A Semi-Implicit Semi-Lagrangian Regional Climate Model: The Canadian RCM

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ABSTRACT

A new regional climate model (RCM) is presented in this paper and its performance is investigated through a pair of 60-day simulations. This new model is based on the dynamical formulation of the Cooperative Centre for Research in Mesometeorology (CCRM) mesoscale nonhydrostatic community model and on the complete subgrid-scale physical parameterization package of the second-generation Canadian Centre for Climate modeling and analysis General Circulation Model (CCCma GCMII). The main feature of the Canadian RCM (CRCM) comes from the very efficient semi-implicit and semi-Lagrangian (SISL) numerical scheme used for the integration of the fully elastic nonhydrostatic Euler equations. The efficiency of the SISL scheme allows the use of longer time steps (at least by a factor of 5) for the integration of this model (e.g., the 45-km resolution version of the model uses a 15-min time step). A complete description of the numerical formulation of the model is presented with a review of the principal characteristics of the physical package. A pair of two-month-long winter simulations is also analyzed to investigate the behavior of the model and to evaluate the potential of the SISL integration scheme in the context of regional climate simulation. The two integrations, produced with a 45-km resolution version of the model, developed realistic small-scale details from the low-resolution GCMII fields used to initialize and drive the RCM.

1. Introduction

The pioneering work of Charney et al. (1950) and Phillips (1956) in numerical integration of meteorological equations set the standard for what is now known as global numerical weather prediction and general circulation models (GCMs). The numerical integration of the barotropic quasigeostrophic vorticity equation, made by Charney et al. (1950) over a modest domain of 15 \times 18 grid points for 24 time steps, took 24 h on their computer at Princeton. With the increased computing power available to him, Phillips (1956) was able to integrate a two-level quasigeostrophic baroclinic model on a 17 \times 16 grid point domain for 812 time steps. Phillips included surface friction (in the simplified form of an Ekman-layer dissipation) and he prescribed a net heating rate as a function of latitude (based on observations). These two forcings were included in the model to account for the dominant sources and sinks of energy in the atmosphere. His model was started from an initial state at rest and atmospheric motions are produced by the differential heating. This was the first general circulation simulation.

With ever-increasing computer power, refinements were gradually introduced to overcome known shortcomings of early GCMs. Over the past three decades, GCMs evolved from a very simple representation of the atmosphere toward sophisticated tools that reproduce many characteristics of the observed climate system. One of the major weaknesses that is still identified in modern GCMs, however, is the poor spatial resolution that limits their ability to properly simulate some physical processes such as the hydrological cycle (IPCC 1992). Because the computing time required to run a GCM increases approximately with at least the third power of the inverse of the horizontal grid length (fourth power if the concomitant vertical resolution is included), running a global model with a grid length of the order of 30 km would require roughly two days of computation per simulated day on a fast computer (Giorgi and Mearns 1991). Nevertheless, this kind of spatial resolution is required for a good representation of processes such as those related to the hydrological cycle (Giorgi and Mearns 1991). Since the computing power necessary for routinely modeling global climate at such resolution will not be available in the near future, alternative approaches to model finescale climate have to be explored.

One approach that has been pioneered at the National...
nions required by the regional model are supplied by initial and time-dependent lateral boundary conditions to a regional climate model (RCM). Because of its higher spatial resolution, the RCM then develops regional-scale details superimposed upon the driving large-scale flow. The main improvements in the climate simulated by an RCM appear in the fields related to the hydrological cycle. For example, much more realistic precipitating systems are simulated by the RCM than those from the driving GCM in studies over the western United States (Dickinson et al. 1989; Giorgi 1990; Giorgi and Bates 1989) and over Europe (Marinucci and Giorgi 1992). Following these studies using the NCAR model, other groups have become involved in the development of nested regional climate models: for example, the U.K. Meteorological Office (United Kingdom; Jones et al. 1995), the Commonwealth Scientific and Industrial Research Organisation (Australia; Walsh and McGregor 1995), the Danish Meteorological Institute (Denmark; Dethloff et al. 1996), the National Meteorological Center (now known as the National Centers for Environmental Prediction) (United States; Juang and Kanamitsu 1994), and the Max Planck Institute (Germany; Jacob and Podzun 1997).

The nesting approach has been adopted for the development of the Canadian RCM (CRCM). The CRCM takes advantage of a very efficient semi-Lagrangian and semi-implicit (SISL) marching scheme (Tanguay et al. 1990, hereafter TRL; Laprise et al. 1997). This scheme permits the use of time steps nearly 10 times longer than traditional (Eulerian explicit) schemes, thus increasing the computer efficiency of CRCM compared to others. This dynamical kernel is coupled to the subgrid-scale physical parameterization package of the Canadian Centre for Climate modelling and analysis (CCCma) second-generation GCM (GCMI; McFarlane et al. 1992).

The present paper, divided in two parts, describes this new RCM. The first part consists of a description of the CRCM where dynamics formulation, subgrid-scale physical parameterizations, nesting, and the coupling between dynamics and physics are covered. The second part presents the analysis of two 60-day-long winter simulations of the CRCM coupled to GCMI. This analysis compares CRCM simulation with its parent GCMI for two different CRCM configurations. This analysis also provides a validation of the use of long time steps for regional climate modeling.

2. Model description

Simulation of regional climate with nested models requires two models: a high-resolution nested LAM and a coarser-resolution global driving model. The initial and time-dependent lateral boundary conditions required by the regional model are supplied by the global model. Simulated data from the CCCma GCMII are used to drive our regional model. A description of GCMII and its simulated climate can be found in McFarlane et al. (1992). The nested model (CRCM) results from the coupling of the Cooperative Centre for Research in Mesometeorology mesoscale nonhydrostatic community model dynamics (Bergeron et al. 1994; Laprise et al. 1997) and the complete physical processes parameterization package of GCMII. The use of the same physical parameterization in CRCM and GCMII ensures consistency between the two models and facilitates the transfer of information from the driving to the nested model. Given the targeted grid length of 45 km for CRCM, no adaptation of this physics package was deemed required.

The CRCM dynamics evolved from the nonhydrostatic model originally developed by the late André Robert and his colleagues (TRL). This model solves the fully elastic Euler equations with an efficient semi-implicit semi-Lagrangian marching scheme. Variants of this model have been used to simulate dry convection (Robert 1993), moist convection (Pellerin 1992), supersonic flows (Larocque 1995), and for mesoscale studies (Benoit et al. 1997). The major upgrades from the TRL version are the inclusion of topography in the model by Robert and Denis (Denis 1990) through the use of Gal-Chen vertical coordinates (Gal-Chen and Sommerville 1975), the option of using variable vertical resolution, a modification of the semi-implicit scheme to minimize a problem related to the use of long time steps in the presence of small-scale orography (Tanguay et al. 1992; Hérel and Laprise 1996), and an extension of the semi-Lagrangian transport to the vertical by Robert (Bergeron et al. 1994). The following section summarizes the dynamical and numerical formulation of CRCM.

a. CRCM dynamical formulation

The CRCM is based on the fully elastic three-dimensional Euler equations used to describe the atmospheric flow on the rotating earth. Making use of the fully elastic Euler equations allows the CRCM to be operated at all meteorological (from planetary to micro) scales. The continuous increase in GCMs’ resolution imposes a similar increase in the resolution of RCMs to keep their ability to downscale GCMs simulations. One can expect RCMs to be routinely used at resolutions where nonhydrostatic effects would have to be correctly represented in model’s formulation within the next decade. There is no anticipated advantage or disadvantage of using a nonhydrostatic model at the resolution of the experiments presented in this paper. Nonhydrostatic effects are not expected at a resolution of 45 km. The benefit of using a nonhydrostatic framework is effective when the CRCM is used at higher resolution. The semi-implicit and semi-Lagrangian (SISL) scheme used in
CRCM permits the integration of the complete elastic Euler equations at a computing cost that is even lower than integrating primitive equations with Eulerian schemes (Tanguay et al. 1990). These equations, when expressed in a conformal projection, take the following form (Eq. 5 of TRL):

\[
\frac{dU}{dt} = fV - K \frac{\partial S}{\partial X} - RT \frac{\partial q}{\partial X} + F_u \tag{1}
\]

\[
\frac{dV}{dt} = -fU - K \frac{\partial S}{\partial Y} - RT \frac{\partial q}{\partial Y} + F_v \tag{2}
\]

\[
\frac{dw}{dt} = -g - RT \frac{\partial q}{\partial Z} + F_z \tag{3}
\]

\[
\frac{dq}{dt} = \left\{ -S \left( \frac{\partial U}{\partial X} + \frac{\partial V}{\partial Y} \right) - \frac{\partial w}{\partial Z} + \frac{L}{T} \right\} (1 - \kappa)^{-1} \tag{4}
\]

\[
\frac{dT}{dt} = \kappa T \frac{dq}{dt} + L \tag{5}
\]

\[
\frac{dM}{dt} = E \tag{6}
\]

In (1)–(6), the variables in the regional model have the following definitions:

- \( U, V, w, T, \) and \( M \) are the velocity components along the \( X, Y, \) and \( z \) coordinates, temperature and specific humidity, respectively;
- \( q = \ln(p/p_0) \) with \( p \) for pressure and \( p_0 \) a constant;
- \( f = 2\Omega \sin \varphi \) is the Coriolis parameter;
- \( K = (U^2 + V^2)/2 \) is the specific pseudokinetic energy;
- \( S = m^2 \) is the projection metric term, where \( m = (1 + \sin \varphi_o)/(1 + \sin \varphi) \) is the map scale factor for a polar stereographic projection, with \( \varphi \) the latitude and \( \varphi_o \) the reference latitude for the conformal transformation;
- \( L \) and \( E \) represent heat and moisture sources or sinks;
- \( F_u, F_v, \) and \( F_z \) correspond to sources or sinks of momentum for their respective wind components;
- \( g \) is the gravitational acceleration;
- \( R \) is the gas constant for air; and
- \( \kappa = R/C_p \) with \( C_p \) the heat capacity at constant pressure.

The total derivative in conformal projection is interpreted as follows:

\[
\frac{d}{dt} = \frac{\partial}{\partial t} + S \left( U \frac{\partial}{\partial X} + V \frac{\partial}{\partial Y} \right) + w \frac{\partial}{\partial Z}, \tag{7}
\]

where the \((X, Y)\) components of the polar stereographic coordinate are defined as the following functions of longitude \( \lambda \):

\[
DX = -m(dx \sin \lambda + dy \cos \lambda)
\]

\[
DY = m(dx \cos \lambda - dy \sin \lambda). \tag{8}
\]

The “image” wind components \((U, V)\) are defined in terms of the true horizontal wind components \((u, v)\) expressed in a local \((x, y, z)\) Cartesian reference system:

\[
U = -m^{-1}(u \sin \lambda + v \cos \lambda)
\]

\[
V = m^{-1}(u \cos \lambda - v \sin \lambda). \tag{9}
\]

It is noteworthy that with this notation \( U = S^{-1}(dX/dt) \) and \( V = S^{-1}(dY/dt) \).

1) Vertical coordinate

Equations (1)–(6) are modified by the introduction of a terrain-following vertical coordinate employed to facilitate the implementation of topography in the model. The vertical coordinate used in the CRCM is the Gal-Chen coordinate (Gal-Chen and Sommerville 1975) defined as:

\[
Z(X, Y, z) = \left( \frac{z - h_0(X, Y)}{H - h_0(X, Y)} \right) H, \tag{10}
\]

where \( Z \) is the scaled height coordinate, \( h_0(X, Y) \) is the topographic height, and \( H \) is the top of the model atmosphere. The \( Z \) coordinate is defined such that the surface \( Z = 0 \) corresponds to the ground \( z = h_0(X, Y) \) and the surface \( Z = H \) is at a constant height of \( z = H \). This coordinate has the advantage that the kinematic boundary condition at the surface, expressed by the following relation in \( z \)-coordinate form:

\[
w(X, Y, z = h_0, t) = \nabla (X, Y, z = h_0, t) \cdot \nabla h_0, \tag{11}
\]

takes the form of a homogenous boundary condition in the \( Z \) coordinate:

\[
W(X, Y, Z = 0, t) = \left( \frac{dZ}{dt} \right)_{(X,Y,Z=0)} = 0. \tag{12}
\]

When the geometric height coordinate \( z \) used in TRL is replaced with the Gal-Chen coordinate \( Z \), the chain rule of differentiation modifies partial derivatives as follows (e.g., Kasahara 1974):

\[
\frac{\partial \psi}{\partial c} \bigg|_{z} = \left( \frac{\partial \psi}{\partial c} \right)_Z + \left( \frac{\partial \psi}{\partial Z} \right)_Z \left( \frac{\partial Z}{\partial c} \right)_Z, \tag{13}
\]

\[
\frac{\partial \psi}{\partial Z} \bigg|_{c} = \left( \frac{\partial \psi}{\partial Z} \right)_Z + \left( \frac{\partial \psi}{\partial Z} \right)_Z \left( \frac{\partial Z}{\partial c} \right)_Z, \tag{14}
\]

where \( \psi \) is any dependent variable and \( c \) represents either the \( X \) or \( Y \) coordinate or time \( t \). A subscript beside a partial derivative is used to specify along which constant surface the partial derivative is taken. In the \((X, Y, Z)\) coordinate system, the Euler equations take the following form (Bergeron et al. 1994):
The total derivative must now be interpreted as follows:

\[
\frac{dU}{dt} = fV - K \frac{\partial S}{\partial X} - RT \left( \frac{\partial q}{\partial X} + \frac{G_v \partial q}{G_o \partial Z} \right) + F_x
\]

(15) \hspace{1cm} \frac{d}{dt} = \frac{\partial}{\partial t} + S \left[ U \left( \frac{\partial}{\partial X} + \frac{V}{\partial Y} \right) \right] + W \frac{\partial}{\partial Z}.

(27)

The generalized vertical velocity in this coordinate system is to be interpreted as

\[
\frac{dV}{dt} = -fU - K \frac{\partial S}{\partial Y} - RT \left( \frac{\partial q}{\partial Y} + \frac{G_v \partial q}{G_o \partial Z} \right) + F_y
\]

(16) \hspace{1cm} W = \frac{dZ}{dt} \left[ \frac{\partial Z}{\partial t} + S \left( \frac{\partial Z}{\partial X} + \frac{V}{\partial Y} \right) \right] + \frac{w}{\partial Z}.

(28)

\[
\frac{dW}{dt} = -g - RT \frac{\partial q}{G_o \partial Z} + F_Z
\]

(17) \hspace{1cm} W = \frac{dZ}{dt} \left[ \frac{\partial Z}{\partial t} + S \left( \frac{\partial Z}{\partial X} + \frac{V}{\partial Y} \right) \right] + \frac{w}{\partial Z}.

(28)

\[
(1 - \kappa) \frac{dq}{dt} = S(F_i U + F_z V) - S \left( \frac{\partial U}{\partial X} + \frac{\partial V}{\partial Y} \right)
\]

\[
- \frac{1}{G_o} \left( \frac{\partial G_o W}{\partial Z} \right) + \frac{L}{T}
\]

(18)

where

\[
F_i = \frac{1}{G_i} \left( \frac{\partial h_0}{\partial X} \right)
\]

(22)

\[
F_z = \frac{1}{G_z} \left( \frac{\partial h_0}{\partial Y} \right)
\]

(23)

\[
G_o = \frac{(H - h_0)}{H}
\]

(24)

\[
G_i = - \frac{(H - Z)}{H} \left( \frac{\partial h_0}{\partial X} \right)
\]

(25)

and

\[
G_z = - \frac{(H - Z)}{H} \left( \frac{\partial h_0}{\partial Y} \right)
\]

(26)

The total derivative must now be interpreted as follows:

\[
\frac{dU}{dt} = fV - K \frac{\partial S}{\partial X} - RT \frac{G_v \partial q}{G_o \partial Z} + F_x
\]

(33)

\[
\frac{dV}{dt} = -fU - K \frac{\partial S}{\partial Y} - RT \frac{G_v \partial q}{G_o \partial Z} + F_y
\]

(34)

\[
\frac{dW}{dt} = -g - RT \frac{\partial q}{G_o \partial Z} + F_Z
\]

(35)

\[
(1 - \kappa) \left( \frac{dq}{dt} - \frac{gw}{RT_s} \right) + S \left( \frac{U}{\partial X} + \frac{V}{\partial Y} \right) + \frac{\partial (G_o W)}{\partial Z} = S(F_i U + F_z V) - \frac{(1 - G_o) \partial (G_o W)}{\partial Z} + \frac{L}{T}
\]

(36)
The semi-Lagrangian integration procedure consists in evaluating total derivatives as follows:

$$\frac{d\psi}{dt} \equiv \frac{D\psi}{Dt} = \frac{\psi(X, Y, Z, t + \Delta t) - \psi(X - 2\alpha, Y - 2\beta, Z - 2\gamma, t - \Delta t)}{2\Delta t}.$$  (40)

Upstream values are obtained by cubic interpolation of surrounding gridpoint values. The Lagrangian displacements ($\delta, \beta, \gamma$) are calculated by solving iteratively the following implicit coupled equations system:

$$\alpha(X, Y, Z, t) = \Delta t S U (X - \alpha, Y - \beta, Z - \gamma, t)$$  (41)

$$\beta(X, Y, Z, t) = \Delta t S V (X - \alpha, Y - \beta, Z - \gamma, t)$$  (42)

$$\gamma(X, Y, Z, t) = \Delta t W (X - \alpha, Y - \beta, Z - \gamma, t).$$  (43)

Robert et al. (1985) presents the formalism of the solution of (41)–(43).

The strategy of semi-implicit semi-Lagrangian scheme consists in expressing linear left-hand side terms in (33)–(39) as averages between fields on grid points at forecast time ($\psi_{X, Y, Z, t + \Delta t}$) and their upstream values at past time ($\psi_{X - 2\alpha, Y - 2\beta, Z - 2\gamma, t - \Delta t}$). A partly uncentered time average of the following form is used in CRCM:

$$\overline{\psi} = \frac{(1 + \varepsilon)\psi(X, Y, Z, t + \Delta t) + (1 - \varepsilon)\psi(X - 2\alpha, Y - 2\beta, Z - 2\gamma, t - \Delta t)}{2}.$$  (44)

where $\varepsilon$ represents the degree to which the average is uncentered. This approach is used to reduce problems of instability of topographic waves associated with rapid flows when using long time steps (Tanguay et al. 1992).

The right-hand side terms of (33)–(39) are evaluated based on current time values $\psi(t)$ but as uncentered spatial averages of their values at the end points of the Lagrangian trajectory (Tanguay et al. 1992) as follows:

$$\overline{\overline{\psi}} = \frac{(1 + \varepsilon)\psi(X, Y, Z, t) + (1 - \varepsilon)\psi(X - 2\alpha, Y - 2\beta, Z - 2\gamma, t)}{2}.$$  (45)

The application of the semi-implicit semi-Lagrangian scheme modifies the Euler equations as follows:

$$\frac{DU}{Dt} + RT^* \frac{\partial q'}{\partial X} = \left( f V - K \frac{\partial S}{\partial X} - RT^* \frac{\partial q'}{\partial X} - RT \frac{G_0}{G_0} \frac{\partial q'}{\partial Z} \right)$$  (46)

$$\frac{DV}{Dt} + RT^* \frac{\partial q'}{\partial Y} = \left( -f U - K \frac{\partial S}{\partial Y} - RT^* \frac{\partial q'}{\partial Y} - RT \frac{G_0}{G_0} \frac{\partial q'}{\partial Z} \right)$$  (47)

$$\frac{Dw}{Dt} + RT^* \frac{\partial q'}{\partial Z} - \frac{T^*}{T^*} = \left( RT^* \frac{\partial q'}{\partial Z} - RT \frac{G_0}{G_0} \frac{\partial q'}{\partial Z} \right)$$  (48)

$$\left(1 - \kappa \right) \left[ \frac{Dq'}{Dt} - \frac{gw}{RT^*} \right] + S \left[ \frac{\partial U}{\partial X} + \frac{\partial V}{\partial Y} \right] + \frac{\partial G_0 W}{\partial Z} = \left( S (F, U + F, V) - \frac{(1 - G_0)}{G_0} \frac{\partial G_0 W}{\partial Z} \right)$$  (49)
The terms \( P_i \) can be written in symbolic form as follows:

\[
\frac{D T'}{D t} - \kappa T^R \frac{\partial q'}{\partial X} + \frac{\kappa g}{R} \frac{\partial \overline{w}'}{\partial Y} = \left( \frac{k T}{1 - \kappa} \right) \left( S(F_i U + F_3 V) - \left[ \frac{\partial U}{\partial X} + \frac{\partial V}{\partial Y} \right] - \frac{1}{G_o} \frac{\partial (G_o W)}{\partial Z} \right) \]

(50)

\[
\frac{D M}{D t} = 0 \]

(51)

\[
\overline{w}' - G_o \overline{W}' = -S(G_o U + G_2 V) \]

(52)

Note that physical forcing terms \( F_x, F_y, F_z, L, \) and \( E \) evaluated in the physics module have been omitted in (46)–(52). Their contributions are added as a corrective step via time splitting to get the final values of the dependent variables only after completion of the inviscid adiabatic time step.

3) Temporal discretization

With the definition of total derivatives (40), time average (44), and spatial average between Lagrangian trajectory end points (45) and by grouping terms on the basis of their time level, the set of equations (46)–(52) can be written in symbolic form as follows:

\[
Q_i(X, Y, Z, t + \Delta t) = P_i(X - 2\alpha, Y - 2\beta, Z - 2\gamma, t - \Delta t)
\]

\[
+ 2\Delta t \overline{R}_i(X - \alpha, Y - \beta, Z - \gamma, t). \]

(53)

The terms \( Q_i \), valid at time \( t + \Delta t \), are given by

\[
Q_u = U + (1 + \varepsilon)\Delta t RT^* \frac{\partial q}{\partial X}
\]

(54)

\[
Q_v = V + (1 + \varepsilon)\Delta t RT^* \frac{\partial q}{\partial Y}
\]

(55)

\[
Q_w = w + (1 + \varepsilon)\Delta t RT^* \frac{\partial q}{\partial Z} - (1 + \varepsilon)\Delta t \frac{\partial (G_o W)}{\partial Z} \]

(56)

\[
Q_q = (1 - \kappa) \left( q' - (1 + \varepsilon)\Delta t \frac{g w}{RT^*} \right)
\]

\[
+ (1 + \varepsilon)\Delta t \left[ \frac{\partial U}{\partial X} + \frac{\partial V}{\partial Y} \right]
\]

\[
+ (1 + \varepsilon)\Delta t \frac{\partial (G_o W)}{\partial Z}
\]

(57)

\[
Q_T = T' - \kappa T^* q' - (1 + \varepsilon)\Delta t \frac{\kappa g}{R} w
\]

(58)

\[
Q_M = M
\]

(59)

\[
Q_w = -(1 + \varepsilon)\Delta t (G_o W - w).
\]

(60)

The terms \( P_i \), calculated from variables defined at time \( t - \Delta t \), are given by

\[
P_u = U - (1 - \varepsilon)\Delta t RT^* \frac{\partial q'}{\partial X}
\]

(61)

\[
P_v = V - (1 - \varepsilon)\Delta t RT^* \frac{\partial q'}{\partial Y}
\]

(62)

\[
P_w = w - (1 - \varepsilon)\Delta t RT^* \frac{\partial q'}{\partial Z} - (1 - \varepsilon)\Delta t \frac{\partial (G_o W)}{\partial Z}
\]

(63)

\[
P_q = (1 - \kappa) \left( q' - (1 - \varepsilon)\Delta t \frac{g w}{RT^*} \right)
\]

\[
- (1 - \varepsilon)\Delta t \left[ \frac{\partial U}{\partial X} + \frac{\partial V}{\partial Y} \right]
\]

\[
- (1 - \varepsilon)\Delta t \frac{\partial (G_o W)}{\partial Z}
\]

(64)

\[
P_T = T' - \kappa T^* q' - (1 - \varepsilon)\Delta t \frac{\kappa g}{R} w
\]

(65)

\[
P_M = M
\]

(66)

\[
P_w = -(1 - \varepsilon)\Delta t (w - G_o W).
\]

(67)

The terms \( R_i \), calculated from variables defined at time \( t \), are given by

\[
R_u = f V - K_o \frac{\partial S}{\partial X} - RT^* \frac{\partial q'}{\partial X} - RT \frac{G_i q'}{G_o} \frac{\partial q'}{\partial Z}
\]

(68)

\[
R_v = -f U - K_o \frac{\partial S}{\partial Y} - RT^* \frac{\partial q'}{\partial Y} - RT \frac{G_i q'}{G_o} \frac{\partial q'}{\partial Z}
\]

(69)

\[
R_w = RT^* \frac{\partial q'}{\partial Z} - RT \frac{\partial q'}{G_o} \frac{\partial q'}{\partial Z}
\]

(70)

\[
R_q = S(F_i U + F_3 V) \frac{G_o}{G_o} - \frac{1 - G_o}{G_o} \frac{\partial (G_o W)}{\partial Z}
\]

(71)

\[
R_T = \frac{\kappa T'}{(1 - \kappa)} \left( S(F_i U + F_3 V) - \left[ \frac{\partial U}{\partial X} + \frac{\partial V}{\partial Y} \right] \right)
\]

\[
- \frac{1}{G_o} \frac{\partial (G_o W)}{\partial Z}
\]

(72)

\[
R_M = 0
\]

(73)

\[
R_w = -S(G_i U + G_i V).
\]

(74)
Once the explicit terms $P_t$ and $R_t$ have been computed and interpolated along trajectories, the set of coupled differential equations $Q$, for prognostic variables at time $t + \Delta t$ can be solved. This results in a three-dimensional Helmholtz equation for $q'$ at time $t + \Delta t$ (for details see Bergeron et al. 1994):

$$C_1[(1 - \kappa) - (1 + \varepsilon)^2(\Delta t)^2RT^*5\nabla^2]q' - (1 + \varepsilon)^2(\Delta t)^2RT^*D_i[D_i(q')] = A_2,$$

where

$$A_2 = C_1 \left[ Q_w \nabla \left( \frac{\partial Q_v}{\partial X} + \frac{\partial Q_v}{\partial Y} \right) + Q_w \left( \frac{1 - \kappa}{RT^*} \right) \right]$$

and

$$A_1 = Q_w \left( 1 + \varepsilon \right) \Delta t \frac{g}{RT^*} Q_T - \frac{Q_w}{\left( 1 + \varepsilon \right) \Delta t} C_1$$

$$C_1 = 1 + \left( 1 + \varepsilon \right)^2(\Delta t)^2\frac{\kappa^2 g^2}{RT^*}$$

$$D_i(\psi) = \frac{\partial \psi}{\partial Z} - \frac{\kappa g}{RT^*} \psi$$

$$D_z(\psi) = \frac{\partial \psi}{\partial X^2} + \frac{\partial \psi}{\partial Y^2}$$

Note that the $Q$, terms in Eqs. (76)–(81) are evaluated as the rhs of Eq. (53) and therefore they are obtained from current and past time information. The Helmholtz equation (75) is solved by a variant of the alternating-direction implicit method of Peaceman and Rachford (1955), using the lateral boundary conditions provided through the nesting scheme and the upper and lower boundary conditions $W = 0$ to obtain $q'$ at time $t + \Delta t$. The other variables at $t + \Delta t$ are then calculated via back-substitution with the following relations:

$$W = \frac{1}{C_1G_0} \left[ A_1 - (1 + \varepsilon)\Delta t RT^*D_i(q') \right]$$

$$w = \frac{Q_w}{\left( 1 + \varepsilon \right) \Delta t} + G_0 W$$

$$T' = Q_T + \kappa T^* q' - (1 + \varepsilon)\Delta t\frac{\kappa g}{R} w$$

$$U = Q_v - (1 + \varepsilon)\Delta t RT^* \frac{\partial q'}{\partial X}$$

$$V = Q_v - (1 + \varepsilon)\Delta t RT^* \frac{\partial q'}{\partial Y}.$$

4) **Spatial discretization**

Each term of Eqs. (75)–(86) has to be expressed in a spatially discrete form in order to be integrated numerically on a grid. This section presents the model grid and how the variables are distributed on the grid.

The model uses a staggered Arakawa C-grid (Arakawa and Lamb 1977) in the horizontal as shown in Fig. 1. All horizontal derivatives are evaluated as second-order centered finite differences over one grid length. Hence the application of a spatial derivative yields a result that is staggered with the original field. Cubic interpolation is used to position fields or their derivatives on the appropriate points when needed.

The model also uses a staggered arrangement of variables in the vertical as shown in Fig. 2. Levels with integer $k$ indices are called momentum levels, whereas half-integer levels are called thermodynamic levels. The $U$, $V$, and $q'$ fields are computed on the momentum levels, whereas all other fields are computed on thermodynamic levels. The last level, $k = N_{k+1/2}$, is used solely to apply the upper boundary conditions $w = 0$ and $W = 0$ at $z = Z = H$, no calculation being made at this level. Linear vertical interpolation is used to transfer fields from one type of level to the other when necessary. Details for the calculations on the lowest and highest levels are given in Bergeron et al. (1994).

b. **Physical formulation**

The contributions of physical processes, terms $F_X$, $F_Y$, $F_Z$, $L$, and $E$ in Eqs. (33)–(39), are computed with a parameterization package imported from GCMII (McFarlane et al. 1992). The highlights of this physical package are presented below.

1) **The $F_X$, $F_Y$, and $F_Z$ terms**

Terms $F_X$, $F_Y$, and $F_Z$ are introduced in the equations to represent the effect of the subgrid-scale processes affecting the momentum variables. These processes are the vertical turbulent fluxes of momentum, the gravity wave drag, and the horizontal diffusion. Vertical turbulent fluxes of horizontal momentum are evaluated as a stability-dependent diffusion using an eddy diffusivity formulation in the free atmosphere with a drag coefficient formulation at the surface. The effect of unresolved orographic gravity waves is parameterized following McFarlane (1987) as an additional drag force on the flow. This parameterization is based on a simple linear analysis of a stationary bidimensional flow. A critical Froude number is used to determine regions where wave dissipation and concomitant momentum deposition take place. The horizontal diffusion is treated separately in section 2c below.

2) **The $L$ term**

The $L$ term represents the combined diabatic heating resulting from the effects of radiation (solar and terrestrial), the release (absorption) of latent heat, and the turbulent and convective vertical heat flux.
Solar radiation is treated with an updated version of the Fouquart and Bonnel (1980) scheme and terrestrial radiation follows the method of Morcrette (1984). Because terrestrial and solar heating calculations are computationally very expensive, they are fully calculated at intervals of 6 h for the terrestrial radiation and of 3 h for the solar radiation. Between full calculations, corrections are made to account for the variation of local solar zenith angle (for solar radiation and cloud optical properties) and for the variation of temperature and cloudiness for terrestrial radiation. Surface albedo is defined for two spectral bands and it is a function of surface type, humidity, snow cover, and snow age. Clouds are determined diagnostically and their optical properties are calculated from their liquid water content [see section 2b(6)].

Latent heat released by condensation (freezing) of water vapor in a grid box is also included in the \( L \) term. Whenever the local relative humidity exceeds a critical value, the excess water vapor is precipitated and the associated latent heat is used to locally warm the air. Convective adjustment is applied to unstable layers to restore local lapse rates to a critical value. The amount of energy that is needed to restore the lapse rate to its critical value corresponds to a convective heat flux. The procedure is repeated iteratively until a stable temperature profile is reached. Vertical turbulent flux of heat follows eddy diffusivity and surface exchange formulations that are similar to momentum ones.

3) The \( E \) Term

Again, the vertical turbulent fluxes of moisture follow the formulation used for the heat. The water in the atmosphere is controlled in one part by precipitation (sink) as described in the previous section and by evaporation (source) from the surface (see section 5 for the land surface scheme description). The convective flux of water vapor associated with the convective adjustment between two layers is the minimum required to saturate the upper layer or to remove supersaturation from the bottom layer. If the bottom layer is still supersaturated...
after the upper layer has been saturated, the residual bottom-layer supersaturation is removed by condensation and the latent heat is used to warm the air.

4) OCEANS

The physical package of GCMII includes a mixed-layer ocean model and a thermodynamic sea-ice model. For the experiments described in this paper, however, both of these models were turned off and climatological values of sea surface temperature (SST) and ice coverage were used.

5) LAND SURFACE SCHEME

The land surface scheme consists of four prognostic budget equations for a single soil layer: one equation for heat, one for frozen soil moisture, one for liquid soil moisture, and one for snow amount. The ground temperature is calculated using the force-restore method where a deep soil temperature is defined as the running time-averaged surface temperature over the last 24 h. The surface energy budget equation takes into account heat storage in the soil, net solar flux absorbed by the ground, net terrestrial flux to the ground, terrestrial flux emitted by the ground, sensible and latent heat fluxes, and latent heat associated with the melting of snow and frozen soil moisture or freezing of liquid soil moisture. Complete melting (freezing) is assumed to take place whenever the soil temperature tries to go above (below) the 0°C value. The soil moisture regime is represented in terms of a simple “bucket” method with a field capacity that is a function of vegetation and soil characteristics. When the total soil moisture exceeds a critical value, runoff is assumed to occur instantaneously, returning the water surplus to the ocean. A more detailed description of the land surface scheme can be found in McFarlane et al. (1992).

Vegetation and soil characteristics are specified using the 1° lat × 1° long resolution land surface dataset of Wilson and Henderson-Sellers (1985). The evaporation from the surface varies with the soil moisture content, snow amount on the ground, and the vegetation and soil type. Similarly, surface albedo is a function of snow on the ground, underlying surface, and soil moisture content.

6) CLOUD COVERAGE

An interactive cloud cover is evaluated from the prognostic water vapor and temperature fields. The cloud coverage is evaluated diagnostically for each model layer from the relative humidity field. The vertically integrated cloud amount is dependent on the vertical distribution of clouds in the model layers. Clouds that are present in adjacent layers of the model are assumed to be totally overlapped, whereas clouds layers separated by a clear layer are assumed to be randomly overlapped.

Cloud optical properties are based on cloud liquid water content, which itself is evaluated diagnostically as a function of temperature. The functional dependence has been established to fit with observed cloudiness data (McFarlane et al. 1992).

c. Horizontal diffusion

Once an adiabatic time step of the dynamics has been completed, followed by correction to incorporate physical forcings, the horizontal diffusion is applied in third place in the time splitting procedure adopted in the CRCM. The horizontal diffusion is applied in the following general form:

\[
\psi_H = \psi + 2\Delta t k \nabla^2 \psi, \quad (87)
\]

where \( k \) and \( m \) are user’s adjustable parameters and the subscript \( H \) refers to a variable after diffusion.

d. Nesting

A one-way nesting procedure is used to drive the CRCM with GCMII data; that is, CRCM gets information from GCMII but it does not influence GCMII. Initial and time-dependent lateral boundary conditions (wind components, temperature, geopotential, and water vapor) are extracted from GCMII archived data. The nesting method used is inspired by Davies (1976) and refined by Robert and Yakimiw (1986) and Yakimiw and Robert (1990). Nesting is carried out after the variables have been updated by dynamical and physical processes and horizontal diffusion. Over a specified distance from the lateral boundaries, the variables from the integration of the regional model are gradually blended with the those of the driving (global) model. By the time the CRCM lateral boundaries are reached, the values of the driving variable are imposed. Throughout the rest of the grid (free zone) the variables of the regional model are not affected by those of the global model. A variable resulting from the dynamical and physical processes and from horizontal diffusion, \( \psi_H \), known at time \((t + \Delta t)\) and located within the sponge zone, is then modified as follows:

\[
\psi_H(X, Y, Z) = \psi_H(X, Y, Z) + P(X, Y)(\psi_C(X, Y, Z) - \psi_H(X, Y, Z)),
\]

(88)

where \( \psi_C \) is the resultant variable after the nesting step (nested variable), \( \psi_H \) is the variable from the regional model, \( \psi_C \) is the variable from the driving model (global), and \( P \) is the attenuation function, which varies from a value of 0 in the free zone to a value of 1 at the lateral boundary. The lateral boundary conditions \( \psi_C \) are updated at every time step with linear interpolation in time of the GCMII fields that are archived at 12-h intervals.
350 Volume 127

3. Simulations

To assess the potential of CRCM to downscale GCMII simulations, a qualitative evaluation of regional-scale features generated during modest length integrations of CRCM is presented. Two short winter simulations were conducted to investigate the ability of the CRCM to develop realistic regional-scale features in its simulated meteorological fields. The presence of a signature of these regional-scale features in monthly statistics of CRCM fields is also investigated. For these experiments CRCM is driven by GCMII outputs, therefore allowing the driving and driven model to share the same physical parameterization. The main objective here is to assess the behavior of this new model and to validate, within the context of regional climate simulation, the use of long time steps permitted by the SISL scheme. It should be emphasized that an evaluation of the CRCM climate is not intended in this analysis; such an evaluation can be found in Laprise et al. (1998).

3. Simulations

(a) Experiments design

This subsection describes the steps involved in the preparation, production, and analysis of CRCM-simulated fields and comparable fields from GCMII. A pair of two-month-long winter simulations were conducted to test this new RCM, covering the period beginning December first and extending to the 31st of the following January. The simulations were carried out on a polar stereographic grid of $121 \times 121$ points at a 45-km resolution, resulting in a square domain of $5400 \text{ km}^2$, centered over the state of Vermont. As can be seen in Fig.

---

Fig. 3. Ground cover and computational grid for CRCM. The complete $121 \times 121$ grid is presented; the inner white square delineates the $101 \times 101$ free grid. Ground cover types are land (dark gray), open water (light gray), and sea ice (white).
Fig. 4. Land–sea-ice mask for January and topographic height for CRCM (a) and GCMII (b). Ground cover types are as in Fig. 3. Topographic height differences between CRCM and GCMII are presented in (c). Contour interval is 100 m; negative values appear as broken lines.

Fig. 5. Distribution of staggered vertical levels in CRCM-10 (a) and CRCM-30 (b).

3, the geographical region covered by the grid encompasses the eastern part of North America and part of the Atlantic Ocean. The visible portion of the inner white square in Fig. 3 delineates the lateral sponge zone used to drive the model. The CRCM is designed to allow a variable width of the nesting zone for each of the driven variables. For the simulations presented in this paper, only the horizontal wind components were forced over the indicated nine-gridpoint belt surrounding the free domain, with temperature, pressure, and specific humidity being specified only at the outmost boundary. The ground cover field for the month of January is also presented in Fig. 3. It has to be noted that the land–sea mask in CRCM was interpolated from a 1° lat × 1° long reference dataset, which corresponds approximately to a resolution of 110 km. This explains the apparent mismatch in Fig. 3 between the geographical borders and the ground cover definition. For comparison, Fig. 4 presents the ground cover and topography fields of CRCM and GCMII as well as the height differences. The major improvement in the regional land–sea mask appears in the definition of James Bay, Great Lakes, and Florida...
peninsula. A more realistic representation of the topography is also permitted with the use of higher resolution.

Figure 5 presents the vertical distribution of levels used for the 10- and 30-level simulations. Both simulations use a vertical domain of approximately 35 km ($H = 34.120$ km in CRCM-10; $H = 31.953$ km in CRCM-30) similar to the driving model vertical extension. In the 10-level version, levels were selected to approximate the vertical location of GCMII levels. An exact replica of the GCMII vertical discretization is however impossible because of the different vertical coordinates used by the two models. The 10- and 30-level models will be referred to as CRCM-10 and CRCM-30, respectively. A 15-min time step is used in both simulations and the value of parameter $\varepsilon$ used in Eq. (44) is 0.1. The horizontal diffusion is harmonic ($m = 1$ in Eq. (87)) with a diffusion coefficient $k$ set to $10^3$ m$^2$ s$^{-1}$.

Simulation of the atmospheric circulation with a re-
Fig. 8. January mean of the 500-hPa geopotential height for CRCM-10 (a), GCMII (b), and differences between CRCM-10 and GCMII (c). Contour interval is 4 dam for (a) and (b) and 1 dam for (c).

Fig. 9. January mean of the surface temperature for CRCM-10 (a), GCMII (b), and differences between CRCM-10 and GCMII (c). Contour interval is 5°C for (a) and (b) and 1°C for (c) with negative values in broken line.
Regional model requires many sources of information. The information required to integrate CRCM may be classified in four categories: initial values for both atmospheric and surface prognostic variables, time-dependent lateral boundary conditions for nesting atmospheric fields, and lower boundary conditions for prescribed surface fields.

Prescribed surface fields include time series of sea-ice coverage and SST taken from a $10^\circ \times 10^\circ$ climatological database compiled at CCCma. The same database was used for GCMII simulations. Initial values of prognostic land surface variables of CRCM (ground temperature, snow amount, and liquid and frozen soil water content) came from a $1^\circ \times 1^\circ$ climatological database.

The initial and lateral boundary atmospheric data were extracted from an archive of GCMII simulation corresponding to current climate conditions (McFarlane et al. 1992). This GCMII simulation was produced with a T32 version of this spectral GCM with 10 levels using a 20-min time step. We selected the 12th (December) and 13th (January) months of a simulation that began on 1 January. GCMII simulated data were archived at 12-h intervals (0000 and 1200 UTC) in spectral formats and in GCMII terrain-following vertical coordinates. Interpolation in time and in space was performed to supply initial atmospheric information as well as subsequent nesting data at each time step and for grid points in the CRCM nesting zone. Interpolation in time is linear and a discussion of the spatial interpolation follows.

The first step to be carried out is the transformation of GCMII spectral divergence and vorticity to spectral winds, followed by a transformation of the different fields from spectral coefficients to gridpoint values. Then a vertical interpolation from the GCMII $\eta$ coordinate (Laprise and Girard 1990) to pressure levels is made. The temperature field is extracted from geopotential heights using the hydrostatic relation. Horizontal interpolation is then performed to transfer the GCMII atmospheric fields from a global Gaussian grid to a CRCM regional polar-stereographic grid. This is followed by a further vertical interpolation from pressure levels to CRCM Gal-Chen levels and by staggering the variables in their appropriate location on the computation grid. It is important to mention that using a RCM with geometric-type instead of pressure-type height coordinate does not involve more vertical interpolation in shifting information from the driving to the driven model. Actually, even if regional and global models shared the same vertical coordinate system, the levels would be differently located because of variations in topography ensuing from different horizontal resolutions.

To analyze the CRCM simulation in the following subsection, CRCM-simulated fields will be unstaggered and vertically interpolated from Gal-Chen to pressure levels for ease of comparison with corresponding GCMII fields. The horizontal interpolation of some surface fields is not straightforward since many of them

![Figure 10](image1.png)  
**Fig. 10.** January mean of the 1000-hPa temperature for CRCM-10 (a), GCMII (b), and differences between CRCM-10 and GCMII (c). Contour interval is 2°C for (a) and (b) and 1°C for (c) with negative values indicated by broken line.
(land–sea mask and vegetation type being obvious examples) cannot be interpolated and have to be defined in a coherent way with surface fields extracted from the CCCma climatological database. For fields that cannot be interpolated, the value of the nearest grid point is assigned to the CRCM grid point. For all other fields, a cubic interpolation is used.

b. Analysis

To assess the downscaling ability of the high-resolution CRCM driven by GCMII low-resolution data, we will first analyze some fields taken at one time in the course of the CRCM integration. Figure 6 compares the 700-hPa specific humidity and mean sea level pressure (MSLP) fields for 14 January 0000 UTC in CRCM and GCMII simulations. The water vapor field exhibits much more detail than GCMII, displaying the classical comma shape that is generally observed in mature weather systems. The CRCM MSLP distribution shows a well-defined trough extending from Newfoundland to Florida contrasting with the coarse fields of GCMII. Differences in the CRCM simulation are not as pronounced in all fields. In general, hydrological fields (water vapor,
clouds, and precipitation) are most sensitive to the increased resolution of CRCM.

A central question related to the use of a high-resolution model for climate modeling is whether there remains any effect of these simulated small-scale details in monthly statistics. To evaluate the ability of CRCM to develop regional characteristics in a climatological sense, some basic statistics were computed on the second month (January) data generated by the 2-month runs. The computed statistics are a spatial distribution of monthly means and transient-eddy standard deviations of some model variables. It is clear that one month is far too short to infer any climatology from a model but, at least, it will suffice to see if any regional-scale features are still present on a monthly mean basis. A complete evaluation of CRCM climate and its sensitivity to increased greenhouse gas concentration can be found in Laprise et al. (1998).

Figure 7 shows January mean MSLP field for CRCM and GCMII. Both models present a similar pattern with an elongated subtropical high pressure zone around 30°N with a gradual decrease northward. Figure 7c shows that the averaged MSLP is generally higher in CRCM than in GCMII. This bias in the MSLP is not constant in time; it increases rapidly in the first 10 days of the simulation and then decreases slowly for the rest of the integration (Caya 1996). This behavior seems to be related to the nonhydrostatic framework used for CRCM and the fact that MSLP is not one of the nested fields in the model. This bias in CRCM has been corrected in Laprise et al. (1998) by nesting the domain-averaged MSLP field. This mass bias has repercussions in CRCM geopotential heights as illustrated by Fig. 8 where the 500-hPa height field is shown. A domain-averaged difference of about 20 m is present for January mean values. Since geopotential heights are diagnostically calculated by integration of the hydrostatic equation from the surface upward, their bias is therefore directly related to the MSLP bias.

The increased resolution of CRCM permits a better representation of the ground cover (as was seen in Fig. 4), which creates regional-scale forcings on the atmospheric circulation. As an example, the presence of warm water bodies, such as the sharply defined Great Lakes in the CRCM ground temperature field (Fig. 9a), generates a warmer area in the January averaged CRCM 1000-hPa temperature field (Fig. 10a). Despite the use of the same data for SSTs, the differences of ±1°C in ground temperature that appear over the ocean (Fig. 9c) are solely caused by the double interpolation needed to bring SSTs from the global lat–long grid to the global Gaussian grid of GCMII and then to the CRCM grid used to present results from both models. Also apparent in the CRCM 1000-hPa temperature field is the strong gradient over the east coast of Labrador caused by the quasi-discontinuity in ground temperature between sea-ice and open water, with a variation of 15°C over a single 45-km grid point (Fig. 9a). The same temperature difference is observed in GCMII simulation but its 450-km resolution reduces the gradient by an order of magnitude. In Fig. 10c, differences of 2° and 3°C are seen over the southern Appalachian ridge and Labrador Plateau, respectively. These differences between CRCM and GCMII result from the extrapolation of temperature below the lowest model level when the 1000-hPa level is beneath the surface. It has to be noted that this extrapolation is sensitive to the topographic heights of each model. However, since the extrapolation is made diagnostically from model data, simulations are not affected by it.

Vertical profiles of domain-averaged January mean temperature are presented in Fig. 11. Despite the fact that both models have the same number of levels and are located at approximately the same place, the tro-
The 1 January 0000 UTC RH for GCMII at 900 hPa (a), 800 hPa (b), 700 hPa (c), and vertical profile (d) of RH over the state of New York [indicated by a cross in (a), (b), and (c)]. Contour intervals are 40%, 50%, 60%, 70%, 75%, 80%, 85%, 90%, 95%, and 100% RH.

**Figure 13.** The 1 January 0000 UTC RH for GCMII at 900 hPa (a), 800 hPa (b), 700 hPa (c), and vertical profile (d) of RH over the state of New York [indicated by a cross in (a), (b), and (c)]. Contour intervals are 40%, 50%, 60%, 70%, 75%, 80%, 85%, 90%, 95%, and 100% RH.

The tropopause seems to be better defined in GCMII. It appears that the smoother CRCM profile is caused by the double vertical interpolation needed to bring the GCMII temperature field from the \( \eta \) vertical coordinate to a pressure level and then to a CRCM Gal-Chen level in nesting and diagnostic procedures. The coarse vertical resolution is responsible for this smoothing in regions where vertical temperature gradient reverses such as at the tropopause. The vertical profile of the difference between domain-averaged temperatures (Fig. 11b) shows a warm bias of CRCM near the tropopause.

Figures 11c and 11d present the same information as Figs. 11a and 11b but for the 30-level version of CRCM. It can be seen that increasing the number of vertical levels greatly reduces the tropopause smoothing. It has to be emphasized here that increasing the number of vertical levels did not affect the other fields presented so far in any noticeable way, and hence few of the 30-level results will be shown in the rest of this article.

Figure 12 shows vertical profiles of domain-averaged mean specific humidity for CRCM and GCMII and their difference. Clearly, CRCM is drier than GCMII over a
thick layer (from 950 hPa to 400 hPa). A detailed study of CRCM simulations indicated that strong spurious precipitation events appear in the nesting zone of the model. These are caused by fictitious supersaturation induced by vertical interpolation of GCMII fields needed to drive the regional model. Spectral transform and spatial interpolation (vertical and horizontal) from the $\eta$ levels of GCMII Gaussian grid to the Gal-Chen polar-stereographic grid of CRCM produce regions where the relative humidity can reach almost 110%. This is illustrated in Fig. 13 where GCMII relative humidity for 1 January at 0000 UTC is presented at 900 hPa (a), 800 hPa (b), and 700 hPa (c) once interpolated on the CRCM grid. As can be seen, values as high as 108% are reached over New York State and the southeastern Atlantic Ocean. Figure 13d presents a vertical profile over the point identified by a cross over New York State; supersaturation exists in a deep layer below 550 hPa. Note that the original GCMII data (on Gaussian grid and on $\eta$ coordinates) did not exhibit such excessive supersaturation. When such supersaturated profiles are used to drive the CRCM, the convective adjustment scheme generates intense precipitation, which could possibly trigger deep convection as a response to the unstable atmosphere resulting from latent heat release. The process can withdraw significant amounts of water vapor from the atmosphere explaining the deficit in the 950–400-hPa layer. This artificially induced convection takes place mostly in the nesting zone when the supersaturated air enters the CRCM grid. This problem has been corrected in a more recent version of the model by using dewpoint depression as a moisture variable instead of specific humidity in the interpolation procedure.

Figure 14 shows that the total cloud cover is more important in CRCM than in GCMII, the difference reaching 40% on the southern part of the grid. The vertical profile of domain-averaged cloud cover (Fig. 15a) shows that there is more cloud in CRCM than in GCMII over the complete troposphere. We suspect that this increase in the cloud cover field has something to do with the semi-Lagrangian nature of the moisture advection in CRCM. Such an increase in the cloud cover field has been observed with GCMII when a semi-Lagrangian moisture advection was used instead of the regular spectral advection (R. Harvey 1997, personal communication). We cannot explain this behavior at the moment but an investigation of the moisture budget is under way and should shed some light on this matter.

It can be seen in Fig. 15, where the vertical profile of domain-averaged cloud cover is presented, that the excess in very low-level clouds is larger in the 30-level version of the model than in the 10-level version. This results from the crude representation of vertical moisture transport in the GCMII convective adjustment scheme. As explained in section 2b(3), this vertical transport between two levels takes moisture from the lower level and moves it to the upper level until one of the following is satisfied: 1) the upper layer is saturated; 2) supersat-

![Figure 14](image)
uration is removed from the lower level. If the upper level is saturated before removing supersaturation of the lower level, the remaining supersaturation of the lower level is removed as precipitation. When the number of levels increases, layer thicknesses are diminished and less water is needed to saturate the upper level, therefore reducing the vertical transport. The upward moisture transport is less efficient with many thin layers than with fewer thicker ones. A further and more complete validation of this scheme will be required for its utilization with more than 20 levels.

The January mean precipitation (Fig. 16) exhibits more structure in CRCM than in GCMII. This finescale distribution in CRCM results from a better definition of precipitating areas in weather systems. In particular, the strong convective activity over the Gulf of Mexico in January is clearly visible. Despite CRCM's stronger mean precipitation rates in the southern part of the domain, with values exceeding 6 mm day$^{-1}$, the domain-averaged January mean precipitation rate is lower in CRCM (2.46 mm day$^{-1}$) than in GCMII (2.78 mm day$^{-1}$). A possible explanation for the decrease in the domain-averaged precipitation rate can be the spurious convective activity in the nesting zone that is continu-
Fig. 16. January mean of the precipitation rate for CRCM-10 (a), GCMII (b), and differences between CRCM-10 and GCMII (c). Contour intervals are 1, 2, 3, 4, 5, and 7 mm day$^{-1}$ for (a) and (b). The interval is 5 mm day$^{-1}$ for (c).

Transients-odd standard deviations of the various CRCM fields were also computed [details of transient-odd standard deviation calculation can be found in Laprise et al. (1998)]. In general, the transient-odd variability is stronger in CRCM than in GCMII for fields that are related to moisture and that are most affected by the increased resolution of CRCM, whereas MSLP, geopotential height, and temperature show similar transient-odd variability. As an example of a strongly influenced field, the transient-odd standard deviation variability for January precipitation is shown in Fig. 17. It can be seen that the increased resolution of CRCM influences not only the mean distribution of precipitation but also its temporal variability.

4. Conclusions

A description of the dynamical formulation and physical parameterization of a new RCM based on a very efficient SISL numerical integration scheme is presented. The SISL-based numerics allow CRCM to use very long time steps (more than five times longer than Eulerian models) for the integration of the fully elastic nonhydrostatic Euler equations. The CRCM uses the complete physical parameterization of subgrid-scale processes of GCMII leading to a harmonious regional climate modeling system where both models (driving and driven) share the same physics. The nesting of a height-based vertical discretization regional gridpoint model within a pressure-based vertical discretization spectral global model is obviously not trivial; some problems related to interpolations between these different grids are reviewed.

Two configurations of CRCM vertical levels are used to generate a pair of two-month-long winter simulations in order to analyze the CRCM behavior in climatological integrations. It has been observed in the two simulations that CRCM rapidly develops timescale structures from the low-resolution GCMII data used to initialize and drive the regional model. Basic statistics, computed to evaluate the influence of those timescale details from a climatological point of view, have shown that smallscale features are still present in the monthly statistics. Hydrology-related fields such as water vapor distribution, precipitation, and cloud cover are the most affected by the increased resolution of CRCM. However, despite the increased detail in CRCM simulations, the general GCMII patterns are reproduced. This was an important goal for the CRCM simulations. In that CRCM is being
used to refine low-resolution GCMII simulations, it has to follow the general features of the driving model and not modify them. These short integrations were not intended to evaluate the CRCM climate but mainly to demonstrate the potential of the SISL integration scheme in climate simulations. A complete analysis of the CRCM present and altered climate is presented in Laprise et al. (1998).

The simulations presented in this paper allowed the identification of some deficiencies in CRCM behavior. Among them, the MSLP drift resulting from mass convergence, the spurious precipitation events in the nesting zone, the water vapor concentration in the low troposphere, and the related increased low-level cloud coverage. These main drawbacks have been corrected in later versions of the model. The nesting of MSLP, the use of dewpoint depression instead of specific humidity as a humidity variable for driving the water vapor field, and the correction of the threshold value for cloud are improvements already made to correct identified deficiencies. The convective adjustment, however, seems to need some refinements to account for the increased resolution of CRCM. A completely new third-generation CCCma physical parameterization package (in which a new convection scheme and a more advanced land surface scheme are used) is being implemented in a newer version of CRCM.

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