RAMS

The Regional Atmospheric Modeling System

Technical Description

DRAFT

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1. General equations

The general equations for RAMS are described below. The equations are the standard hydrostatic or non-hydrostatic Reynolds-averaged primitive equations. All variables, unless otherwise denoted, are grid-volume averaged quantities where the overbar has been omitted. The horizontal and vertical grid transformations are omitted in this section for clarity. These transformations will be described later. The symbols are defined in Table 1-1.

The non-hydrostatic equations are:

Equations of motion:

$$\frac{\partial u}{\partial t} = -u\frac{\partial u}{\partial x} - v\frac{\partial u}{\partial y} - w\frac{\partial u}{\partial z} - \theta\frac{\partial \pi'}{\partial x} + fv + \frac{\partial}{\partial x}\left(K_m\frac{\partial u}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_m\frac{\partial u}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_m\frac{\partial u}{\partial z}\right) \quad (1)$$

$$\frac{\partial v}{\partial t} = -u\frac{\partial v}{\partial x} - v\frac{\partial v}{\partial y} - w\frac{\partial v}{\partial z} - \theta\frac{\partial \pi'}{\partial y} - fu + \frac{\partial}{\partial x}\left(K_m\frac{\partial v}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_m\frac{\partial v}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_m\frac{\partial v}{\partial z}\right) \quad (2)$$

$$\frac{\partial w}{\partial t} = -u\frac{\partial w}{\partial x} - v\frac{\partial w}{\partial y} - w\frac{\partial w}{\partial z} - \theta\frac{\partial \pi'}{\partial z} - \frac{g\theta_v'}{\theta_0} + \frac{\partial}{\partial x}\left(K_m\frac{\partial w}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_m\frac{\partial w}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_m\frac{\partial w}{\partial z}\right) \quad (3)$$

Thermodynamic equation:

$$\frac{\partial \theta_{il}}{\partial t} = -u\frac{\partial \theta_{il}}{\partial x} - v\frac{\partial \theta_{il}}{\partial y} - w\frac{\partial \theta_{il}}{\partial z} + \frac{\partial}{\partial x}\left(K_h\frac{\partial \theta_{il}}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_h\frac{\partial \theta_{il}}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_h\frac{\partial \theta_{il}}{\partial z}\right) + \left(\frac{\partial \theta_{il}}{\partial t}\right)_{rad} \quad (4)$$

Water species mixing ratio continuity equation:

$$\frac{\partial r_n}{\partial t} = -u\frac{\partial r_n}{\partial x} - v\frac{\partial r_n}{\partial y} - w\frac{\partial r_n}{\partial z} + \frac{\partial}{\partial x}\left(K_h\frac{\partial r_n}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_h\frac{\partial r_n}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_h\frac{\partial r_n}{\partial z}\right) \quad (5)$$

Mass continuity equation:

$$\frac{\partial \pi'}{\partial t} = -\frac{R\pi_0}{c_v\rho_0\theta_0}\left(\frac{\partial \rho_0\theta_0 u}{\partial x} + \frac{\partial \rho_0\theta_0 v}{\partial y} + \frac{\partial \rho_0\theta_0 w}{\partial z}\right) \quad (6)$$

The hydrostatic option in RAMS replaces the vertical equation of motion and the mass continuity equation with:

Hydrostatic equation:

$$\frac{\partial \pi}{\partial z} = -\frac{g}{\theta_v} + g\left(r_T - r_v\right) \tag{7}$$

$$\frac{\partial \rho u}{\partial x} + \frac{\partial \rho v}{\partial y} + \frac{\partial \rho w}{\partial z} = 0 \tag{8}$$

Table 1-1 Symbols used in this paper.

Symbol	Definition
u	east-west wind component
v	north-south wind component
w	vertical wind component
f	Coriolis parameter
K_m	eddy viscosity coefficient for momentum
K_h	eddy viscosity coefficient for heat and moisture
θ_{il}	ice-liquid water potential temperature
r_n	water mixing ratio species of total water, rain, pristine crystals, aggregates, and snow
ρ	density
con	subscript denoting tendency from convective parameterization
rad	subscript denoting tendency from radiation parameterization
res	subscript denoting tendency from resolvable scale microphysical parameterization
g	gravity
r_t	total water mixing ratio
r_v	water vapor mixing ratio
π	total Exner function
π’	perturbation Exner function
θ_v	virtual potential temperature
p	pressure

2. Grid structure

2.1. Grid stagger

The grid stagger used in RAMS is the standard C grid (Mesinger and Arakawa, 1976). All thermodynamic and moisture variables are defined at the same point with the velocity components, u, v, and w staggered 1/2 Δx, 1/2 Δy, and 1/2 Δz, respectively.

2.2. Map projection

The horizontal grid uses a rotated polar-stereographic projection, where the pole of the projection is rotated to an area near the center of the domain, thus minimizing the distortion of the projection in the main area of interest. The appropriate "map factors" are used in all horizontal derivative terms.

2.3. Terrain-following coordinate

The vertical structure of the grid uses the σ_z terrain-following coordinate system (Gal-Chen and Somerville, 1975; Clark, 1977; Tripoli and Cotton, 1982). It is a terrain-following σ_z coordinate system where the top of the model domain is exactly flat and the bottom follows the terrain. The coordinates in this system are defined as:

$$\begin{aligned} x^* &= x \\ y^* &= y \\ z^* &= H\left(\frac{z - z_g}{H - z_g}\right) \end{aligned} \tag{9}$$

where H is the height of the top of the grid and z_g is local topography height as a function of x and y. Derivative terms can be written (Clark, 1977) in tensor notation as:

$$\frac{\partial \phi}{\partial x_i} = \left(\frac{1}{a}\right)\left(\frac{\partial a b^{ij} \phi}{\partial x_j^*}\right) \tag{10}$$

$$a\left(x^*, y^*\right) = 1 - \frac{z_g\left(x^*, y^*\right)}{H}$$

and the tensor b_{ij} is written as:

$$b^{ij} = \begin{bmatrix} 1 & 0 & \frac{1}{a}\frac{\partial z_g}{\partial x}\left(\frac{z^*}{H}-1\right) \\ 0 & 1 & \frac{1}{a}\frac{\partial z_g}{\partial y}\left(\frac{z^*}{H}-1\right) \\ 0 & 0 & \frac{1}{a} \end{bmatrix} \tag{11}$$

The relationship between the Cartesian wind components and the components in the transformed system is:

$$\begin{aligned} u^* &= u \\ v^* &= v \\ w^* &= \left(uab^{13} + vab^{23} + w\right)/a \end{aligned} \tag{12}$$

This transformation can be viewed as a simple mapping of the horizontal velocity components to the terrain-following system since they remain horizontal in Cartesian space. This eliminates the complication of dealing with non-orthogonal velocity components. The vertical component, w^*, has an imposed value of 0 at z*, which implies no mass flux through the ground surface.

2.4. Vertical levels

In using variable vertical grid spacing to increase resolution near the ground, it is preferable to avoid large expansion ratios, $\tilde{R}_j$, between consecutive levels, where $\tilde{R}_j$ is defined for any level j as

$$\tilde{R}_j \equiv \frac{\left(\Delta\tilde{Z}\right)_j}{\left(\Delta\tilde{Z}\right)_{j-1}} \tag{13}$$

and

$$\left(\Delta\tilde{Z}\right)_j \equiv \tilde{Z}_{j+1} - \tilde{Z}_j \tag{14}$$

using the notation in Fig. 1. (??????) We thus usually employ a fairly small fixed value of $\tilde{R}_j$ over many levels such that the grid varies geometrically from small $\tilde{Z}_j$ values near the ground to much larger ones aloft. For example, with $\tilde{R}_j = 1.15$, $\tilde{Z}_j$ increases tenfold in about 16 levels.

Often, we impose an upper bound on $\tilde{Z}_j$ which leads to $\tilde{R}_j = 1$ when the bound is reached. We also desire that the nested grid (NG) levels be stretched geometrically when those on the parent grid (PG) are. Since it is not always required that $\tilde{R}_j$ be constant, we define an average PG stretch ratio $\tilde{R}_j$ for the grid cell between $\tilde{Z}_{j+1}$ and $\tilde{Z}_j$ as

$$R_j = \left[\frac{\left(\Delta \tilde{Z}\right)_{j+1}}{\left(\Delta \tilde{Z}\right)_{j-1}} \right]^{0.5} \tag{15}$$

which is the geometric mean of $\tilde{R}_{j+1}$ and $\tilde{R}_j$. The appropriate constant stretch ratio $\tilde{r}_k$ applied to all NG cells inside this PG cell is

$$\tilde{r}_k = \left(R_j\right)^{1/N} \tag{16}$$

where

$$\tilde{r}_k \equiv \frac{\left(\Delta \tilde{z}\right)_k}{\left(\Delta \tilde{z}\right)_{k-1}} \tag{17}$$

and

$$\left(\Delta \tilde{z}\right)_k \equiv \tilde{z}_{k+1} - \tilde{z}_k \tag{18}$$

This relation, coupled with the requirement

$$\sum_{k=1}^{N} \left(\Delta \tilde{z}\right)_k = \left(\Delta \tilde{Z}\right)_j \tag{19}$$

determines all levels $\tilde{z}_k$. In cases where $\tilde{R}_j$ is constant with height, this formulation causes $\tilde{r}_k$ to also be constant, even between PG cells.

Similarly, we make distances between $\tilde{Z}$ and Z levels (and $\tilde{z}$ and z levels) expand geometrically by setting

$$\frac{\tilde{Z}_{j+1} - Z_j}{Z_j - \tilde{Z}_j} = \left(R_j\right)^{0.5} \tag{20}$$

and

$$\frac{\tilde{z}_{k+1} - z_k}{z_k - \tilde{z}_k} = \left(r_k\right)^{0.5} \tag{21}$$

3. Time differencing

RAMS has several options for the basic time differencing schemes. For the non-hydrostatic model, the user has a choice of either a forward-backward first order scheme, a leapfrog scheme (Tripoli and Cotton, 1982), or a hybrid scheme (Tripoli, 1992, Clark ????) which consists of forward time differencing for the thermodynamic variables and leapfrog differencing for the velocity components and pressure. The hydrostatic model uses the forward-backward scheme.

All of the above options are formulated with "time-split" time differencing schemes (Tripoli and Cotton, 1982; Tremback et al., 1985) that are similar to the time-split scheme of Klemp and Wilhelmson (1978) and the split explicit scheme of Gadd (1978). The basic idea behind these schemes is to "split" off in a series of smaller timesteps those terms in the equation that are responsible for the propagation of the fast wave modes. In the non-hydrostatic model, the fast modes are the acoustic and gravity waves while in a hydrostatic model, the fast modes are the external gravity wave and the Lamb wave.

3.1. Non-hydrostatic

The time differencing schemes can be demonstrated as follows for simplified two-dimensional, dry, inviscid, quasi-Boussinesq equation sets where the vertical and horizontal coordinate transformations have been removed for clarity. For the non-hydrostatic option, the basic equations are:

$$\frac{\partial u}{\partial t} + \theta_0 \frac{\partial \pi'}{\partial x} = -u\frac{\partial u}{\partial x} - w\frac{\partial u}{\partial z} + fv = F_u \tag{22}$$

$$\frac{\partial w}{\partial t} + \theta_0 \frac{\partial \pi'}{\partial z} = -u\frac{\partial w}{\partial x} - w\frac{\partial w}{\partial z} + g\left(\frac{\theta'}{\theta_0}\right) = F_w \tag{23}$$

$$\frac{\partial \theta}{\partial t} = -u\frac{\partial \theta}{\partial x} - w\frac{\partial \theta}{\partial z} = F_\theta \tag{24}$$

$$\frac{\partial \pi'}{\partial t} = -\frac{c^2}{\rho_0 \theta_0^2}\left[\frac{\partial u \rho_0 \theta_0}{\partial x} + \frac{\partial w \rho_0 \theta_0}{\partial z}\right] \tag{25}$$

where c is the speed of sound.

As mentioned the object of the time-split scheme is to compute the terms that are primarily responsible for the propagation of the fast modes on a smaller timestep than the slower modes (such as horizontal advection and the Coriolis force). The computational procedure is then as follows for a forward-backward time differencing scheme.

1. The right hand side of [22] and [24], F_u, F_w, and F_θ are computed.

2. The velocity components are stepped to $t + \Delta t_s$.

$$u^{t+\Delta t_s} = u^t - \Delta t_s \left[\theta_0 \frac{\partial \pi'}{\partial x} - F_u \right]$$
$$w^{t+\Delta t_s} = w^t - \Delta t_s \left[\theta_0 \frac{\partial \pi'}{\partial z} - F_w \right] \tag{26}$$

3. Pressure at $t + \Delta t_s$ is computed with [25], using the newly updated velocity components at $t + \Delta t_s$.

4. The small timestep is repeated *n* times until $n\Delta t_s = \Delta t_L$.

In addition, there is a vertical implicit scheme (Tripoli and Cotton, 1982; Durran, 1983) which allows a longer small time step by solving the vertical pressure gradient and vertical divergence terms in an implicit numerical manner. In finite difference form, (23) and (25) can be written in implicit form as:

$$w_k^{t+\Delta t_s} = w_k^* + d_k \left(\pi'_k \right)^{t+\Delta t_s} + e_k \left(\pi'_{k+1} \right)^{t+\Delta t_s} \tag{27}$$

$$\left(\pi'_k \right)^{t+\Delta t_s} = \pi'^*_k + f_k \left(w \rho_0 \theta_0 \right)_k^{t+\Delta t_s} + g_k \left(w \rho_0 \theta_0 \right)_{k+1}^{t+\Delta t_s} \tag{28}$$

where the subscript *k* refers to a vertical level index and:

$$w_k^* = w_k^t + \Delta t_s F_w$$
$$\pi'^*_k = \left(\pi'_k \right)^t + \Delta t_s F_\pi$$
$$d_k = \frac{\Delta t_s \theta_0}{\Delta z_k}$$
$$e_k = -c_k \tag{29}$$
$$f_k = -\frac{c^2}{\rho_0 \theta_0^2} \frac{\Delta t_s}{\Delta z_k}$$
$$g_k = -f_k$$

Substituting (28) into (27) and rearranging yields:

$$A_k w_{k-1}^{t+\Delta t_s} + B_k w_k^{t+\Delta t_s} + C_k w_{k+1}^{t+\Delta t_s} = D_k \tag{30}$$

where

$$
\begin{aligned}
A_k &= d_k g_k \\
B_k &= d_k f_k + d_k g_{k+1} - 1 \\
C_k &= e_k f_{k+1} \\
D_k &= w_k^* + d_k \pi_k'^* + e_k \pi_{k+1}'^*
\end{aligned}
\tag{31}
$$

Equation (30) can be easily solved with a tri-diagonal matrix technique. These results for w_k^* would then be used in (25) to solve for $\left(\pi_k'\right)^{t+\Delta t_s}$ to complete the small timestep. In RAMS, there are some other considerations such as the interaction of the top boundary conditions with the implicit scheme and the inclusion of a Crank-Nicholson scheme in the matrix solution.

RAMS also contains a modification of *c*, the speed of sound, to allow a longer small timestep. This technique (Chorin, 1967; Drogemeier and Anderson,19??; Tripoli, 1992) lowers *c* in (25). Experience has shown that a reduction of *c* by a factor between .4 and 1 usually has no significant effect on the model solution.

3.2. Hydrostatic

For the hydrostatic option, the time differencing schemes can be demonstrated for a simplified two-dimensional, dry, inviscid equation set.

$$
\begin{aligned}
&\frac{\partial u}{\partial t} + \theta \frac{\partial \pi}{\partial x} = F_u \\
&F_u = -u \frac{\partial u}{\partial x} - w \frac{\partial u}{\partial z} + fv
\end{aligned}
\tag{32}
$$

$$
\begin{aligned}
&\frac{\partial \theta}{\partial t} + w \frac{\partial \theta}{\partial z} = F_\theta \\
&F_\theta = -u \frac{\partial \theta}{\partial x}
\end{aligned}
\tag{33}
$$

$$
\frac{\partial \rho u}{\partial t} + \frac{\partial \rho w}{\partial z} = 0 \tag{34}
$$

$$
\frac{\partial \pi}{\partial z} = -\frac{g}{\theta} \tag{35}
$$

The computational procedure for the hydrostatic equations is then as follows for a forward-backward time differencing scheme.

a) The right hand side of (32) and (33), F_u and F_θ are computed.

b) θ is stepped forward to time level $t+\Delta t_L$ where Δt_L is the long timestep.

c) θ is stepped forward to time level $t+\Delta t_s$ where Δt_s is the small timestep.

d) Pressure at $t+\Delta t_s$ is computed with (35).

e) The horizontal velocity is stepped to $t+\Delta t_s$.

f) The vertical velocity at $t+\Delta t_s$ is computed with (34).

g) The pressure boundary condition is updated to $t+\Delta t_s$ with (34) using the divergence at the $t+\Delta t_s$ level.

h) The small timestep, c) through g), is repeated n times until $n\Delta t_s = \Delta t_L$.

4. Nested grids

Grid nesting is used in RAMS to provide high spatial resolution in selected locations, while covering a large domain at lower resolution. A nested grid occupies a region within the computational domain of its coarser parent grid, and coincides with, rather than replaces, the parent grid mesh in that region. Any number of nested grids may be used, within the limits of available computer memory. The grids may be configured as a telescoping sequence where parent grids are themselves nested within coarser grids, with several nested grids having the same parent, or as a combination of these. A nested grid may, but need not, extend from the ground to the model top, and it can be moved horizontally during a model simulation with the velocity of a traveling system such as a thunderstorm or tropical cyclone in order to keep the system within the high resolution grid. Mesh refinement on a nested grid may be prescribed for the horizontal or vertical directions, or both. The refinement is always required to be an integer ratio, such that a whole number of nested grid cells are contained within a parent grid cell. Since RAMS uses a terrain-following vertical coordinate, the terrain definition must be compatible between grids such that the vertical levels on both grids coincide properly. Vertical nesting in RAMS may be used on stretched grids to provide even greater vertical resolution on a nested grid (NG) relative to its parent grid (PG). In addition, the vertical nesting ratio may be varied with height.

RAMS employs the grid nesting technique described in Clark and Farley (1984) and Clark and Hall (1991), with a generalization for stretched grids and a spatially variable nesting ratio described in Walko et al. (1993). The technique carries out a two-way communication of all prognostic variables between any nested grid and its parent grid. Communication from the parent to the nested grid is accomplished immediately following a timestep on the parent grid which updates the prognostic fields. The updated values are interpolated sequentially in the coordinate directions, to the locations where they are defined on the boundaries of the nested grid. The NG values are replaced by the interpolated values. A NG usually runs with a shorter time step than the parent grid. Thus, linear interpolation in time of the PG values is also performed for intermediate nested grid timesteps. The NG is then updated in a series of smaller steps until it has caught up to the simulation time of the PG. At this time, the reverse communication is accomplished by averaging the prognostic variables over each set of NG cells which occupy a single parent grid cell, and replacing the PG cell value with that average.

The prognostic velocity components, potential temperature, and moisture variables are multiplied by density prior to their communication from either grid to the other. The interpolation and the averaging operators are both designed to conserve the volume integrals of these density-weighted quantities between the grids. This implies that the nesting algorithm conserves mass, momentum, and internal thermodynamic energy.

Most spatial interpolations are quadratic, except for a correction which maintains conservation of the volume integral. A cubic interpolation is performed for velocity components in the direction of the given component, for consistency with other velocity interpolations and the constraint of mass continuity (Walko et al., 1993). However, some NG velocity component values are defined on the same plane (oriented normal to the velocity component) as the PG values; in such cases, interpolation and averaging are performed only within the two dimensions of the plane.

If a nested grid extends to the top and bottom of the parent grid, interpolations from the are applied only to its lateral boundaries during model integration. Interpolation to the interior of the nested grid is performed only for initialization. For any top or bottom nested grid boundary which does

not coincide with a boundary of the parent grid, interpolations are carried out to define the nested grid values from the parent grid.

5. Advection

RAMS uses two types of advection schemes for the time differencing options in the model simulations, the standard leapfrog-type schemes and the forward-upstream schemes (Tremback, et al, 1987). The leapfrog schemes are used for all variables in the leapfrog time differencing option and for the velocity components in the hybrid time differencing option.

The advective schemes are configured in flux form in order to conserve mass and momentum. Considering the *x*-direction, the advective terms in (1)-(4) then can be generically written assuming constant grid spacing and omitting topographical and spherical transformations for clarity, as:

$$\begin{aligned} -u\frac{\partial \phi}{\partial x} &= -\frac{1}{\rho}\frac{\partial \rho u \phi}{\partial x} + \phi\frac{\partial \rho u}{\partial x} \\ &= -\frac{1}{\rho_j \Delta x}\left[\left(\left(\rho F\right)_{j+1/2} - \left(\rho F\right)_{j-1/2}\right) - \phi_j\left(\left(\rho u\right)_{j+1/2} - \left(\rho u\right)_{j-1/2}\right)\right] \end{aligned} \tag{36}$$

where u is the wind component in the x direction, ρ is the air density, and ϕ is the variable to be advected. The subscript j references a particular grid point.

5.1. Leapfrog advection

The leapfrog advection fluxes are either the second-order or fourth-order of the centered-in-time, centered-in-space leapfrog advection scheme. The expressions for the fluxes, F, are

Second-order leapfrog fluxes:

$$F_{j+1/2} = \left(u\phi\right)_{j+1/2} \tag{37}$$

Fourth-order leapfrog fluxes:

$$F_{j+1/2} = u_{j+1/2}\left(-\frac{1}{12}\phi_{j-1} + \frac{7}{12}\phi_j + \frac{7}{12}\phi_{j+1} - \frac{1}{12}\phi_{j+2}\right) \tag{38}$$

5.2. Forward advection

The forward advection fluxes are computed with the second-order or the sixth-order forward upstream advection schemes that were derived and tested by Tremback et al. (1987). These schemes are in the same family of schemes as the classical first-order forward upstream scheme and the much used Crowley (1968) second-order scheme. As described in detail by Tremback et al. (1987), two different forms of the flux scheme can be derived. The first, following the methodology of Crowley (1968), fits a polynomial to the field being advected then integrates the function. However, for schemes of order three and greater, this form does not reduce to the advective form for constant grid spacing and advecting velocity. The second form, also described by Bott (1989), makes this requirement. It is more accurate than the first but does require that the

grid spacing be constant. Therefore, the first form, called the integrated flux form, is used in the vertical where the grid spacing is stretched to provide higher resolution near the ground. The second form is used in the horizontal where the grid spacing is constant in any one direction. The forward advection fluxes can be written as (where $\alpha = u\,\Delta t/\Delta x$) :

Second-order fluxes:

$$F_{j+1/2}\frac{\Delta t}{\Delta x} = \frac{\alpha}{2}\left(\phi_j + \phi_{j+1}\right) + \frac{\alpha^2}{2}\left(\phi_j - \phi_{j+1}\right) \tag{39}$$

Sixth-order integrated fluxes:

$$\begin{aligned} F_{j+1/2}\frac{\Delta t}{\Delta x} = {} & \frac{\alpha}{256}\left(3\phi_{j-2} - 25\phi_{j-1} + 150\phi_j + 150\phi_{j+1} - 25\phi_{j+2} + 3\phi_{j+3}\right) + \\ & + \frac{\alpha^2}{3840}\left(9\phi_{j-2} - 125\phi_{j-1} + 2250\phi_j - 2250\phi_{j+1} + 125\phi_{j+2} - 9\phi_{j+3}\right) + \\ & + \frac{\alpha^3}{288}\left(-5\phi_{j-2} + 39\phi_{j-1} - 34\phi_j - 34\phi_{j+1} + 39\phi_{j+2} - 5\phi_{j+3}\right) + \\ & + \frac{\alpha^4}{192}\left(-\phi_{j-2} + 13\phi_{j-1} - 34\phi_j + 34\phi_{j+1} - 13\phi_{j+2} + \phi_{j+3}\right) + \\ & + \frac{\alpha^5}{240}\left(\phi_{j-2} - 3\phi_{j-1} + 2\phi_j + 2\phi_{j+1} - 3\phi_{j+2} + \phi_{j+3}\right) + \\ & + \frac{\alpha^6}{720}\left(\phi_{j-2} - 5\phi_{j-1} + 10\phi_j - 10\phi_{j+1} + 5\phi_{j+2} - \phi_{j+3}\right) + \end{aligned} \tag{40}$$

Sixth-order constant grid fluxes:

$$\begin{aligned} F_{j+1/2}\frac{\Delta t}{\Delta x} = {} & \frac{\alpha}{60}\left(\phi_{j-2} - 8\phi_{j-1} + 37\phi_j + 37\phi_{j+1} - 8\phi_{j+2} + \phi_{j+3}\right) + \\ & + \frac{\alpha^2}{360}\left(2\phi_{j-2} - 25\phi_{j-1} + 245\phi_j - 245\phi_{j+1} + 25\phi_{j+2} - 2\phi_{j+3}\right) + \\ & + \frac{\alpha^3}{48}\left(-\phi_{j-2} + 7\phi_{j-1} - 6\phi_j - 6\phi_{j+1} + 7\phi_{j+2} - \phi_{j+3}\right) + \\ & + \frac{\alpha^4}{144}\left(-\phi_{j-2} + 11\phi_{j-1} - 28\phi_j + 28\phi_{j+1} - 11\phi_{j+2} + \phi_{j+3}\right) + \\ & + \frac{\alpha^5}{240}\left(\phi_{j-2} - 3\phi_{j-1} + 2\phi_j + 2\phi_{j+1} - 3\phi_{j+2} + \phi_{j+3}\right) + \\ & + \frac{\alpha^6}{720}\left(\phi_{j-2} - 5\phi_{j-1} + 10\phi_j - 10\phi_{j+1} + 5\phi_{j+2} - \phi_{j+3}\right) + \end{aligned} \tag{41}$$

6. Turbulent mixing parameterization

6.1. Deformation-based parameterization

The horizontal and vertical grid spacings configured in the model determine the spatial scales of prognostic field variables that can be explicitly resolved and those which cannot. Reynolds averaging of the prognostic differential equations for momentum and conservative scalars is performed to partition advective transport into resolved and unresolved components.

The unresolved flux components may be expressed in terms of covariances of the form $\overline{u_i' u_j'}$ for momentum, and $\overline{u_i' \phi'}$ for scalars, where subscripts *i* and *j* denote spatial directions *[1,2,3]*, u_i is the transporting velocity component, u_j is the transported velocity component, ϕ represents the transported scalar, an overbar represents the Reynolds average, and a prime the deviation from that average.

The contribution to the tendency of the resolved variables due to turbulent transport is given by the convergence of the turbulent fluxes

$$\left(\frac{\partial u_j}{\partial t}\right)_{TURB} = \frac{\partial}{\partial x_i}\left(\overline{u_i' u_j'}\right) \tag{42}$$

and

$$\left(\frac{\partial \phi}{\partial t}\right)_{TURB} = \frac{\partial}{\partial x_i}\left(\overline{u_i' \phi'}\right) \tag{43}$$

The summation convention applies in [1], [2], and [8] but not in the remainder of this Section.

RAMS parameterizes the unresolved transport using K-theory, in which the covariances are evaluated as the product of an eddy mixing coefficient and the gradient of the transported quantity. For scalars, this parameterization takes the form

$$\overline{u_i' \phi'} = -K_{hi} \frac{\partial \phi}{\partial x_i} \tag{44}$$

where K_{hi} is the eddy mixing coefficient for scalars which applies to the *i*-direction. K_{hi} is never negative, which restricts the parameterized eddy fluxes to always be down-gradient.

For velocity components, two different forms are used, depending on the scales of motion resolved by the model grid. When the horizontal grid spacing is comparable to the vertical spacing so that convective motions are resolved, the Reynolds stresses are evaluated from

$$\overline{u_i' u_j'} = -K_{mi}\left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i}\right) \tag{45}$$

which applies to the *i*-direction for *i* and *j = [1,2,3]*, where K_{mi} is the eddy mixing coefficient for momentum. In this case, it is assumed that $K_{mi} = K_{mj}$, and therefore, $\overline{u_i' u_j'} = \overline{u_j' u_i'}$.

This symmetry of the Reynolds stress components is a basic physical property of a fluid. Violation of this property in numerical models is equivalent to applying a fictitious external torque to the fluid wherever a violation occurs. In simulations where convective motions are resolvable, this may lead to significant errors in the numerical solution.

If the horizontal grid spacing is much larger than the vertical spacing, which prevents explicit representation of convective vertical motion, it is not essential that Reynolds stresses be symmetric between the vertical and a horizontal direction. The constraint imposed by low horizontal resolution in the model prevents the solution from being strongly affected by fictitious external horizontal torque. On the other hand, relatively coarse horizontal grids require a larger value of K_{mi} in the horizontal directions than in the vertical. This horizontal mixing coefficient is larger in magnitude than any physical transport by turbulent eddies, and is required purely for numerical stability. Consequently, asymmetry of the Reynolds stresses which involve the vertical direction is a practical requirement. Hence, for coarse horizontal grid spacing we apply [4] only in the horizontal directions by restricting *i* and *j* to *[1,2],* and use the following expression whenever i and/or *j* is 3,

$$\overline{u_i' u_j'} = -K_{mi}\left(\frac{\partial u_i}{\partial x_j}\right) \tag{46}$$

There are currently four basic options for computing K_{mi} and K_{hi}. Two of these are based on the Smagorinsky (1963) scheme which relates the mixing coefficients to the fluid strain or deformation rate, and include corrections for the influence of Brunt-Vaisala frequency (Hill, 1974) and Richardson number (Lilly, 1962).

These are purely local schemes in which the mixing coefficients depend only on the local and current flow properties. The other two options diagnose mixing coefficients from a parameterized subgrid-scale turbulent kinetic energy (TKE) which is prognosed in the model. Although the diagnosis is based on the local current value of TKE, these schemes are regarded as non-local because the TKE can be generated elsewhere in the domain and transported to new locations.

Of the two local schemes, and of the two TKE schemes, one applies in cases of small horizontal grid spacing where [4] is used, and the other applies with large horizontal grid spacing where [5] is used.

The eddy mixing coefficient for the local deformational scheme with small horizontal grid spacing is given by

$$K_{mi} = \rho\left(C_x \Delta x\right)\left(C_z \Delta z\right)\left\{S_3 + F_H\left[\max\left(0,-F_B\right)\right]^{0.5}\right\}\left[\max\left(0,1-R_{hm}R_i\right)\right]^{0.5} \tag{47}$$

$$K_{hi} = -R_{hm}K_{mi} \tag{48}$$

where C_x and C_z are dimensionless coefficients multiplying the horizontal and vertical grid spacings Δx and Δz to obtain characteristic horizontal and vertical mixing length scales, respectively, F_H is an optional flag set to either 0 or 1, F_B is the Brunt-Vaisala frequency, R_i is the Richardson number, and R_{hm} is a specified ratio of scalar to momentum mixing coefficients.

S_3 is the magnitude of the three-dimensional rate-of-strain tensor, given by

$$S_3 = \left[\left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right)^2 \right]^{0.5} \tag{49}$$

The bracketed term multiplied by F_H is an optional enhancement of mixing in convectively unstable cases proposed by Hill (1974). It is usually included in RAMS to reduce the degree of superadiabatic lapse rate required in a convective boundary layer to transport heat upward at a given rate. The bracketed term containing the Richardson number has a minimum value of zero for stable situations, in which case K_{mi} becomes zero, and becomes greater than 1 in superadiabatic conditions, which amplifies K_{mi}.

When the horizontal grid spacing is large, the local deformational scheme uses a horizontal eddy mixing coefficient given by

$$K_i = \rho_0 \max\left[K_{mh}, \left(C_x \Delta x\right)^2 \left\{ S_2^{0.5} \right\} \right] \tag{50}$$

where the horizontal strain rate S_2 is given by [8] with i and j limited to the horizontal directions *[1,2]*. K_{mh} is a minimum value imposed on the horizontal mixing coefficient, and is given by

$$K_{mh} = 0.075 K_A \left(\Delta x^{4/3} \right) \tag{51}$$

where K_A is a user-specified coefficient of order 1. The proportionality between the mixing coefficient and the *4/3* power of grid spacing is derived from scaling laws of turbulence in the inertial subrange, and has been verified empirically as valid for grid spacing as small as 2 cm and as large as 100 km, with $K_A = 1$. With large horizontal grid spacing, the vertical mixing coefficient is computed from [6] with C_Z replacing C_x, Δz replacing Δx, and S_1 replacing S_3, where S_1 is given by [8] with *i=[1,2]* and *j=[3]*.

6.2. Turbulent kinetic energy parameterization

We now consider the case where prognostic TKE is used to evaluate eddy mixing coefficients. If the horizontal grid spacing is large, the Mellor and Yamada (1974) scheme is used to compute the vertical mixing coefficients, while the local deformational scheme described above is used for horizontal mixing. The Mellor and Yamada scheme is an ensemble closure, which assumes that the Reynold's-averaged flow cannot resolve convection so that parameterized convection performs all vertical transport. If the horizontal grid spacing is small, so that convective motions are resolvable, the Deardorff (1980) scheme is used. This is a subgrid closure which parameterizes turbulent transport only by eddy motions smaller in scale than the smallest resolvable scales, and assumes that the resolved motions are capable of carrying the bulk of the turbulent transport. This transport is applied to both the horizontal and vertical directions.

The technique developed by Mellor and Yamada (1974, 1982) is a so-called level 2.5 scheme with modifications for a case of growing turbulence (Helfand and Labraga, 1988). The fields of wind (u and v), potential temperature (θ), and turbulent kinetic energy (e) are provided by the prognostic fields in RAMS. This scheme is based on the prognostic equation for the turbulent kinetic energy which is solved in the meteorological model.

Define the turbulent kinetic energy (TKE), *e*, as:

$$e = 0.5\left(\overline{u'^2} + \overline{v'^2} + \overline{w'^2}\right) \tag{52}$$

Following is the prognostic equation for *e*:

$$\begin{aligned}\frac{\partial e}{\partial t} = & -u\frac{\partial e}{\partial x} - v\frac{\partial e}{\partial y} - w\frac{\partial e}{\partial z} + \\ & \frac{\partial}{\partial x}K_e\frac{\partial e}{\partial x} + \frac{\partial}{\partial y}K_e\frac{\partial e}{\partial y} + \frac{\partial}{\partial z}K_e\frac{\partial e}{\partial z} + P_s + P_b + \varepsilon\end{aligned} \tag{53}$$

where P_s is the shear production term

$$P_s = K_m\left[\left(\frac{\partial u}{\partial x}\right)^2 + \left(\frac{\partial v}{\partial y}\right)^2\right] \tag{54}$$

and P_b is the buoyancy production term

$$P_b = -\frac{g}{\theta}K_h\frac{\partial \theta_v}{\partial z} \tag{55}$$

The expression for the dissipation term, ε, is given by

$$\varepsilon = a_e\frac{e^{3/2}}{l} \tag{56}$$

The vertical eddy diffusivities for momentum, heat, and TKE are computed by:

$$\begin{aligned}K_m &= S_m l\sqrt{2e} \\ K_h &= S_h l\sqrt{2e} \\ K_e &= S_e l\sqrt{2e}\end{aligned} \tag{57}$$

The wind and temperature enter these calculations in the form of nondimensional vertical gradients:

$$
\begin{aligned}
G_u &= \frac{l}{\sqrt{2e}}\frac{\partial u}{\partial z} \\
G_v &= \frac{l}{\sqrt{2e}}\frac{\partial v}{\partial z} \\
G_m &= G_u^2 + G_v^2 \\
G_h &= -\frac{g}{\theta}\frac{l^2}{2e}\frac{\partial \theta}{\partial z}
\end{aligned}
\tag{58}
$$

The turbulent length scale, *l*, is assumed after Mellor and Yamada (1982).

$$
\begin{aligned}
l &= \frac{\kappa(z + z_0)}{1 + \kappa(z + z_0)/l_\infty} \\
l_\infty &= 0.1\frac{\int_0^H z\sqrt{e}\,dz}{\int_0^H \sqrt{e}\,dz}
\end{aligned}
\tag{59}
$$

where κ is the Von Karman constant and z_0 is the roughness length.

An upper limit for *l* in stable conditions proposed by André et al. (1978) is given by:

$$
l \le 0.75\left[\frac{2e}{\left(\frac{g}{\theta}\frac{\partial \theta}{\partial z}\right)}\right]^{1/2}
\tag{60}
$$

The above condition implies the constraint: $G_h \ge -0.75^2$.

In the level 2.5 scheme, the functions S_m and S_h (nondimensional eddy diffusivities) depend on nondimensional gradients of wind and potential temperature:

$$
S_m = \frac{A_1\{1 - 3C_1 - 3A_2[B_2(1 - 3C_1) - 12A_1C_1 - 3A_2]G_h\}}{1 - 3A_2(7A_1 + B_2)G_h + 27A_1A_2^2(4A_1 + B_2)G_h^2 + 6A_1^2[1 - 3A_2(B_2 - 3A_2)G_h]G_m}
\tag{61}
$$

$$
S_h = A_2\frac{1 - 6A_1S_mG_m}{1 - 3A_2(4A_1 + B_2)G_h}
\tag{62}
$$

Empirical constants are assigned values following Mellor and Yamada (1982):

$$
\{A_1, A_2, B_1, B_2, C_1, S_e, a_e\} = \{0.92, 0.74, 16.6, 10.1, 0.08, 0.20, 2^{2/3}/16.6\}
\tag{63}
$$

7. Boundary Conditions

The following section will detail the lateral, top, and bottom boundary conditions used in RAMS.

7.1. Lateral boundaries

RAMS contains several options for the specific form of the lateral boundary conditions. The general form on the C-grid stagger that is used in RAMS is the basic radiative condition (Sommerfeld, 1???):

$$\frac{\partial u}{\partial t} = -(u + c)\frac{\partial u}{\partial x} \tag{64}$$

where u is the wind component normal to the boundary. The problem is then to specify the form of c, the phase velocity. Options exist in RAMS for the Orlanski (1976) scheme, the modified Klemp and Lilly (1978) scheme (Durran, 1981), and the Klemp and Wilhelmson (1977) scheme.

The Orlanski (1976) computes c from:

$$c = \left(\frac{\partial u}{\partial t}\right) \Big/ \left(\frac{\partial u}{\partial x}\right) \tag{65}$$

The Klemp and Lilly (1978) scheme averages the Orlanski phase velocities in the vertical, then applies the averaged velocities to the entire vertical column. The Klemp and Wilhelmson (1978) scheme simply specifies a constant value as a typical gravity wave phase velocity (10 - 30 m/s).

The radiative boundary condition is only applied to the normal velocity components. Due to the grid stagger, the other variables are defined 1/2 Δx outside of the normal velocity components. These variables also must be specified in some manner. In RAMS, options exist for zero gradient conditions, constant inflow and/or outflow conditions, or radiative outflow conditions.

When a simulation is initialized from observational data, a form of the Davies (1978) nudging scheme is utilized. See the section on four-dimensional data assimilation for a description of the implementation of this scheme.

7.2. Vertical boundaries

For the lower boundary conditions at the surface, the surface layer fluxes provide the main exchange of information between the atmosphere and the surface. However, it is convenient for numerical reasons to define variable values at a fictitious 1/2 Δz under the surface.

For the top boundary conditions, there are several choices depending on type of simulation and preference. For the non-hydrostatic equation set, there are two choices for defining the normal velocity component to the top of the domain. Either a simple wall (w=0) or the Klemp and Durran (1983) gravity wave radiative condition is applied. The Klemp-Durran condition is derived for linear, steady-state, hydrostatic gravity waves and has been shown to work adequately in some situations, especially when RAMS is initialized in a horizontally-homogeneous mode and the flow

is forced over topography. In our experience, there is considerably less success in simulations where the flow is very non-steady (such as a simulation of deep convection). The condition is inappropriate for a non-homogeneous initialization, since the condition assumes that all vertical motion at the model top is the result of vertical gravity wave propagation.

Therefore, in conjunction with either the wall or the Klemp-Durran condition, an absorbing layer can be utilized extending from above the domain of interest up to the model top. In the horizontally-homogeneous case, the form is a simple Rayleigh friction layer where an extra term has been added to the basic prognostic equations:

$$\frac{\partial \phi}{\partial t} = -\frac{\left(\phi - \phi_0\right)}{\tau} \tag{66}$$

where ϕ represents the prognostic variables of u, v, w, and θ, ϕ_0 is the initial value of these variables, and τ is a timescale which is defined as a linear function of height, varying from a value of infinity at the bottom of the absorbing layer to a maximum value at the top of the model domain which is usually set between 60 and 300 seconds.

For non-homogeneous initializations, a similar nudging condition, which can be compared to either the Davies (1978) or the Rayleigh friction schemes, can be used as an absorbing layer at the model top. See the section on four dimensional data assimilation for a description of the implementation of this scheme.

Similar to the lateral boundary conditions, the other variables are defined 1/2 Δz above the normal velocity components due to the grid stagger. These variables are specified by a simple extrapolation.

For the hydrostatic model, the vertical direction is handled somewhat differently than with a non-hydrostatic model. RAMS contains two options for the boundary condition of the hydrostatic equation. Either the top boundary pressure is computed with the Klemp-Durran condition described above or the pressure at the ground surface is predicted in the following manner.

The hydrostatic equation is written in the nonlinear form with the Exner function.

$$\begin{aligned} \frac{\partial \pi}{\partial z} &= -\frac{g}{\theta_v}\left(1 + r_T - r_v\right) \\ \pi &\equiv C_p\left(\frac{p}{p_{00}}\right)^{R/C_p} \end{aligned} \tag{67}$$

The boundary condition for the hydrostatic equation in the model integration is the prognostic surface pressure equation that is derived from substitution of the hydrostatic equation into the fully elastic mass continuity equation.

$$\frac{\partial p_g}{\partial t} = -\frac{1}{g}\int_{z_g}^{z_T}\left(\frac{\partial \rho u}{\partial x} + \frac{\partial \rho v}{\partial y}\right)dz \tag{68}$$

where z_g is the height of the ground and z_T is the height of the model top.

The boundary condition assumes that the divergence above the model top is small compared to the divergence in the domain. This assumption is consistent with many σ_p coordinate models in use that set the top pressure of the model at 100 mb and define $\partial\sigma/\partial t = 0$ at the top.

Both forms of the top absorbing layer are available for the hydrostatic model, also.

8 Surface layer parameterization

The surface layer fluxes of heat, momentum, and water vapor into the atmosphere were computed with the scheme of Louis (1979). This scheme approximates the profile functions of Businger et al. (1971) (which need to be solved iteratively) with non-iterative analytic expressions. This scheme is used to determine the fluxes from the land surface into the atmosphere and is used for water surfaces, bare soil, and vegetated surfaces. The computed fluxes serve as the lower boundary for the sub-grid diffusion scheme for the atmosphere.

The expressions for the surface layer fluxes can be written as:

$$\begin{aligned} u_*^2 &= a^2 u^2 F_m\left(\frac{z}{z_0}, Ri_B\right) \\ u_*\theta_* &= \frac{a^2}{R} u\Delta\theta F_h\left(\frac{z}{z_0}, Ri_B\right) \\ u_* r_* &= \frac{a^2}{R} u\Delta r F_h\left(\frac{z}{z_0}, Ri_B\right) \end{aligned} \tag{69}$$

where

$$a^2 = \frac{k^2}{\left(\ln\frac{z}{z_0}\right)^2} \tag{70}$$

In the unstable case, the expression for F can be written as:

$$F = 1 - \frac{bRi_B}{1 + c\left|Ri_B\right|^{1/2}} \tag{71}$$

and for the stable case

$$F = \frac{1}{\left(1 + b'Ri_B\right)^2} \tag{72}$$

The empirical constants b and b' are set to 9.4 and 4.7, respectively, in the expressions for F_m and F_h. The constant c is given by the expression:

$$c = C^* a^2 b\left(\frac{z}{z_0}\right)^{1/2} \tag{73}$$

where C^* is set to 7.4 for F_m and 5.3 for F_h.

9. Soil and vegetation parameterizations

Each grid area in RAMS is logically divided into three different classes: water, bare soil, and vegetated surface. The surface layer parameterization requires surface temperature and moisture values for all three classes. For the water surfaces, since the majority of model simulations last a few days or less of simulation time, the temperature is usually held at a temporally constant (but spatially variable) value, while the moisture value used in the surface layer parameterization is the saturation mixing ratio defined at the surface pressure and water temperature.

For land surfaces, the surface values are provided with prognostic temperature and moisture models for both the soil and vegetation. The land surface can be bare soil or vegetated surface over a shaded soil. For bare soil, RAMS uses the multi-layer soil model described by Tremback and Kessler (1985). This scheme is a modification of the schemes described by Mahrer and Pielke (1977) and McCumber and Pielke (1981) in which the numerous iterative processes have been removed. This involved formulating prognostic equations for the soil surface temperature and water content by assuming a finite depth soil/atmosphere interface layer. The soil model normally is run with between seven and twelve layers to a depth of about .5 meters in a typical simulation.

The moisture diffusivity, hydraulic conductivity, and moisture potential are given by (Clapp and Hornberger, 1978):

$$D_\eta = \frac{-bK_f\psi_f}{\eta}\left[\frac{\eta}{\eta_f}\right]^{b+3}$$

$$K_\eta = K_f\left[\frac{\eta}{\eta_f}\right]^{2b+3} \qquad (74)$$

$$\psi = \psi_f\left[\frac{\eta_f}{\eta}\right]^{b}$$

where D_η is the moisture diffusivity, η is the soil moisture content expressed in terms of volume of water per volume of soil, K_η is the hydraulic conductivity, and ψ is the moisture potential. K_f, ψ_f, and η_f, are values for a soil at saturation. The constant b is dependent on soil textural class. The soil types and values for the constants that RAMS uses are given in Table 9-1.

Table 9-1 Soil parameters for USDA textural classes used in RAMS

Soil class	$\eta_f\,(m^3m^{-3})$	$\psi_f\,(m)$	$K_f\,(ms^{-1})$	b	$C_d\,(Jm^{-3}K^{-1})$
Sand	0.395	-.121	1.760×10^{-4}	4.05	1.465×10^{6}
Loamy sand	0.410	-.090	1.563×10^{-4}	4.38	1.407×10^{6}
Sandy loam	0.435	-.218	0.347×10^{-4}	4.90	1.344×10^{6}
Silt loam	0.485	-.786	0.072×10^{-4}	5.30	1.273×10^{6}
Loam	0.451	-.478	0.070×10^{-4}	5.39	1.214×10^{6}
Sandy clay loam	0.420	-.299	0.063×10^{-4}	7.12	1.177×10^{6}

Silty clay loam	0.477	-.356	0.017×10^{-4}	7.75	1.319×10^{6}
Clay loam	0.476	-.630	0.025×10^{-4}	8.52	1.227×10^{6}
Sandy clay	0.426	-.153	0.022×10^{-4}	10.40	1.177×10^{6}
Silty clay	0.492	-.490	0.010×10^{-4}	10.40	1.151×10^{6}
Clay	0.482	-.405	0.013×10^{-4}	11.40	1.088×10^{6}
Peat	0.863	-.356	0.080×10^{-4}	7.75	0.874×10^{6}

The thermal properties of the soil is dependent on the moisture characteristics. The volumetric specific heat and the thermal diffusivity of the soil are written as:

$$C_s = (1-\eta)C_d + \eta C_w$$
$$\lambda = \frac{e^{-[\log_{10}|\psi|+2.7]}}{C_s} \times 4.186\times10^{7}, \log_{10}|\psi| \le 5.1 \tag{75}$$
$$\lambda = \frac{0.00041}{C_s} \times 4.186\times10^{7}, \log_{10}|\psi| > 5.1$$

where C_s is the volumetric specific heat of the moist soil, λ is the thermal diffusivity ($m^2\ s^{-1}$), C_d is the specific heat of the dry soil and C_w is the specific heat of water.

Heat is diffused in the soil by:

$$\frac{\partial\theta_s}{\partial t} = \frac{\partial}{\partial z}\left[\lambda\frac{\partial\theta_s}{\partial z}\right] \tag{76}$$

where θ_s is the potential temperature of the soil. The boundary conditions for the above equation are that θ_s at the deepest soil level is held constant at the initial value or externally varied (*e.g.*, as a function of time) and the potential temperature at the ground surface, θ_g , is computed with the prognostic equation:

$$C_s\Delta z_g\frac{\partial\theta_g}{\partial t} = \alpha_g R_s\downarrow + R_l\downarrow - \sigma T_g^4 + \rho_a C_p u_*\theta_* + \rho_a C_p u_* r_* - C_s\lambda\left.\frac{\partial\theta_s}{\partial z}\right|_g \tag{77}$$

where $R_s\downarrow$ and $R_l\downarrow$ are the downward shortwave and longwave atmospheric radiative fluxes, α_g is the albedo of the soil surface, σ is the Stefan-Boltzman constant, $\rho_a C_p u_*\theta_*$ is the sensible heat flux to the atmosphere, $\rho_a C_p u_* r_*$ is the latent heat flux, and Δz_g is the depth of the top soil layer.

Moisture at all soil levels is computed also with prognostic equations. For the top soil layer, the equation is

$$\frac{\partial\eta_s}{\partial t} = \frac{\left(\rho_a u_* r_* \big/ \rho_w\right) - D_\eta\dfrac{\partial\eta}{\partial z} - K_\eta}{\Delta z_g} \tag{78}$$

where ρ_w is the density of water. The moisture diffusion in all other soil levels is done by:

$$\frac{\partial \eta_s}{\partial t} = \frac{\partial}{\partial z}\left[D_\eta \frac{\partial \eta}{\partial z} + K_\eta \right] \tag{79}$$

The mixing ratio at the ground surface is computed from Philip (1957) as:

$$r_g = e^{\left[\frac{g\psi_g}{R_v T_g}\right]} r_s\left(T_g P_g\right) \tag{80}$$

where g is gravity, R_v is the gas constant for water vapor, and r_s is the saturation mixing ratio at temperature T_g and pressure P_g. This water vapor mixing ratio can be considered to be the mixing ratio of the air within the top layer of soil. To compute an "effective" mixing ratio to be used for the surface layer parameterization, a β weighting factor is computed following as:

$$\beta = .25\left(1 - \cos\left(\min\left[1, \frac{\eta_g}{\eta_{f_c}}\pi\right]\right)\right)^2 \tag{81}$$

where η_{fc} is the soil moisture at field capacity. The effective mixing ratio is then defined as:

$$r_v = \beta r_g + (1-\beta) r_a \tag{82}$$

where r_a is the vapor mixing ratio of the air.

For the vegetated surface, the conceptual model RAMS uses is the "big leaf" approach where there is a layer of vegetation overlying a shaded soil. The shaded soil is computed in a similar manner as the bare soil, except with the following differences in the surface energy budget equation.

$$C_s \Delta z_g \frac{\partial \theta_g}{\partial t} = \tau_{veg}\alpha_g R_s \downarrow + \sigma T_{veg}^4 - \sigma T_g^4 + \rho_a C_p u_* \theta_* + \rho_a C_p u_* r_* - C_s \lambda \left.\frac{\partial \theta_s}{\partial z}\right|_g \tag{83}$$

where τ_{veg} is the shortwave transmissivity through the vegetation layer and T_{veg} is the temperature of the vegetation layer. These differences will account for the shading of the soil by the vegetation and the additional source of energy to the soil by the longwave radiation from the vegetation. Note also it is assumed that all downward atmospheric longwave radiation is intercepted by the vegetation layer.

The vegetation temperature and moisture model currently used in RAMS is loosely based on Avissar (????). A prognostic temperature equation is developed for the vegetation layer, which, just like the soil model, is a basic surface energy budget equation.

$$C_{veg}\Delta z_{veg}\frac{\partial \theta_{veg}}{\partial t} = (1-\tau_{veg})\alpha_{veg} R_s \downarrow + R_l \downarrow + \sigma T_g^4 - 2\sigma T_{veg}^4 + 2\rho_a C_p u_* \theta_* + \rho_a C_p u_* r_* \tag{84}$$

where all variables retain their previous meanings except for the *veg* subscript which refers to the values for the vegetation. Note the factor of two in the longwave emissions and the sensible heat

terms, denoting transfers of energy from both sides of the leaf surface. The latent heat term does not contain this factor since stomata usually occur only on one side of the leaf.

The effective moisture for the vegetation is based on the saturation mixing ratio at the temperature of the vegetation and a stomatal resistance function which indicates how opened or closed the vegetation's stoma are based on several atmospheric and soil properties. The expression for the effective mixing ratio is:

$$r_{veg} = \gamma r_{veg_s} + (1-\gamma) r_a \tag{85}$$

where r_{veg} is the effective vegetation mixing ratio, r_{vegs} is the saturation mixing ratio as a function of the vegetation temperature and atmospheric pressure, and γ is the stomatal resistance function given by:

$$\gamma = \tag{86}$$

Any moisture that transpirates from the vegetation must be taken from the soil. This is accomplished by defining a vertical ``root" profile (from the Biosphere-Atmosphere Transfer Scheme, BATS; Dickinson, et al., 1986) and extracting the water mass from the soil depending on the fraction of roots in each soil layer.

RAMS currently uses several other parameters defined in BATS. These include 18 vegetation classifications and other parameters defined for each class including the vegetation albedo, emissivity, roughness height, displacement, fractional coverage, and leaf area index. Note that RAMS only uses the values of these parameters from BATS, not the BATS parameterization itself.

Once the temperatures and effective moisture values for each of the surface categories are computed, a surface layer flux for each of the categories is found with the Louis (1979) scheme. The fluxes must then be averaged to provide the grid-area average flux which will be applied to the atmosphere. The grid-area averaged velocity fluxes can be written as:

$$\overline{u'w'} = -\cos\left(\frac{u}{\sqrt{u^2+v^2}}\right)\left(f_w\left(u_*^2\right)_w + f_g\left(u_*^2\right)_g + f_{veg}\left(u_*^2\right)_{veg} + f_s\left(u_*^2\right)_s\right) \tag{87}$$

$$\overline{v'w'} = -\sin\left(\frac{v}{\sqrt{u^2+v^2}}\right)\left(f_w\left(u_*^2\right)_w + f_g\left(u_*^2\right)_g + f_{veg}\left(u_*^2\right)_{veg} + f_s\left(u_*^2\right)_s\right) \tag{88}$$

and the average potential temperature and water vapor fluxes are:

$$\overline{w'\theta'} = -\left(f_w\left(u_*\theta_*\right)_w + f_g\left(u_*\theta_*\right)_g + f_{veg}LAI\left(u_*\theta_*\right)_{veg} + f_s\left(u_*\theta_*\right)_s\right) \tag{89}$$

$$\overline{w'r_v'} = -\left(f_w\left(u_*r_*\right)_w + f_g\left(u_*r_*\right)_g + f_{veg}LAI\left(u_*r_*\right)_{veg} + f_s\left(u_*r_*\right)_s\right) \tag{90}$$

where the subscripts *w*, *g*, *veg*, and *s* refer to water, bare ground, vegetation, and shaded ground, respectively, *f* is the grid fraction of each component, and *LAI* is the leaf area index which is loosely defined as the average number of leaves in a vertical column in the vegetated area. Note

that the *LAI* does not effect the momentum fluxes as the roughness length is set to account for leafier vegetation in a bulk sense. Also, $f_{veg} = f_s$ and $f_w = 1 - f_g - f_{veg}$.

10. Data analysis technique

For simulations using observational data, RAMS needs data analyses for initial conditions, large scale lateral boundary tendencies, and the four-dimensional data assimilation scheme. Various observational datasets are combined and processed with a mesoscale isentropic data analysis package (Tremback, 1990) which has been termed RAMS/ISAN (ISentropic ANalysis package). Recent improvements to ISAN have been made which greatly increase the accuracy and functionality of the data analysis.

Isentropic coordinates have many advantages over other coordinate systems when applied to data analysis. Since the synoptic- scale flow is, to a first approximation, adiabatic, an objective analysis performed on an isentropic surface will better approximate the interstation variability of the atmospheric fields. Also, isentropes tend to be "packed" in frontal areas, thus providing enhanced resolution along discontinuities. In addition, because isentropes are sloped in the vicinity of fronts, short wavelength features are transformed into longer wavelengths that can be more accurately analyzed objectively with much less smoothing than with other coordinate systems. However, there are some disadvantages to isentropic coordinates, namely that the vertical resolution decreases as the atmospheric stability decreases (i.e., in the planetary boundary layer) and that the isentropes frequently intersect the ground. Therefore, one of the main improvements to ISAN has been the inclusion of a "hybrid" vertical coordinate, a mixture of isentropic and the terrain-following, σ_z coordinates which will help to alleviate the problems that occur with isentropic coordinates near the ground.

This section will detail the current version of the package and describe the analysis procedure. We will focus on the procedure for making simulations with historical data; however, similar procedures are used when using RAMS in an operational forecast mode or when only local data are available. Also, ISAN is able to create a data analysis from any or all of the described datasets, but obviously, for initializing the atmospheric model, some upper air data is necessary.

10.1. Nested grids and data analyses

ISAN will perform a separate data analysis to any or all of the nested grids specified for the RAMS simulation. This allows the user to create a higher resolution analysis over, for instance, an intensive field site. If data resolution does not warrant a separate analysis, a nested grid may be interpolated from the next coarser grid.

10.2. Access of large scale data

The first step in the analysis procedure is usually to access the available global gridded datasets that are routinely produced by the National Meteorological Center (NMC) or the European Center for Medium-range Weather Forecasting (ECMWF) which are archived at various institutions including the National Center for Atmospheric Research (NCAR). These data are usually defined on a global 2.5° grid on a number of mandatory pressure levels. These data are accessed over the area of interest and interpolated onto the RAMS polar-stereographic grids, creating a polar-

stereographic/pressure coordinate dataset. Then, the data are interpolated vertically (linearly in p^{R/C_p} or height) to both the isentropic vertical coordinate and the terrain-following, σ_z coordinate.

10.3. Processing of rawinsondes

Once the large scale data has been processed, any available (routine, special, or bogus) rawinsondes observations may be accessed. All significant and mandatory level wind, temperature, and moisture data can be used from the rawinsonde reports. The horizontal wind components, pressure, and relative humidity are interpolated vertically (linearly in p^{R/C_p}) to the same isentropic levels or the σ_z levels as was the large scale data.

An separate objective analysis is then performed on the isentropic and σ_z datasets. The Barnes (1973) objective analysis scheme is applied to the wind, pressure, and relative humidity on the isentropes and the wind, temperature, and relative humidity on the σ_z levels. User-specified parameters control the smoothing characteristics of the objective analysis and the relative importance between the rawinsondes and the large-scale data.

10.4. Processing of surface observations

The atmospheric variables at the earth's surface are analyzed in a similar manner to the upper-air variables. Wind components, potential temperature, and relative humidity are objectively analyzed using the Barnes scheme.

10.5. Blending of analyzed datasets

At this point, there are three objectively-analyzed, gridded datasets on the polar-stereographic RAMS grids, the isentropic and σ_z upper-air datasets and the surface dataset. These data are blended in the following manner. First, a layer is chosen, usually starting at a level just above the boundary layer, extending for 1-3 km. From the surface to the bottom of this layer, the analysis will be defined completely from the σ_z data. From the top of this layer to the top of the model domain, the analysis will be defined completely from the isentropic data. In the layer, there is a simple weighted average of the isentropic and σ_z data where the weighting function is a linear function of height.

This defines the final upper-air analysis to be used for the RAMS simulations. It still remains to Blending the surface analysis into the upper-air fields still remains. There are two main conceptual problems that are of concern: 1) it is not obvious how deep the effect of the surface data should extend even in ideal situations, and 2) surface observations are not necessarily representative of all of the surface area, especially in complex terrain, where the observations are usually taken in the lower elevation areas. However, the surface observation network is usually the highest resolution, spatially and temporally, of any of the datasets that are routinely accessible, therefore, we would like to use them as best we can.

For the first problem, there are a few objective techniques that are used. If the potential temperature of the surface analysis is greater than the upper-air analysis right above the ground, a neutral layer is created which extends up to the level where the upper-air potential temperature

becomes greater than the surface. Assuming that this is a well-mixed layer, the winds and relative humidity are set to the surface analysis values through the depth of the layer. If the surface potential temperature is less than the upper-air analysis, a user-specified layer depth is set(usually 200-1000 m). At the ground, the blended analysis is set to the surface analysis, with the weighting between the surface analysis and upper-air analysis linearly changing with height through the depth of the layer.

The second problem mentioned above is addressed in the following manner. The station elevations of the surface observations are objectively-analyzed in the same way as the data fields, with the Barnes scheme and the same response functions. This creates a new "ground surface" on which we assume the observations are valid. If the model topography is higher than this new surface, the final blended analysis will be weighted toward the upper-air data if the model topography height is within the blending layer mentioned above. If the model topography is higher above the new surface than the blending layer, the final analysis will be completely defined from the upper air analysis. As an example of this technique, lets take the situation in northern Colorado where there are surface stations located at Fort Collins, CO (elevation ≈ 1524 m) and Walden, CO (elevation ≈ 2500 m) which is the next closest surface observation location, about 100 km west of Fort Collins. Assume that both locations reported a surface temperature of 26° C. A standard objective analysis procedure would give all locations between Fort Collins and Walden the same temperature of 26° C. However, the Front Range of the Rockies lies between Fort Collins and Walden with the lowest pass elevation of 3140 m and several higher peaks and ridges with elevations up to 3960 m. The technique described above will assume that the temperatures at the ridge line and for some distance on either side will be better described by the upper air analysis.

10.6. Surface characteristic initialization

Because of the lack of routine measurements, the initialization of the soil temperature and moisture cannot be done accurately. Therefore, a simple method was used that accounted for the actual variations to a first order only. It was assumed that the surface analysis of atmospheric relative humidity at 1200 UTC also described the relative humidity of the air in the top layer of soil. In addition, the temperature in the top layer of soil was assumed to be 4 K lower than the surface analysis since the model runs began at about local sunrise (1200 UTC). With these values, the soil moisture content in the top soil layer was computed using the parameterized equations from the soil model that was used as a lower boundary condition for the atmospheric model.

The profile of soil temperature was defined by assuming a linear increase of 7 K from the surface value to a depth of 20 cm, then constant to the bottom soil level at 50 cm. This profile was modified from the climatological profile given by Sellers (1965) and crudely accounts for the fact that the top soil layer is colder than the air at sunrise but the temperature increases with depth from heat storage of the previous days' radiation. The soil moisture profile was defined assuming that the top-layer moisture value doubled linearly from the surface to the bottom soil level. The soil moisture values were limited between 40% and 75% of the saturation value at the surface and the saturation value under the surface. A clay loam soil textural type was assumed for all land areas.

10.7. Topographic initialization

Terrain height has an important influence on many meteorological phenomena, and every effort is made to represent terrain effects as accurately as possible in RAMS. Certain constraints and rules must be observed, however. Most important is that since terrain height is used in defining the model vertical coordinate, compatibility of terrain heights between different model grids is required for proper grid nesting communication. This is equivalent to requiring that terrain heights between grids satisfy the averaging and boundary interpolation operators which are applied to prognostic variables (see nesting section). A second concern is that in representing terrain on a discrete model grid, terrain features of importance may be strongly filtered. A common example is a long, narrow mountain chain that acts as a barrier to transverse flow, but which loses much of its effective barrier height when averaged to a coarse model grid using a standard mass-preserving operator. A third concern is that the finest resolvable spatial scales in the model (i.e., those having a wavelength close to twice the grid spacing), are not accurately advected, and are in fact moderately filtered to keep their amplitudes low. These small scales should be absent from the model terrain, so that terrain does not directly induce such scales into the prognostic fields. This section describes how these issues are met in the process of defining model terrain heights from a raw dataset.

The transfer of terrain height information from standard datasets to model grids is normally a three-step process. The datasets usually contain terrain heights defined at regular intervals of latitude and longitude or of some other quantity such as UTM coordinate. The first step is a horizontal interpolation of these data to a temporary grid of comparable resolution to the data, and in the polar stereographic coordinates used in RAMS. Second, data are averaged from this temporary grid to a second temporary polar stereographic grid whose mesh size is coarser, commonly by a factor of 2, than the model grid. This automatically filters out the undesirable small scales from the terrain height data. Since the smallest remaining scale has a wavelength equal to twice the grid spacing of the coarse temporary grid, this would be four times the grid spacing on the model grid. In this second averaging step, both the conventional and a silhouette average are performed. The conventional average is a simple summation over all terrain heights on the finer temporary grid, and a division of this sum by the total number of fine grid cells contained in a single coarse-grid cell, to obtain the terrain height to be assigned to that coarse-grid cell. The silhouette average, on the other hand, finds the mean height of the silhouette, as viewed from the east or west, of the set of fine-grid terrain heights contained within a single coarse-grid cell, and the silhouette of the same points as viewed from the north or south, and averages the two mean silhouette heights together. This becomes the computed silhouette height for that coarse-grid cell. While the conventional average preserves total terrain volume above sea level, the silhouette average adds mass by filling in valleys. It is used to maintain the effective mean barrier height that air must rise to when crossing a topographic barrier such as a ridge. The conventional average lowers this barrier height, particularly when the grid is too coarse to resolve the barrier properly. A weighted average between the conventional and silhouette averages is taken as the final height for the coarse grid cell, where the weights are specified by the user. The third and final step in computing terrain heights for a given model grid is to interpolate the values from the temporary coarse grid to the model grid. The interpolations of the first and third step use the overlapping quadratic scheme of REFERENCE( ) (also described in isan section?).

When nested grids are used in RAMS, the terrain on all of them may be defined from standard datasets in the above manner, thus allowing the finer nested grids to have higher resolution representation of terrain than the coarser grids. When this is done, however, further processing is required to make the terrain heights compatible between different grids. The compatibility requirement is two-sided: Fine grid values corresponding to a single coarse grid cell must locally average to the coarse grid terrain height value in that cell, and the fine grid values near the lateral boundary of the fine grid must be those which would result from a bi-quadratic interpolation from

coarse grid values, according to the standard nesting procedure (see nesting section). The required processing is carried out in three stages. First, all grids are filled from datasets, as described above. Second, starting from the finest nested grids and working toward the coarsest grid, the grid nesting averaging operator is applied to obtain averages that are used to replace the coarser grid values. Finally, starting from the coarsest grid and working toward the finest grids, the nesting interpolation operator is applied to re-define terrain heights at the lateral boundaries of all nested grids.

Other surface data types include the percentage of land in a grid cell, and the surface water temperature. These are interpolated from standard datasets to the model grids in the same manner as terrain height data, except that silhouette averaging is never done. Compatibility of these data between grids is not a requirement but is desirable, so the grid nesting averaging operator is applied to replace values on coarser grids with averages from the finer grids.

11. Four-dimensional data assimilation and nudging boundary conditions

A "nudging" type of four-dimensional data assimilation (4DDA) scheme has been implemented in RAMS in which the model fields can be nudged toward observational data as a simulation progresses. Since the lateral boundary condition of Davies (1978) and the nudging top boundary condition are examples of this type of nudging also, these will be also described in this section.

The 4DDA scheme, as mentioned, is a simple nudging-type scheme where an additional term is added to the model prognostic equations. This term can be written as:

$$\frac{\partial \phi}{\partial t} = \frac{\left(\phi_{obs} - \phi_m\right)}{\tau} \tag{91}$$

where ϕ represents the prognostic variables of u, v, θ_{il}, π, and r_T. Note that vertical motion is not nudged since it is inadequately observed and would lead to divergent fields. τ is a timescale which controls the strength of the nudging term and varies in three dimensions. In RAMS, the configuration of the timescale structure can be broken into three parts, the lateral boundary, the top boundary, and the domain interior. The final timescale that is defined for any grid point, then, is the minimum of the three computed timescales or the one that will provide the maximum nudging strength.

11.1. Lateral boundaries

The lateral boundary nudging is an implementation of the Davies (1978) scheme where a number of grid points in a boundary region (of only the coarsest resolution grid in a nested grid run) are nudged toward the data analysis. This serves two purposes: 1) introducing time-varying information into the model domain, and 2) damping information propagating from the model interior toward the lateral boundary. In the x (east-west) direction, the form of the timescale structure is a parabolic function defined as:

$$\tau_{lat} = \tau_B \left[\frac{\left(x - x_I\right)^2}{\left(x_I - x_B\right)^2} \right] \tag{92}$$

where τ_{lat} is the nudging timescale for the lateral boundary regions, τ_B is the timescale specified for the actual boundary point, x_B is the x coordinate of the boundary point, and x_I is the x coordinate of the interior point where the lateral boundary timescale goes to infinity. This equation is applied only between x_B and x_I .

11.2. Top boundary

This boundary condition is patterned after the Rayleigh friction absorbing layer that has been used by many modelers (e.g., Tripoli and Cotton, 1982; Durran, 1983) in which a layer below the model

top is forced back to an initial horizontally-homogeneous state in order to damp vertically-propagating gravity waves, thus reducing wave reflections from the model top. RAMS continues to use this type of absorbing layer for horizontally-homogeneous simulations. However, when the model is run with observational data initialization, a modification of this technique is used in which the absorbing layer is nudged to a inhomogeneous observed state rather than a horizontally-homogeneous initial state.

In the nudging region near the model top, the nudging timescale is defined as a simple linear function of height:

$$\tau_{top} = \tau_T \left[\frac{\left(z - z_I\right)}{\left(z_T - z_I\right)} \right] \tag{93}$$

where τ_{top} is the nudging timescale for the top boundary region, τ_T is the timescale specified for the actual top boundary point, z_T is the height of the top boundary point, and z_I is the height of the base of the nudging layer where the timescale goes to infinity. This equation is applied only between z_T and z_I .

11.3. 4DDA nudging technique

RAMS currently uses the technique that has been termed *analysis nudging* for its 4DDA scheme where the observational data is first objectively analyzed to the model grid, then the model field is nudged to the gridded analysis. This contrasts with *observational nudging* where the model fields are nudged to the observational data only at those grid points which are in the vicinity of the observations. Both techniques have their advantages and disadvantages with the analysis nudging generally being somewhat more efficient and easier to apply. Observational nudging has the advantage of only performing the nudging where the observations warrant.

RAMS combines the advantages of both of these techniques into a general analysis nudging scheme. This is accomplished by the computation of a generalized three-dimensional weighting function, $\varepsilon(x, y, z)$, which is used as added weighting factor in the specification of a timescale. This function can be used to combine several factors, including station proximity, instrument types, observational confidence, etc. For example, consider a function that is to account for proximity to rawinsonde locations. A value of 1 can be defined at the rawinsonde locations, then objectively-analyzed to the model grid, creating a field that would vary from 1 near the rawinsonde locations to a value of 0 far away from the locations. This weighting function can then be applied to the nudging term as:

$$\frac{\partial \phi}{\partial t} \quad \varepsilon(x\ y\ z) \frac{\left(\phi \qquad \phi\ \right)}{\tau}$$

11.4. Nesting considerations

Our experience has shown that it is important to maintain a smooth transition of the nudging timescales across nested grid boundaries. Therefore, the timescales are defined on the coarsest grid and interpolated to each of the nested grids.

12. Convective parameterization

Probably the most complex problem in mesoscale, synoptic and, global modeling is the parameterization of convection. Unfortunately, the convective terms are some of the most significant forcing terms in the equations that describe the atmospheric motions, even on the larger scales. Also, the fact that there have not been many observational studies of the effects of convection on the larger scales (especially in mid-latitudes) compounds the difficulty of the problem.

Various types of schemes of convective parameterizations have been devised over the years, from rather simple convective adjustment schemes (Manabe et al., 1965; Krishnamurti et al., 1980) to schemes based on equilibrium or quasi-equilibrium assumptions (Arakawa and Schubert, 1974; Kuo, 1965, 1974) to schemes derived specifically for mesoscale models (Kreitzberg and Perkey, 1976; Fritsch and Chappell, 1980a). The adjustment schemes force the grid scale thermodynamic structure toward a moist adiabat in the presence of certain conditions (conditional instability, low level convergence, etc.). The equilibrium-type schemes assume that the convection consumes the conditional instability at the rate at which the grid scale supplies the instability, while the convection consumes the instability that is present in a grid column in the mesoscale schemes.

Several general problems with the current classes of convective parameterizations can be identified:

- Our general knowledge of atmospheric moist convection, its controlling factors, and its effects on the larger scales of motion is very incomplete, at best. This makes the parameterization of convection extremely difficult since it is not clearly known what we are parameterizing. *Intuition, rather than fact, is frequently relied upon.*
- Partly because of the lack of knowledge about convection and partly because of the wide range of forms that convection can take, the current parameterization schemes lack the generality to handle the different forms of convection. Seasonal variations, environmental variations, and even diurnal variations are not well treated. There are many more subtle distinctions which current schemes also have difficulty in handling, such as convection that is not surface based, convergence that is forced by transient gravity waves which may or may not couple with the boundary layer to produce convection, and cases where conditional instability coexists with conditional symmetric instability where the schemes usually put too much energy into the convective modes rather than the slantwise modes.
- Almost all parameterization schemes currently in use treat convection as a sub-grid scale process. As computer capabilities and grid resolutions have increased, this assumption is starting to be violated frequently. For example, it is frequently assumed in the parameterizations that the updrafts, downdrafts, and compensating subsidence are all contained within a single grid column. At less than 15 km grid spacing, it is very unlikely that the subsidence will occur in the same column as the convection. For some types of convection, even a single updraft may not be contained in a grid column.
- The temporal resolution also needs to be addressed. The parameterizations provide more or less time-integrated convective effects over a few to several hours. With current emphasis in the mesoscale meteorological community on nowcasting and four-dimensional data assimilation, a more accurate temporal evolution of the convection must be depicted.

Unfortunately, in spite of all these problems, convective parameterization must still be used in many atmospheric simulations. Computer technology is still at least 2-3 orders of magnitude away from being able to explicitly simulate the convection properly and at the same time simulate the three-dimensional mesoscale interactions of a mesoscale convective system, for example. Therefore, the remainder of this chapter will describe the convective parameterization schemes that have been implemented in RAMS.

The two most used schemes today in mesoscale modeling are the Kuo (1974)-type scheme and the Fritsch and Chappell (1980a, hereafter, FC) scheme. Modified versions of both of these schemes have been implemented into RAMS at Colorado State University (Tremback, 1990) over the past decade. Other parameterizations including the Arakawa-Schubert scheme (OTHERS??) have been implemented at various other institutions. A simplified Kuo scheme (Tremback, 1990) is the standard scheme used in RAMS and will be described in the following section.

12.1. A simplified Kuo convective parameterization

This convective parameterization is a modification of the generalized form of the Kuo (1974) parameterization described by Molinari (1985). The Kuo-type scheme is an equilibrium scheme; convection acts to consume the convective instability that is supplied by the larger scales. The terms in the thermodynamic and moisture equations due to moist convection are written as:

$$\left(\frac{\partial \theta}{\partial t}\right)_{con} = L(1-b)\pi^{-1} I \frac{Q_1}{\int_{z_g}^{z_{ct}} Q_1 dz} \tag{95}$$

$$\left(\frac{\partial r_T}{\partial t}\right)_{con} = bI \frac{Q_2}{\int_{z_g}^{z_{ct}} Q_2 dz} \tag{96}$$

The computation of each quantity in the right hand side is described below.

The quantity I is the rate at which the resolvable scale is supplying moisture to a particular grid column. This is parameterized as suggested by Molinari and Corsetti (1985) as the resolved vertical flux of water vapor through the lifting condensation level (LCL). The quantity b was defined by Kuo (1974) as a moisture partitioning parameter which determines what fraction of I is used to increase the moisture of the column. The remainder of the moisture, $1-b$, is precipitated and its latent heat warms the grid column. In this scheme, the quantity $1-b$ (which also can be interpreted as the precipitation efficiency) is computed according to the empirical function given by Fritsch and Chappell (1980a).

Q_1 and Q_2 are vertical profiles of the convective heating and moistening, respectively. For Q_1, the difference between the environmental potential temperature and a convective potential temperature profile is used. To compute this convective temperature profile, a weighted average between the updraft and downdraft profiles is performed. For the updraft, the potential temperature of the moist adiabat of the source level air lifted to its LCL is used. Downdrafts are handled in an even more approximate way. Downdrafts are defined to begin at the level of the θ_E minimum of the sounding at the same temperature as the environment. At cloud base, they are 2 K colder than the environment and at the surface the temperature deficit is 5 K. Other levels are interpolated linearly

in height. The weighting function for the downdraft relative to the updraft is also somewhat arbitrarily defined to be 1% of the updraft at their beginning level, 10% at the LCL, 20% at the level of maximum downdraft mass flux (about 800 m as described by Knupp, 1985) and 100% at the surface. This weighting function is then used to define the convective temperature profile. Below cloud base, the environmental temperature is used instead of the updraft.

For Q_2, two actual regions of moisture tendency are defined. The region below cloud base is dried at the rate I. The profile of this drying is the total water mixing ratio which forces the vertical profile of the mixing ratio toward a constant value below cloud base. The anvil region is moistened by the rate bI with the moistening profile constant from 2/3 of the height between the convective source level and the cloud top (consistent with the profile of English, 1973.).

As pointed out by Molinari (1985), there is no requirement built into the scheme to reach the moist adiabat in the limit. Therefore, a check is made on the potential temperature profile with the convective tendencies added to see if any level will exceed the moist adiabatic value after the total convective tendencies are applied. The moisture supply rate, I, is reduced if this is the case and all convective tendencies are recomputed with this new value to ensure mass and energy conservation.

The steps in the computation of the convective tendencies are then:

- Convection is activated if the grid column is convectively unstable and there is resolved upward vertical motion above a specified threshold at the LCL.
- The source level air for the convection is defined as the highest θ_E air that is less than 3 km above the ground. The LCL of the source level air is found.
- Cloud top is defined as the level above which the potential temperature of the moist adiabat becomes less than the grid temperature.
- The vertical profiles of heating and moistening are computed.
- The convective tendency terms for potential and total water are determined.

13. Radiation parameterizations

RAMS currently contains two options for the longwave and two options for the shortwave radiation parameterizations. The first options are taken from Mahrer and Pielke (1977) and are rather simple and efficient, but do not account for clouds. The second set from Chen and Cotton (1983) takes into account the amount of condensate present in the simulated atmosphere but at a larger computer cost.

13.1. Shortwave radiation scheme (without cloud effects)

The shortwave scheme described in Mahrer and Pielke (1977) is a simple scheme which accounts for the scattering by oxygen, ozone, and carbon dioxide in an empirical manner and treats absorption by water vapor, but does not treat clouds or condensate in any manner. Defining the solar flux on a horizontal surface at the top of the atmosphere as a function of time of day and year,

$$S = S_0 \cos Z \tag{97}$$

$$\cos Z = \sin\phi \sin\delta + \cos\phi \cos\delta \cos\psi \tag{98}$$

where Z is the zenith angle, S_0 is the solar constant, ϕ is the latitude, δ is the solar declination, and ψ is the hour angle. The downward flux of shortwave radiation at the earth's surface if a function of two empirical functions. The first function, given by Kondrat'yev (1969) and modified by Atwater and Brown (1974), treats oxygen, ozone, and carbon dioxide.

$$G = 0.485 + 0.515\left[1.041 - 0.16\left(\frac{0.000949\,p + 0.051}{\cos Z}\right)^{1/2}\right] \tag{99}$$

where p is the pressure in millibars. The second empirical function is from McDonald (1960) and accounts for the absorption of shortwave radiation by water vapor.

$$a_w = 0.077\left[\frac{\tau(z)}{\cos Z}\right]^{0.3} \tag{100}$$

where τ is the optical path of water vapor above a vertical level z and is given by

$$\tau(z) = \int_z^{top} \rho r_v dz \tag{101}$$

The net shortwave radiative flux at the surface is given by

$$R_s = S(1-A)\left(G - a_w\right) \tag{102}$$

where A is the albedo. The change in atmospheric temperature due to absorption by water vapor is given by

$$\left(\frac{\partial T}{\partial t}\right)_S = 0.0231\frac{S}{\rho c_p}\left[\frac{\tau(z)}{\cos Z}\right]^{-0.7}\frac{\partial \tau}{\partial z} \tag{103}$$

The downward shortwave radiative flux at the ground is further modified to account for topography. As described in Mahrer and Pielke (1977), Kondrat'yev (1969) gives the expression for the solar radiation on a slanted surface as:

$$S_{sl} = S_0 \cos i \tag{104}$$

$$\cos i = \cos\alpha\cos Z + \sin\alpha\sin Z\cos(\beta - \eta) \tag{105}$$

where cos *i* is the angle of incidence of the solar rays on the sloped surface where the slope angle α and the solar and slope azimuths, β and η are given by

$$\alpha = \tan^{-1}\left[\left(\left(\frac{\partial z_g}{\partial x}\right)^2 + \left(\frac{\partial z_g}{\partial y}\right)^2\right)^{1/2}\right] \tag{106}$$

$$\beta = \sin^{-1}\left[\frac{\cos\delta\sin\phi}{\sin Z}\right] \tag{107}$$

$$\eta = \pi/2 - \tan^{-1}\left[\left(\frac{\partial z_g}{\partial x}\right)\Big/\left(\frac{\partial z_g}{\partial y}\right)\right] \tag{108}$$

Therefore, the downward solar flux at the ground is given by

$$R_s\big|_{sl} = R_s\frac{\cos i}{\cos Z} \tag{109}$$

13.2. Longwave radiation scheme (without cloud effects)

The longwave scheme described in Mahrer and Pielke (1977) is a simple scheme which accounts for the infrared emission and absorption of water vapor and carbon dioxide, but does not treat clouds or condensate in any manner. Defining the optical path lengths for water vapor and CO_2 for each model layer,

$$\Delta\tau_v = r_v\rho\Delta z \tag{110}$$

$$\Delta\tau_c = 0.0004148239\, g\rho\Delta z \tag{111}$$

The optical paths can be summed between any layer *i* and *j* for an total optical path between any two levels.

Based on data by Kuhn (1963), Jacobs, et al. (1974) gave values for the water vapor emissivities of a layer where $\overline{\tau_v}$ is the total optical path between levels *i* and *j*:

$$\varepsilon_v(i,j)=\begin{cases}0.11288\log_{10}\left(1+12.63\overline{\tau_v}\right) & \log_{10}\overline{\tau_v}\le -4\\ 0.104\log_{10}\overline{\tau_v}+0.440 & -4<\log_{10}\overline{\tau_v}\le -3\\ 0.121\log_{10}\overline{\tau_v}+0.491 & -3<\log_{10}\overline{\tau_v}\le -1.5\\ 0.146\log_{10}\overline{\tau_v}+0.527 & -1.5<\log_{10}\overline{\tau_v}\le -1\\ 0.161\log_{10}\overline{\tau_v}+0.542 & -1<\log_{10}\overline{\tau_v}\le 0\\ 0.136\log_{10}\overline{\tau_v}+0.542 & \log_{10}\overline{\tau_v}>0\end{cases} \tag{112}$$

The expression given by Kondrat'yev (1969) for the emissivity due to CO_2 is

$$\varepsilon_c(i,j)=0.185\left[1-\exp\left(-0.03919\tau_c^{0.4}\right)\right] \tag{113}$$

Defining $\varepsilon(i,j)$ to be the sum of $\varepsilon_v(i,j)$ and $\varepsilon_c(i,j)$, the atmospheric temperature tendency due to longwave radiative flux divergence at level *N* (using Sasamori's (1972) approximation for efficiency) is:

$$\left(\frac{\partial T}{\partial t}\right)_L=-\frac{1}{\rho C_p\Delta z}\left[\left(\sigma T_N^4-\sigma T_g^4\right)\left(\varepsilon(N+1,0)-\varepsilon(N,0)\right)+\left(\sigma T_{top}^4-\sigma T_N^4\right)\left(\varepsilon(N+1,top)-\varepsilon(N,top)\right)\right] \tag{114}$$

and the downward radiative flux to the earth's surface is written as:

$$R_L=\left(\sigma T_{top}^4-\sigma T_N^4\right)\left(\varepsilon(N+1,top)-\varepsilon(N,top)\right) \tag{115}$$

13.3. Longwave radiation scheme (with cloud effects)

The longwave radiation model described by Chen and Cotton (1983) is a full solution of the radiative transfer equation using an emissivity approach. The effects of condensate are taken into account in this scheme.

13.4. Shortwave radiation scheme (with cloud effects)

The shortwave scheme described by Chen and Cotton (1983) is a 3-band scheme which, like the longwave scheme, parameterizes the effects of condensate.

14. References

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