and deviations from the mean respectively. After some algebra and writing \( \sigma^2 = \overline{f'}^2 \), \( \sigma^2_m = \overline{f'}^2_m \), and \( \sigma^2_o = \overline{f'}^2_o \), the mean-square-error is written as

\[
e^2 = \overline{e}^2 + e'^2 = \overline{e}^2 + \left( \sigma^2_m - \sigma^2_o \right)^2 + 2\sigma_o \sigma_m (1 - R)
\]

(5)

or as a fraction as

\[
\frac{e^2_m}{e^2} + \frac{e^2_o}{e^2} = 1.
\]

(6)

### Table 4.

Comparison of routed and unrouted streamflow with observation-based estimates of streamflow.

<table>
<thead>
<tr>
<th>River Basin</th>
<th>Correlation with observed streamflow (( R ))</th>
<th>( rmse ) (m( ^3 ) s(^{-1} ))</th>
<th>( rmse ) (%)</th>
<th>Percentage of error associated with Annual Cycle</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fraser</td>
<td>Unrouted: 0.38, Routed: 0.79</td>
<td>3238</td>
<td>3321</td>
<td>(Annual Mean: 2.2, Amplitude: 42.4, Phase: 55.4)</td>
</tr>
<tr>
<td>Mackenzie</td>
<td>Unrouted: 0.84, Routed: 0.81</td>
<td>22892</td>
<td>7756</td>
<td>(Annual Mean: 3.1, Amplitude: 46.7, Phase: 50.2)</td>
</tr>
<tr>
<td>Mississippi</td>
<td>Unrouted: 0.87, Routed: 0.89</td>
<td>2940</td>
<td>7759</td>
<td>(Annual Mean: 65.1, Amplitude: 11.0, Phase: 23.9)</td>
</tr>
<tr>
<td>St. Lawrence</td>
<td>Unrouted: 0.24, Routed: 0.62</td>
<td>17137</td>
<td>7420</td>
<td>(Annual Mean: 1.7, Amplitude: 78.3, Phase: 19.9)</td>
</tr>
</tbody>
</table>

Fig. 10 Comparison of observed and model-simulated mean monthly streamflow at the mouth of the Fraser, Mackenzie, Mississippi, and St. Lawrence rivers.
These three terms give the fraction of the mean-square error $\bar{e}^2$, associated with the error in the annual mean, and the amplitude and the phase of the annual cycle. The root-mean-square error, $rmse = \sqrt{\bar{e}^2}$, and the relative $rmse$, $rrmse = \frac{rmse}{\bar{x}}$, are also calculated.

Table 4 shows the $rmse$ and $rrmse$ values and the percentage error associated with the mean, amplitude and phase of the simulated values for streamflow for the four selected major river basins. The average $rmse$ for these rivers is about 35%. This value may be compared with the $rmse$ of approximately 40% for runoff from 11 participating land surface schemes in the Global Soil Wetness Project (GSWP) for areas with good precipitation forcing (Oki, 2001). The percentage errors associated with the annual mean, and the amplitude and phase of the annual cycle about the mean, give further insight into the model simulations. Since the simulated annual mean discharge for the Fraser and Mackenzie rivers compares well with observations (see Fig. 10), little percentage $rmse$ error is associated with the mean. For both of these rivers, most of the error is associated with the phase and the amplitude of the annual cycle. The VIC model simulated annual runoff for the Mississippi River is about 33% higher than the observation-based estimate and the largest fraction of error is therefore associated with the annual mean. The St. Lawrence River has the largest fraction of error associated with the amplitude of the annual cycle because the simulated streamflow shows a slight seasonality while the observed streamflow at the mouth of the St. Lawrence River shows almost no seasonality (because of the damping effect of the Great Lakes and man-made flow regulation by dams along the St. Lawrence River).

Model-simulated annual mean basin-wide averaged flow velocities over land are shown in Table 2, and range from 0.75 m s$^{-1}$ for the St. Lawrence River to 1.73 m s$^{-1}$ for the Fraser River. These velocities fall within the range of velocities (0.14 to 2.13 m s$^{-1}$) obtained by AB99 in a global simulation. While flow velocities also depend on the amount of runoff, Table 2 shows that river basins characterized by high topography and steep river slopes (such as the Fraser and the Yukon) generally experience higher flow velocities compared to river basins characterized by relatively mild slopes (such as the St. Lawrence and the Mackenzie).

5 Discussion and summary
The freshwater flux from rivers is an important forcing for the oceans, which influence the climate. This paper describes the implementation of the variable velocity flow routing scheme of AB99, which transports runoff from the land surface to the continental edges, in the CRCM. Basin discretizations and river flow directions at the polar stereographic resolution of the CRCM are based on the 5’ global river flow direction data of G99. G99 provide river basin discretization of 10 major river basins in North America and of these, 8 lie completely within the CRCM operational North American domain. Flow velocity is determined using Manning’s formula that relates velocity to discharge, slope, and river cross-section. River cross-sections are approximated as rectangular in shape, and the river width along the various sections of the digital river network is obtained from a geomorphological relationship between width and annual mean discharge. Estimates of annual mean discharge passing through the various sections of the river network are obtained with the help of the global runoff dataset of Cogley (1998). A mean characteristic value of Manning’s roughness coefficient for natural river channels is used, and the river slopes are determined from the mean elevations of the CRCM grid cells. Groundwater delay factors corresponding to the dominant soil type within a CRCM grid cell are used for coarse-, medium- and fine-textured soils. With all its parameter values determined from existing literature and based on model flow directions, the scheme does not require optimization or calibration of any parameters. The effect of the Great Lakes on the flow regime of the St. Lawrence River is parametrized in a simple manner. Flow directions are prescribed for Great Lakes cells such that flow continuity is maintained and the damping effect of the lakes is taken into account by assuming that each lake cell is a storage reservoir that acts to attenuate the flow.

In the absence of observation-based gridded estimates of runoff which could be used as input to the flow routing scheme, we use gridded runoff estimates from a global simulation of the calibrated VIC hydrological model forced with observation-based meteorological data. However, despite the prior calibration of the VIC model’s parameters to produce realistic runoff estimates for major river basins, some discrepancies remain. Compared to other observation-based runoff estimates, the VIC model runoff is larger for the Churchill and Yukon river basins and smaller for the Columbia. The primary reason for bias in annual runoff for these river basins is the problem associated with precipitation forcing (Nijssen et al., 2001b). We therefore do not analyse monthly streamflow for these rivers. In addition, we do not perform monthly streamflow analysis for the Nelson River since its streamflow is influenced by Lake Winnipeg which is not represented in the CRCM. For the remaining four rivers (the Fraser, Mackenzie, Mississippi and St. Lawrence), the simulated monthly streamflow compares reasonably well with observation-based estimates with an average $rrmse$ of around 35%. For these rivers, the inclusion of flow routing generally improves the agreement with observation-based estimates compared to the unrouted runoff. Simulated values of river flow velocity provide an additional measure to test the adequacy of the flow routing scheme. Although observation-based estimates of basin-wide averaged flow velocities are difficult to obtain, the flow velocities simulated in this study fall within the range of flow velocities obtained by AB99 for a global simulation. One drawback of the flow routing scheme used in this study is that it does not take into account the attenuation of flow due to the presence of dams and other man-made reservoirs. In the absence of detailed simulations with and without dams it is difficult to assess their effect on flow regimes. The errors associated with model precipitation and the manner in which this precipitation is
partitioned into evapotranspiration and runoff by the climate model land surface scheme, however, are likely to be larger than the errors associated with neglecting dams.

The primary objective of implementing a flow routing scheme in the CRCM is to close the hydrological cycle by transporting land-surface runoff to the continental edges and providing freshwater flux to the oceans. To this end, the results presented in this paper suggest that the flow routing scheme performs in a reasonable manner. Future combined atmospheric-hydrologic studies with the CRCM are likely to involve validation of CRCM simulated precipitation and water balance at river basin scales via comparisons with observed streamflow, and analysis of streamflow from enhanced greenhouse gas warming simulations to assess the impact of climate change on the hydrology of major North American river basins.

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Appendix
a The Flow Routing Algorithm
The water balance within a grid cell for the surface water stream channel storage, \( S \), is given by

\[
\frac{dS}{dt} = I - Q
\]

(7)

where \( I \) is the inflow into, and \( Q \) is the outflow from, a grid cell. The inflow \( I \) is given by

\[
I = f_s + f_n + f_g
\]

(8)

and consists of surface runoff generated within a grid cell \( f_s \), the inflow from neighbouring upstream cells \( f_n \), and the outflow from the groundwater store \( f_g \) within a grid cell which simulates the groundwater contribution to the river channel. The scheme uses Manning’s equation to determine time-evolving channel flow velocities, which depend on the amount of streamflow in the river channel:

\[
V = \frac{1}{n} R^{2/3} S^{1/2} = \frac{1}{n} \left( \frac{Wh}{W + 2h} \right)^{2/3} S^{1/2}
\]

(9)

where \( V \) is the channel velocity, \( n \) the Manning’s roughness coefficient, \( R \) the hydraulic radius (area of flow divided by the wetted perimeter), \( W \) and \( h \) are the river width and flow depth, respectively, and \( s \) the channel slope. The surface storage is assumed to be a linear function of outflow discharge \( Q \):

\[
S = \frac{L}{V} AV = LA = LWh
\]

(10)

where \( \tau \) is the travel time between the grid cell under consideration and its downstream neighbour, \( L \) is the distance between the grid cells, and \( A \) the cross-sectional area of the river. The outflow discharge is given by

\[
Q = AV = Wh\left( \frac{Wh}{n(W + 2h)} \right)^{2/3} s^{1/2}
\]

(11)

Substituting Eq. (10) and Eq. (11) into the continuity equation Eq. (7) yields,

\[
\frac{d(h)}{dt} = \frac{1}{LW} \left( I - \frac{W^{5/3} h^{5/3}}{n(W + 2h)^{2/3} s^{1/2}} \right)
\]

(12)

Equation (12) describes the flow in terms of the rate of change of flow depth for a given river section. An explicit forward step finite difference approximation of Eq. (12) is used to determine flow depth

\[
h_{t+1} = h_t + \frac{\Delta t}{LW} \left( I_t - \frac{W^{5/3} h_t^{5/3}}{n(W + 2h_t)^{2/3} s^{1/2}} \right)
\]

(13)

Once flow depth is known, the outflow discharge \( Q \) is estimated using Eq. (11). Since flow depth \( h \) in Eq. (13) depends on the inflow \( I \), the flow velocity \( V \), which depends on flow depth via Eq. (9), is simulated to be variable in time. In Manning’s equation, the flow velocity is directly proportional to the channel roughness coefficient \( n \). Manning’s roughness coefficient values for natural channels from 0.025 to 0.035 have been adopted in hydrological literature (Chow, 1964). Here we use a value of \( n \) equal to 0.03.

The groundwater model assumes that the storage \( G \) is a linear function of outflow \( f_g \) discharge:

\[
G = \tau_g f_g
\]

(14)

where \( \tau_g \) is the residence time in the groundwater reservoir and is related to the major soil type in a CRCM grid cell following Arora et al. (1999). The water balance of the groundwater reservoir \( G \) is determined by drainage input \( f_p \) and the outflow groundwater discharge \( f_g \):

\[
\frac{dG}{dt} = f_p - f_g
\]

(15)

Substitution of \( G \) in Eq. (15) yields

\[
\tau_g \frac{df_g}{dt} = f_p - f_g
\]

(16)

and, following AB99, we use the expression
\[
    f_{R(t+1)} = e^{-\Delta t/\tau_f} f_{R(t)} + \left(1 - e^{-\Delta t/\tau_f}\right) f_{p(t)} \tag{17}
\]
to determine discharge from the groundwater reservoir to the river stream channel and to step forward in time. In this paper, the time step \(\Delta t\) is equal to 1 day.

References


