

# The CSIRO 9-level Atmospheric General Circulation Model

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CSIRO DIVISION OF ATMOSPHERIC RESEARCH TECHNICAL PAPER No. 26



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CSIRO9 500 hPa winds (JJA, m s<sup>-1</sup>)

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## National Library of Australia Cataloguing-in-Publication Entry

The CSIRO 9-level atmospheric general circulation model

Bibliography ISBN 0 643 05250 X

 Atmospheric circulation - Mathematical models.
 McGregor, J.L. II. CSIRO Division of Atmospheric Research. (Series: CSIRO Division of Atmospheric Research Technical Paper; no. 26).

551.517

Front cover: Mean 500 hPa winds for June-August as simulated by the CSIRO9 model, showing the split jet. The arrows show the wind direction. The contours show the magnitude of the wind, with light shading used for 15-20 m s<sup>-1</sup> and dark shading above 20 m s<sup>-1</sup>.

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## 1. Introduction

Numerical models of the general circulation of the atmosphere are an important tool in climate research. They have been used to investigate the dynamical and physical processes controlling the atmosphere. The incorporation of representations of the cryosphere and oceans into such a model allows it to be used for forecasting climate anomalies and climate change. The spectral 9-level Atmospheric General Circulation Model (AGCM) described in this report has been developed at the CSIRO Division of Atmospheric Research to provide the basis for the current greenhouse research project and for future research.

The spectral method for modelling the general circulation of the atmosphere is now firmly established (Bourke 1974; McAvaney et al. 1978) and has been adopted at several major research centres including the National Center for Atmospheric Research (NCAR), the European Centre for Medium-range Weather Forecasting (ECMWF), and the Canadian Climate Centre. It provides a cost effective means of atmospheric modelling which is essential for climate research.

The original CSIRO spectral AGCM had 2 vertical levels, and was developed at the Australian Numerical Meteorology Research Centre (Gordon 1983; Gordon and Hunt 1987; Hunt and Gordon 1988, 1989). From this model a 4-vertical-level model (CSIRO4) was developed at the CSIRO Division of Atmospheric Research (Gordon and Hunt 1991; Hunt and Gordon 1991; Smith and Gordon 1992), as documented by Gordon (1993). The present 9-level model (hereafter referred to as CSIRO9) was subsequently developed from this model.

The version of the CSIRO9 model described here has been used to generate the implied ocean heat transports required to enable computation of sea temperatures by a slab ocean model rather than using prescribed temperatures. The inclusion of this simple "mixed-layer" ocean into the model allows the lower boundary condition involving the land surfaces, the polar ice caps, and now the oceans to be self-determining. This form of the CSIRO9 model is currently being used for greenhouse research and will be documented elsewhere.

This publication documents the formulation of CSIRO9 and is intended as a general guide to its contents and formulation. It does not give details of the model's computer coding as would be required by users wishing to modify the model. The description of the dynamical framework of the model has been kept to a minimum since standard spectral techniques have been used. On the other hand, the physical processes are comprehensively covered.

The physical structure of the model (horizontal and vertical resolution) is detailed in the next section, and provides an overview of the formulation used in the computer code. The subsequent sections have been given in a sequence which follows as closely as possible the steps involved in the computer code to complete a timestep. Sections 3 - 14 contain details of the methods used to implement the gives brief the model. details physical parameterizations in Section 15 of considerations used when formulating the non-linear dynamics so that energy conservation is achieved. Section 16 and 17 describe the final time integration of application of spectral horizontal the main prognostic variables, and the diffusion.

The model description is followed by a short section on the model climatology (Section 18). A range of important climatic variables produced by the model over a 10-year run is presented. The observed fields are displayed for comparison wherever possible. The overall climate produced by this version of the model is acceptable by current standards of the international climate modelling community. Areas of the model simulation which might be improved are discussed where appropriate.

A trajectory (tracer) facility has been incorporated into the model, with a view to using the CSIRO9 model to study phenomena such as the emission of debris from volcanoes, and the release of possibly harmful materials into the atmosphere (e.g. during the "Gulf" war). Details are given in Appendix D.

The model has a comprehensive physical and dynamical diagnostics package, with a convenient print out of major fields (some as global maps and some as zonal averages). Data files of monthly averages of the most important global dynamical and physical quantities are created. There is also a facility for storing data over a specific region on a daily basis. Various surface fields for a small number of individual grid-points can also be saved at each timestep.

### 2. Model structure

The CSIRO9 model performs a forward-in-time integration of the primitive equations describing the motion of the global atmosphere. The model simulates a comprehensive range of "physical" processes including radiation and precipitation which act as forcings of the dynamical equations. The model is intended for general climate simulation and thus represents full annual and diurnal cycles. The lower boundary condition for the atmosphere is determined by an interactive land surface scheme but sea-surface temperatures are prescribed in the model version described here. All other major properties such as cloud amount, snow, and sea-ice are self-determining.

The model utilizes the "flux" form of the dynamical equations (Gordon 1981) rather than the "advective" form (see for example, Bourke 1974). The flux formulation ensures that conservation of mass and energy can be readily achieved (unlike the advective formulation). This conservation is vital for an AGCM which is to be used for the multi-annual integrations required for climate investigation. Details of the derivation of the model dynamical equations are given by Gordon (1981, 1993).

## Horizontal and vertical resolution

The vertical structure of the model is given in Figure 1. Standard notation is used for variables wherever possible, and a listing of all variables is given in Appendix A. The model utilizes the sigma ( $\sigma = p/p_s$ ) coordinate in the vertical. The vertical level spacing need not be uniform. The main prognostic variables of the model are the surface pressure  $p_s$ , and the surface pressure weighted divergence (D), vorticity ( $\zeta$ ), temperature (T), and moisture (q). The divergence and vorticity have an associated velocity potential ( $\chi$ ) and stream function ( $\psi$ ). The variables are carried as spectral (complex or split real/imaginary) fields except for moisture which is a grid variable. The main prognostic variables are carried at full-levels, whilst the diagnostics of "vertical" velocity ( $\dot{\sigma}$ ) and geopotential height are essentially derived at half-levels; the full-levels are located midway between the half-levels.

The model has been coded for variable spectral (horizontal) resolution, and the most appropriate resolutions are usually based upon the number of east-west gridpoints being some power of 2. This enables an efficient usage of currently available Fast Fourier Transform routines (FFTs). The current CSIRO 9-level model is run at a spectral resolution of R21 (rhomboidal truncation at 21 waves) which utilizes an equally spaced east-west grid of 64 and a pole to equator grid of 28 unevenly spaced latitudes per hemisphere. This grid resolution is sufficient to give alias-free evaluation of quadratic terms via the grid transform method of Bourke (1974). A semi-implicit leapfrog time scheme is used (current and previous timestep values are retained) together with a Robert (Asselin) time filter. The R21 model timestep is 30 minutes.



Figure 1. Level structure of the CSIRO9 model. Variables are defined in Appendix A.

## Model flow diagram

The sequence of operations during each timestep is illustrated in Figure 2, where some subroutine names have been included as a guide for model users. A more complete list of subroutine names is given in Appendix E. The model has been coded for use with a coupled ocean model, but in this document only the atmospheric component is discussed. The atmospheric part of the combined model is controlled by the routine *CSIRO9*. This routine controls the initialization of the model (the main model constants are set in *inital* and *initax*, the two restart files are read via *filerd*, and the Gaussian latitudes and Legendre polynomials are created in *gauleg*). There then follows a sequence of subroutine calls which takes the model through repeated timesteps - the Timestep loop.

The major components involved in each timestep are as follows. From the spectral input data for the stream function and velocity potential, the spectral fields for U and V are obtained (*uvharm*). At the same time, the spectral equivalents for  $\partial U_{\rm fr}/\partial t$ ,  $\partial V_{\rm fr}/\partial t$  (which are the frictional dissipation terms from the previous timestep) are also created. These will be used to determine the frictional heating of the atmosphere. This particular part of the model physics is discussed more fully in Section 14.

The model then enters the Physics transform loop. Noting that the main prognostic variables of the model  $(\chi, \psi, T, p_s, q)$  have just been updated via the previous time integration (or the equivalent fields read from the restart file), the temperature and moisture fields in particular then need to be adjusted for the physical parameterizations of rainfall, convection and vertical mixing. There will also be implied adjustments to the momentum fields via surface drag, turbulent mixing and gravity wave drag. All of the physical parameterizations (see Sections 3 - 14) are achieved during the Physics loop which transforms spectral data to equivalent grid-point fields in order to perform these adjustments. More details of the methodology used in the Physics transform loop, and a subsequent Dynamics transform loop are given in the next sub-section entitled *Grid transforms*.

The next part of the timestep evaluates the non-linear part of the tendencies for the  $\hat{\chi}, \hat{\psi}, \hat{T}$  and  $\hat{q}$  fields. This is achieved in the Dynamics loop which, like the Physics loop, takes spectral fields and creates grid-point equivalents. The gridpoint values are used to determine multiple products on the grid, and by an inverse transform, the relevant spectral tendencies are evaluated. This is the standard spectral technique for such evaluations.

Following the Physics and Dynamics transform loops (Figure 2), the linear part of the spectral tendency equations are added to the non-linear components derived during the Dynamics loop (subroutine *linear*). These linear terms affect the vorticity and divergence equations only. In order to prevent decoupling of the time integrated solution at odd and even timesteps, a Robert time filter is used (see Section 16). This filter is applied in two parts - the first stage being during the time integration (*semii*) and a subsequent part is applied following the physical adjustments (in *assel* on the flow diagram), but before the next timestep.

The time integration of the main atmospheric prognostic variables is then performed (*semii*). The grid-point moisture field is integrated in a simple leapfrog manner. The same method is applied to the spectral vorticity equation (or the stream function equivalent). In the case of the spectral divergence, temperature, and surface pressure equations, a semi-implicit time integration method is used in order to handle gravity waves; see Gordon (1981, 1993) for details. Following the time integration, the spectral horizontal diffusion for the temperature, vorticity and divergence fields are applied in a forward implicit manner for numerical stability (see Section 17). This completes a model timestep.

Initialize model (*inital*, *initax*, *filerd*, *gauleg*) **Timestep** loop Form spectral  $U_1^m$ ,  $V_1^m$  and frictional components (*uvharm*) Physics loop over latitude pairs, poles to equator (phys) Convert relevant fields from spectral to grid form (ptog):  $\hat{\mathbf{U}}, \hat{\mathbf{V}}, \mathbf{p}_{s}, \hat{\mathbf{T}}, \frac{\partial \mathbf{U}_{fr}}{\partial t}, \frac{\partial \mathbf{V}_{fr}}{\partial t}$ Store  $\hat{U}$ ,  $\hat{V}$ ,  $p_s$  for use in Dynamics loop Compute physical parameterizations per latitude (radin which calls surfset, hsflux, radfs, surfupa, hvertmx, gwdrag, rainda, conv, cvmix, surfupb) Update the grid moisture field  $\hat{q}$  following evaporation and rain Reform the spectral temperature  $\hat{T}$ Create spectral components for horizontal moisture flux  $\nabla_{\mathbf{x}}(\mathbf{V}|\mathbf{q})$ , and apply moisture diffusion on pressure surfaces  $(\nabla^2 p_{,,} \nabla^2 q)$ - End of **Physics** loop **Dynamics loop** over latitude pairs, poles to equator (*dynm*) Convert requisite fields from spectral to grid form (*dtog*):  $\hat{D}, \ \hat{\xi}, \ \hat{T}, \ \nabla p_s, \ \nabla.(\underline{V} \ \hat{q}), \ \nabla^2 p_s, \ \nabla^2 q$ Compute non-linear flux terms on the grid (dvnmnl) Store pressure level data every 6 h (dynmst) Create spectral components of the non-linear parts of the tendency equations • End of Dynamics loop Add the linear spectral tendencies (linear) Apply the Robert (Asselin) time filter (assel) Perform the semi-implicit time integration (matset, semii) Incorporate the forward implicit horizontal diffusion (diffn) End of Timestep loop Print/save statistics (filest). Update restart file (filewr)

Figure 2. CSIRO9 model flow diagram. A few key subroutine names are given in italics.

#### Grid transforms

For each timestep, the CSIRO9 model uses two grid transforms. The first is termed the Physics transform loop during which various physical parameterizations are implemented. A second grid transform termed the Dynamics loop follows, in which the non-linear dynamical tendencies are evaluated.

The temperature field will be used as an example to demonstrate the method used in the two grid transforms. During the Physics loop, the temperature is transformed from spectral space (following time integration) into its gridded form. This is achieved in a sequential manner - first the northern-most latitude, then the southern-most latitude, and so on up to the adjacent equatorial latitudes. The temperature field is modified by the physical processes (e.g. rainfall, convection, vertical mixing), and the new spectral field is generated by an inverse transform technique. This new field is then exactly fitted to the spectral resolution of the model.

The updated form of the temperature field is now in a form suitable for application of the horizontal flux calculations (the Dynamics transform) for evaluation of the non-linear part of the temperature tendencies. This again involves a full spectral-to-grid, and subsequent grid-to-spectral transform as in the Physics loop. The use of two transform loops ensures that a spectrally-fitted temperature field is used during the calculation of advective tendencies on the grid. There is only a small computational overhead associated with the second transform loop for temperature.

The transform from complex spectral space to values on the grid is performed by an algorithm which is efficient on vector processing machines. The complex fields are first split into separate real and imaginary components. The Fourier coefficients for a northern hemisphere (NH) latitude and equivalent southern hemisphere (SH) latitude can be obtained at the same time by summing separately the odd and even components (for both real and imaginary parts) for the Legendre part of the transform. This is due to the fact that

$$\mathbf{P}_{l}^{m}\{\sin(-\phi)\} = (-1)^{l+m} \mathbf{P}_{l}^{m}\{\sin(\phi)\}$$
(2.1)

where  $P_l^m{sin(\phi)}$  is an associated Legendre polynomial of the first kind normalized to unity. The resultant odd and even sums can be either added or subtracted to give Fourier components for the NH or SH. The FFT routines then generate grid values at the latitude.

The inverse transform, whereby a spectral field is resynthesized from the data on the grid at every latitude, is essentially the reverse of the above. The same efficient odd/even, real/imaginary method is used. A substantial increase in efficiency on vector machines is achieved by using rotated indices for these fields. To clarify this, note that the spectral arrays are normally held as (l, m, k) where l, m denote the Legendre components and k is the vertical level. This is efficient for the spectral-to-grid transform where summation is first carried out over the odd/even l components (a partial cascade sum method is used here). But for the inverse transform, the grid data after using the FFT has Fourier components in the array form (m, k). So for vector efficiency reasons, the spectral resynthesis which involves the sequential addition from contributions at all latitudes is done using temporary arrays of the form (m, k, l), noting that the odd/even technique is now carried out over the *last* index.

### Moisture considerations

Whilst the main dynamical fields of vorticity, divergence, and temperature are carried as spectral fields, the moisture is held as a grid-point field. This is done since moisture is such a highly variable quantity in the horizontal, it suffers unduly from spectral fitting problems. Such problems are especially apparent in the polar regions where spectral fitting can cause some locations to have excessive rainfall.

While the gridded form of the moisture field is ideal for the calculation of physical processes, the horizontal transport of moisture needs special treatment. It is evaluated by a pseudo-spectral technique, whereby the horizontal moisture flux divergence term  $\nabla .(p_s \nabla q)$  is spectrally determined by synthesis during the Physics loop (a standard spectral procedure is used for such divergence evaluations; see e.g. Bourke 1974). The equivalent grid-point values of this tendency are then computed during the subsequent Dynamics loop.

In modelling the transport of a field with large gradients, whether by spectral methods or by grid-point methods, negative values may develop even though the field should be positive-definite. Atmospheric moisture is a difficult variable to model accurately since it has both a large vertical gradient and a large pole to equator gradient. The parameterization of sub-grid-scale horizontal mixing, as described in Section 17, helps to smooth the horizontal structure of the moisture field. Similarly, vertical mixing of moisture (Section 9) also assists. However, there are occasions when the divergence of the moisture field is such that a negative value value occur. To counter this, if the moisture drops below 2x10-6 may kg(water)/kg(air) by vertical advection, then the vertical transport is inhibited. Also, following the time integration, the global moisture field at each level is checked for the presence of negative values, and these are removed by a proportional adjustment method, whilst maintaining conservation of global mean moisture.

#### 3. Interface to the physical processes (phys)

Most of the model computation is carried out during the Physics transform loop (at R21 resolution the largest portion is concerned with radiative transfer calculations). Because of their complexity each major component of the parameterizations is described in a separate section of this report.

For each latitude row certain grid-point values are evaluated from spectral space (subroutine *ptog*). These include  $\hat{U}$ ,  $\hat{V}$ ,  $p_s$  and  $\hat{T}$ . The momentum tendencies due to the horizontal diffusion of momentum are also obtained. The mixing ratio is already available in grid form. The values of  $\hat{U}$ ,  $\hat{V}$ ,  $p_s$  are not altered by the Physics loop and are retained in grid form for the Dynamics loop.

An important physical parameterization calculates the turbulent vertical mixing of momentum, which includes the effect of surface stresses. Special consideration

is required in a spectral model, because the implied changes to the velocities would require the resynthesis of the associated spectral vorticity and divergence fields. To improve computational stability these tendencies must be either calculated implicitly (similar to the procedure the model uses for temperature and moisture), or *backward* in time. By taking the latter course, we can apply the vertical mixing tendencies as a grid-point addition to the non-linear dynamics terms, and avoid this resynthesis. It is necessary that these quantities be saved between timesteps.

#### 4. Surface characteristics (surfset)

The distribution of land and non-land R21 model grid-points is shown in Figure 3. The spectral method requires that the surface topography be spectrally fitted to a resolution of R21 for use by the model. This initial topography is derived from a  $1^{\circ} \times 1^{\circ}$  data set, area averaged to the 64 x 56 Gaussian grid, and then spectrally resynthesized to R21 resolution. A consequence of this procedure is non-zero sea elevation (due to the Gibbs phenomenon). The resultant topography is shown in Figure 4.

There are 4 types of surface. These are referenced by a mask (*imsl*) which has values 1 for Ice, 2 for Mixed-Layer Ocean (MLO), 3 for Sea, and 4 for Land. This is *not* a static mask since the model allows for the growth and decay of ice (see the description of sea grid-points below).

### Land

In the current model, all land grid-points are assumed to have constant properties, except for the occurrence of snow. No modelling of the biosphere is included. Soil temperatures and soil moisture are computed for land grid-points (see Sections 6, 7 and 8), as is snow cover. A constant roughness length  $z_0 = 0.168$  m is used in the determination of the surface fluxes as described in Section 5. Details regarding surface albedo are given in Section 11 for radiation.

### Snow

If the surface conditions are sufficiently cold, then precipitation falling on the surface is converted to snow. This snow alters the prescribed surface according to the depth of snow. The albedo of snow is reduced when the snow is melting. The maximum allowable snow depth is set at 4 m.

#### Sea grid-points

The sea surface has its temperature  $(T_s)$  interpolated daily from monthly data. There is no allowance for diurnal variation of sea surface temperature. Near the poles, the sea grid-points may be converted to mixed-layer ocean grid-points with self-computed temperatures, and then to ice grid-points.

Due to the presence of non-zero elevation for sea grid-points, the atmospheric temperature and moisture fields will tend to adjust to this elevation effect. The use of observed sea level (elevation = 0) temperatures will give rise to incorrect gradients between the surface and the first model level which are used in the calculation of surface fluxes. In order to correct for this, the SSTs are adjusted to account for the spectral elevations by use of a constant lapse rate of  $6.5^{\circ}$ C km<sup>-1</sup>. These adjusted SSTs are then used in the calculation of sensible and latent heat flux.



Figure 3. Outline of land masses on the R21 model grid.



Figure 4. Topography (m) on the R21 model grid. The contour interval is 250 m.

## Sea-ice grid-points

Sea-ice is formed if the temperature of the ocean (a mixed-layer point) falls below the freezing point of sea water. A simple thermodynamic ice model based largely on Parkinson and Washington (1979) is then used to allow for ice growth and decay. The sea-ice grid-points allow for snow cover, similar to land grid-points. The temperature at the air-surface interface  $T_s$  (either ice or snow) is computed as a result of the net flux of energy (from radiation, sensible heat flux, heat of sublimation and heat conduction through the ice) into the surface layer. Sublimation reduces the snow cover at grid-points with snow, and the ice amount at snow-free ice grid-points.

Heat conduction through the ice is proportional to the temperature difference between the surface and the underlying sea water (assumed to be at freezing point). There is also a prescribed flux of 2 W  $m^{-2}$  into the ice from the ocean below the ice which is included to represent the lateral convergence of heat transport by the ocean below the ice. These two fluxes are applied to heat the ice and thus, in addition to the sublimation at the surface, control the growth/decay of the ice thickness. For a description of how ice changes its horizontal extent, see the next subsection detailing the function of the mixed-layer ocean grid-points. Some constraints imposed on the sea-ice are:

- i) the maximum snow depth is 4 m with the excess being compressed into ice below the snow,
- ii) a maximum ice depth of 4 m is allowed.

## Mixed-layer ocean grid-points

The mixed-layer ocean (MLO) grid-points act as a buffer between the sea gridgrid-points. Note that the sea grid-points take their points and the ice temperature T<sub>s</sub> from the observed data set, whereas for ice grid-points the temperature of the sea below the sea-ice is at the freezing point of sea water (the ice/snow surface temperature is computed). For MLO grid-points, a 50 m depth is assumed and from the net energy flux at the surface the evolution of temperature for the MLO point can be obtained. However, in reality the temperature of the MLO point is not only influenced by the surface energy flux but also by the influx of heat from the surrounding sea (lateral and from below by overturning). In order that the response of the MLO grid-points be realistic (and also in part because of the diurnal forcing of the model), these effects are parameterized by a relaxation back to the observed SST for that point (with an exponential decay period of about 23 days).

As the model proceeds through an annual cycle, the MLO grid-points can reach freezing point. When this occurs, a MLO point changes to an ice point. If the equatorward point is a sea point, then this point now changes status to a MLO point. The reverse of this occurs for melting. Note that for both cases the current and the *equatorward* grid-points only change status. Since the transform loops compute a latitude at a time from each pole towards the equator the surface mask can be updated in the correct sequence.

#### 5. Surface fluxes (hsflux)

If the flux into the ground is denoted by G, the net downward short-wave flux by S, the downward long-wave flux by R, the upward sensible heat flux by  $H_0$ , the upward latent heat flux by LE, then the energy balance equation linking these quantities may be written as

$$G = S + R - \sigma T_s^4 - H_0 - LE .$$
 (5.1)

Here  $T_s$  represents the effective surface temperature (for long-wave radiation purposes) and  $\sigma$  is the Stefan-Boltzmann constant. The evaluation of S, R and  $T_s$  will be given in subsequent sections.

The surface fluxes of heat and moisture, and that of momentum are parameterized following Monin-Obukhov similarity theory. This assumes a surface layer within which the fluxes of heat and momentum are constant in the vertical. The scaling velocity  $u_*$  and temperature  $\theta_*$  are defined from the heat and momentum fluxes; these are constants applying to the whole surface layer. The fluxes can be written as

$$H_0/(\rho c_p) = \overline{\theta' w'} = u_* \theta_*$$
(5.2)

$$|\underline{\tau}_{s}|/\rho = \left\{ \left( \overline{u'w'} \right)^{2} + \left( \overline{v'w'} \right)^{2} \right\}^{1/2} = u_{*}^{2}.$$
(5.3)

In the Louis (1979) method these equations are rewritten as functions of the bulk Richardson number

$$\operatorname{Ri}_{\mathbf{b}} = g \frac{\partial \Theta}{\partial z} / \left\{ \Theta \left| \frac{\partial \mathbf{V}}{\partial z} \right|^2 \right\}$$
(5.4)

$$u_*^2 = C_{DN} |\underline{V}| F_m(z/z_0, Ri_b) |\underline{V}|$$
(5.5)

$$\mathbf{u}_* \mathbf{\theta}_* = \mathbf{C}_{\mathrm{HN}} |\underline{\mathbf{V}}| \mathbf{F}_{\mathrm{h}}(\mathbf{z}/\mathbf{z}_{\mathrm{T}}, \mathbf{R}\mathbf{i}_{\mathrm{b}}) (\mathbf{\theta}_{\mathrm{s}} - \mathbf{\theta}_{\mathrm{l}})$$
(5.6)

where  $C_{DN}$  and  $C_{HN}$  are the neutral transfer coefficients for momentum and heat respectively corresponding to height z. In this section and also in Appendix B and Section 9,  $\theta$  is a column-wise potential temperature defined to equal  $T_s$  at the surface ( $p_s$  is used rather than  $p_{1000}$ ); this provides the proper units for the sensible heat flux equation (5.2) to be compatible with the soil fluxes,

$$\theta = T \left(\frac{P_s}{p}\right)^{R/c_p} .$$
(5.7)

Note that the separate components of the momentum flux are given by

$$\rho \overline{\mathbf{u}'\mathbf{w}'} = \rho \ \mathbf{C}_{\mathrm{DN}} \ |\underline{\mathbf{V}}| \ \mathbf{F}_{\mathrm{m}} \ (\mathbf{u}_{\mathrm{s}} - \mathbf{u}_{\mathrm{1}})$$
(5.8)

$$\rho \overline{\mathbf{v}' \mathbf{w}'} = \rho \ \mathbf{C}_{\mathrm{DN}} \ \left| \underline{\mathbf{V}} \right| \ \mathbf{F}_{\mathrm{m}} \ \left( \mathbf{v}_{\mathrm{s}} - \mathbf{v}_{\mathrm{l}} \right)$$
(5.9)

where the surface velocity components u, v are taken to be zero.

The roughness lengths for heat  $(z_T)$  and momentum  $(z_0)$  are different over land, with  $z_0/z_T = 7.4 \approx e^2$ ; this corresponds to the currently accepted value of 0.4 for the von Kármán constant and follows a suggestion by J.R. Garratt (personal communication, 1991). The transfer coefficients are defined by

$$C_{DN} = k^2 / \{\ln(z/z_0)\}^2$$
(5.10)

$$C_{\rm HN} = k^2 / \{ \ln(z/z_0) \ \ln(z/z_{\rm T}) \} = k^2 / \{ \ln(z/z_0) \ (2 + \ln(z/z_0)) \}.$$
(5.11)

For the stable case the functions  $F_m$  and  $F_h$  are approximated by

$$F_{m} = (1 + b'_{m} Ri_{b})^{-2}$$
(5.12)

$$F_{\rm h} = (1 + b'_{\rm h} Ri_{\rm h})^{-2}$$
(5.13)

and for the unstable case

$$F_{m} = 1 - b_{m} \operatorname{Ri}_{h} / (1 + c_{m} |\operatorname{Ri}_{h}|^{1/2})$$
(5.14)

$$F_{\rm h} = 1 - b_{\rm h} Ri_{\rm h} / (1 + c_{\rm h} |Ri_{\rm h}|^{1/2})$$
(5.15)

where

$$c_{\rm m} = c_{\rm m}^* C_{\rm DN} b_{\rm m} (z/z_0)^{1/2}$$
 (5.16)

$$c_h = c_h^* C_{HN} b_h (z/z_T)^{1/2}$$
 (5.17)

The constants are:  $b_m = b_h = 10$ ,  $b'_m = b'_h = 5$ , with  $c^*_m = 5.0$  and  $c^*_h = 2.6$ . The values were suggested by J.R. Garratt (personal communication, 1991).

In the above equations, all vertical derivatives are evaluated between the surface and the first model level; all other variables are specified at height z, which is here taken to be the height of the first model level above the surface. In practice, there is negligible error in using a value for z calculated from the hydrostatic equation at the middle of each latitude row.

#### Surface latent heat fluxes

Similarly to the sensible heat flux expression derived above, the surface latent heat flux is written as

$$LE = L \rho \overline{q'w'} = L \rho C_{HN} |\underline{V}| F_h(z/z_T, Ri_b) (q_s - q_1)$$
(5.18)

where  $L = 2.5 \times 10^6 \text{ J kg}^{-1}$  is the latent heat of evaporation. The effective value of surface mixing ratio is denoted by  $q_s$  and parameterized as

$$q_s = \alpha \ q_{sat}(T_s) \tag{5.19}$$

where  $\alpha$  is a function of soil moisture over land (see Section 7), while  $\alpha = 1$  for water and ice surfaces.

## Surface fluxes over ice

Over ice the fluxes are determined by the above equations, with the exception that the roughness lengths  $z_0$  and  $z_T$  are both set to 0.001 m.

#### Surface fluxes over the sea

Over the sea the above equations require several modifications. The roughness length  $(z_0)$  is diagnosed from the Charnock (1955) formula

$$z_0 = 0.018 |\tau_s| / (\rho g)$$
 (5.20)

This formula is combined with (5.3) and (5.5) for  $|\underline{\tau}_s|$  and solved iteratively via 3 Newton-Raphson iterations at each sea point. The first guess is 0.01 m and a minimum value of 0.000015 m is imposed. The roughness length for heat  $z_T$  is presently set equal to  $z_0$  over the sea. The roughness lengths are used in the calculation of  $C_{DN}$  and in the unstable calculation of  $F_m$  and  $F_h$  for the momentum, heat and moisture fluxes.

Guided by experimental results (Bunker 1976; Liu et al. 1979; J.R. Garratt, personal communication), we take  $C_{\rm HN}$  over the sea to be constant rather than use (5.11); a value of 0.00085 is considered appropriate for the present height of the lowest model level.

### 6. Soil and surface temperatures (surfupa and surfupb)

There are 3 layers used for calculating soil temperatures (Figure 5). Their thicknesses were chosen to enable a reasonable representation of both the diurnal and seasonal temperature waves through the soil. The surface temperature  $(T_s)$  is taken to be the temperature of the top layer having thickness,  $h_1 = 0.03$  m. In the case of snow, the thickness is taken as 0.23 m. The value of  $T_s$  is determined by the downward ground flux (G) entering from the atmosphere above, and a downward flux ( $G_{12}$ ) going into a second soil layer. Its prognostic equation is

$$\frac{\partial T_s}{\partial t} = (G - G_{12}) / (\rho_{s1} c_{s1} h_1)$$
(6.1)

where flux G is given by (5.1),  $\rho_s$  is the soil (or snow) density 1600 (or 100) kg m<sup>-3</sup> and  $c_s$  is the soil specific heat 1000 (or 2090) J kg<sup>-1</sup> K<sup>-1</sup>. The two lowest layers are assumed to be soil for purposes of heat transfer, and have thicknesses  $h_2 = 0.255$  m and  $h_3 = 2.5$  m respectively. Their prognostic temperature equations are

$$\frac{\partial T_{s^2}}{\partial t} = (G_{12} - G_{23}) / (\rho_{s^2} c_{s^2} h_2)$$
(6.2)

$$\frac{\partial T_{s^3}}{\partial t} = G_{23} / (\rho_{s3} c_{s3} h_3) .$$
 (6.3)

The bottom of the lowest layer is assumed to be insulated. The second layer thickness was chosen such that its temperature would have a lag of about 6 h compared to the top layer. The interlayer fluxes are  $G_{12}$  and  $G_{23}$ . We give a derivation for  $G_{12}$  which allows for mixed media (although in the present model this may only arise for  $G_{12}$  in the case of snow/soil). Denoting the mid-layer depths by  $z_1$  and  $z_2$  respectively (positive upwards) and the interlayer depth and temperature by  $z_{12}$  and  $T_{s12}$  we can write two equivalent uncentred expressions

## Soil temperatures



## Soil moistures

 $(snow) = Evap \uparrow Rain, Snowmelt \downarrow Runoff \rightarrow$   $\uparrow d_1 = 12 \text{ cm} \qquad 0 < w_g < 0.36$   $d_2 = 50 \text{ cm} \qquad 0 < w_b < 0.32$ (Deardorf f force-restore soil moisture scheme)

Figure 5. Configuration of levels for soil temperatures and moistures.

$$G_{12} = \rho_1 c_{s1} K_{s1}(T_{s1} - T_{s12}) / (z_1 - z_{12})$$
  
=  $\rho_2 c_{s2} K_{s2} (T_{s12} - T_{s2}) / (z_{12} - z_2)$  (6.4)

where  $T_{s1}$  is synonymous with  $T_s$  and  $K_s$  is the soil (or snow) thermal diffusivity taken to be 0.3 x 10<sup>-6</sup> (0.1 x 10<sup>-6</sup>) m<sup>2</sup> s<sup>-1</sup>. Eliminating  $T_{s12}$  between these expressions produces

$$G_{12} = \alpha_{12}(T_{s1} - T_{s2}) / \left\{ \frac{z_{12} - z_2}{\rho_2 c_{s2} K_{s2}} + \frac{z_1 - z_{12}}{\rho_1 c_{s1} K_{s1}} \right\}.$$
(6.5)

A similar expression applies for  $G_{23}$ . The soil level assignments are:  $z_1 = -0.015$  m,  $z_{12} = -0.03$  m,  $z_2 = -0.1575$  m,  $z_{23} = -0.285$  m,  $z_3 = -1.535$  m. An enhancement factor ( $\alpha_{12}$ , of order 1) is introduced in (6.5) so that the temperature derivatives are corrected for the coarse vertical resolution, and so that they still represent the typical daily (or annual in the case of  $\alpha_{23}$ ) temperature variation with depth. The derivation of this factor follows. From standard heat conduction theory for a sinusoidally varying source having angular frequency ( $\omega$ ), one would expect the soil temperature to satisfy an expression of the form

$$T = A \exp(\gamma z) + B \tag{6.6}$$

at any given time, where A and B are arbitrary constants and  $\gamma = (0.5 \ \omega \ / \ K_s)^{0.5}$ . This expression may be integrated over a layer, for example the second layer, to give an average temperature

$$\overline{T}_{s2} = B + A \{ \exp(\gamma z_{12}) - \exp(\gamma z_{23}) \} / (\gamma h_2) .$$
(6.7)

The soil temperatures used in the above formulae (6.1) to (6.5) really represent layer averages of this form. Comparing the finite differencing of such layer averages with the analytic evaluation of the derivative of (6.6) at  $z = z_{23}$ , yields that the enhancement factor should be

$$\alpha_{23} = \gamma A \exp(\gamma z_{23}) (z_2 - z_3) / (T_{s2} - T_{s3}) .$$
 (6.8)

It is found that  $\alpha_{12} = 1.85$  (daily temperature wave) and  $\alpha_{23} = 1.45$  (annual temperature wave), when the above numerical values for layer depths and soil thermal diffusivity are substituted.

#### 7. Soil moisture (surfupa and surfupb)

The treatment of soil moisture follows Deardorff (1977). The total moisture content of the soil (volume of water per volume of soil) is represented by a reservoir value  $w_b$ . The range for  $w_b$  is  $0 \le w_b \le w_{bsat}$ ; the saturation value  $w_{bsat} = 0.32$  corresponds to 0.16 m of water saturating a layer of soil with thickness  $d_2 = 0.5$  m (Figure 5). The thin surface layer has thickness 0.005 m and its response is parameterized in terms of the depth of penetration of the diurnal moisture cycle,  $d_1 = 0.12$  m. The thin layer has moisture content  $w_g$ , which represents the ground wetness for use in evaporation calculations; its saturation value is  $w_{gsat} = 0.36$ .

As given by (5.18) and (5.19), the evaporation is parameterized using the so-called " $\alpha$  method" (Kondo et al. 1990) as

$$E = \rho C_{HN} |\underline{V}| F_h(z/z_T, Ri_b) (\alpha q_{sat} - q_1)$$
(7.1)

where  $q_{sal}(T_s)$  is the saturation mixing ratio evaluated for the surface temperature  $T_s$  and  $q_1$  is the mixing ratio at the first model level. Over land the surface wetness factor  $\alpha$  is given by

$$\alpha = \min(w_g/w_{gsat}, 1.0) . \tag{7.2}$$

For snow covered surfaces,  $w_g$  is given an empirical temperature dependence as

$$w_{\rm g} = w_{\rm gsat} \{ 1 - 0.008 \ (T_{\rm frz} - T_{\rm s}) \}$$
(7.3)

where of necessity  $T_s < T_{frz} = 273.16$ .

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If (7.1) is used directly to calculate negative evaporation (i.e. dew), then excessive values may occur. At present, there is a pragmatic solution to this problem over land, obtained by setting

$$E = \max(E, 0.1 E).$$
 (7.4)

If the precipitation rate is written as  $P_r$ , then  $w_g$  is parameterized by a forcerestore procedure, whilst  $w_b$  satisfies a continuity equation

$$\frac{\partial w_g}{\partial t} = -\frac{C_1 (E - P_r)}{\rho_w d_1} - \frac{C_2 (w_g - w_b)}{\tau_w}$$
(7.5)

$$\frac{\partial w_{\rm b}}{\partial t} = -\frac{(\rm E-P_{\rm r})}{\rho_{\rm w} d_2} \,. \tag{7.6}$$

Here  $\rho_w=1000~kg~m^{-3}$  is the density of liquid water and the force-restore time scale  $\tau_w=86400~s$  (i.e. 1 day). The empirically ascribed expressions for  $C_1$  and  $C_2$  are

$$C_1 = \min \{ 14.0, \max [ 0.5, 14.0 - 22.5 ( w_g/w_{gmax} - 0.15 ) ] \}$$
  
and  
 $C_2 = 0.9$ . (7.8)

If the field capacity is exceeded due to rainfall or snow melt  $(w_b > w_{bsa})$  runoff occurs. Runoff is also allowed to occur if the rainfall rate at any given timestep exceeds an equivalent rate of 15 mm day<sup>-1</sup>.

### 8. Time integration of surface properties (surfupa and surfupb)

In early integrations with the model, an explicit procedure was used for the soil moisture calculations. However evaporation from the thin surface layer exhibited great sensitivity to  $T_s$ , and sometimes this created extreme oscillatory feedbacks in the calculation of  $T_s$  in subsequent timesteps. The solution we have adopted is to implicitly couple together the calculations for  $T_s$  and  $w_g$ . These variables and the other subsoil temperatures and  $w_b$  are all computed as two-time-level prognostic variables.

In achieving this coupled solution, some of the non-linear terms must still be handled explicitly using current timestep  $(\tau)$  values, since only linear or linearized terms can be solved implicitly in terms of the next timestep  $(\tau+1)$  variables. Two assumptions are made:

- i) the Richardson numbers and effective drag coefficients remain constant over the  $(\tau, \tau+1)$  time interval at their  $\tau$  values,
- ii) atmospheric values of T, q, and the winds maintain their  $\tau$  values during the solution for  $T_s^{\tau+1}$  and  $w_s^{\tau+1}$ .

We will define

$$\Delta T_s = T_s^{\tau+1} - T_s^{\tau} \tag{8.1}$$

and

$$\Delta w = w_{g}^{\tau+1} - w_{g}^{\tau} \tag{8.2}$$

recognizing that the implicit treatment will be carried out in terms of these variables. Equation (6.1) for  $T_s$  may be written as

$$\frac{\partial T_s}{\partial t} = \gamma_1 G_* \tag{8.3}$$

where  $\gamma_1 = 1/(\rho_{s1}c_{s1}h_1)$  and

$$G_* = G - G_{12} . (8.4)$$

Writing the RHS terms involving  $T_s$  at timestep  $\tau$ +1 wherever possible, by expanding them in terms of the first derivatives of a Taylor series expansion, gives

$$\frac{\Delta T_s}{\Delta t} = \gamma_1 \left( G_*^{\tau} + \Delta T_s \frac{\partial G_*^{\tau}}{\partial T_s} + \Delta w \frac{\partial G_*^{\tau}}{\partial w} \right) .$$
(8.5)

The ground flux, G, can be substituted from (5.1) and its derivatives written as

$$\frac{\partial G_*}{\partial T_s} = -4 \sigma T_s^3 - \frac{\partial H_0}{\partial T_s} - L \frac{\partial E}{\partial T_s} - \frac{\partial G_{12}^T}{\partial T_s}$$
(8.6)

and

$$\frac{\partial G_{\star}}{\partial w} = -L \frac{\partial E}{\partial w}$$
(8.7)

being evaluated at timestep  $\tau$ . The sensible heat flux derivative can be substituted from (5.2) and (5.6) as

$$\frac{\partial H_0}{\partial T_s} = \rho c_p C_{HN} |\underline{u}| F_h$$
(8.8)

and the evaporation derivatives from (5.18) and (5.19) as

$$\frac{\partial \mathbf{E}}{\partial \mathbf{T}_{s}} = \rho \ \mathbf{C}_{\mathrm{HN}} \ |\underline{\mathbf{u}}| \ \mathbf{F}_{\mathrm{h}} \ \alpha \ \frac{\partial \mathbf{q}_{\mathrm{sat}}}{\partial \mathbf{T}_{\mathrm{s}}}$$
(8.9)

and

$$\frac{\partial \mathbf{E}}{\partial w} = \rho \ \mathbf{C}_{\mathrm{HN}} \ \left| \underline{\mathbf{u}} \right| \ \mathbf{F}_{\mathrm{h}} \ \mathbf{q}_{\mathrm{sat}} \ \frac{\partial \alpha}{\partial w}$$
(8.10)

where  $\partial \alpha / \partial w = 1 / w_{gsat}$  from (7.1).

It now remains to discretize the soil moisture equation (7.5) in a similar implicit manner as

$$\Delta w/\Delta t = -C_1 \left( E^{\tau} + \Delta T_s \frac{\partial E^{\tau}}{\partial T_s} + \Delta w \frac{\partial E^{\tau}}{\partial w} - P_r \right) / (\rho_w d_1) - C_2 (w_g^{\tau} + \Delta w - w_b) / \tau . \quad (8.11)$$

This is now a system of two linear equations (8.5) and (8.8). They are solved by substituting  $\Delta T_s$  from (8.5) into (8.8), whence  $\Delta w$  follows and then the final  $\Delta T_s$ . It is possible to modify (8.8) to include a dependence of  $C_1$  on  $w_g$ , but this option is not presently implemented.

## Some coding details

In (8.5) the  $\Delta T_s$  terms are grouped with the common factor

gbot = 
$$\frac{1}{\gamma_1 \Delta t} - \frac{\partial G_*^{\tau}}{\partial T_s}$$

The T<sub>s</sub> equation is then

gbot 
$$\Delta T_s = G_s^{\tau} + \Delta w \frac{\partial G_s^{\tau}}{\partial w}$$
.

Rewriting (8.8) as

$$\Delta w = - \text{ clfact } L \left\{ E^{\mathsf{T}} + \Delta T_{s} \frac{\partial E^{\mathsf{T}}}{\partial T_{s}} + \Delta w \frac{\partial E^{\mathsf{T}}}{\partial w} - \frac{P_{r}}{L} \right\} - C_{2} \left( w_{g}^{\mathsf{T}} + \Delta w - w_{b} \right)$$

where clfact = C<sub>1</sub>  $\Delta t/(\rho_w d_1 L)$ , substitution gives

$$\Delta w = - \operatorname{clfact} L\left( E^{\tau} + \frac{\partial E^{\tau}}{\partial T_{s}} \left( G_{*} + \Delta w \frac{\partial G_{*}^{\tau}}{\partial w} \right) / \operatorname{gbot} + \Delta w \frac{\partial E^{\tau}}{\partial w} - \frac{P_{r}}{L} \right) - C_{2} \left( w_{g}^{\tau} + \Delta w - w_{b} \right)$$

## 9. Vertical mixing and shallow convection (hvertmx)

Turbulent vertical mixing in the model is parameterized in terms of stabilitydependent K theory and follows Blackadar (1962). The diffusion coefficients are expressed in the form

$$K_{\rm m} = l^2 \left| \frac{\partial v}{\partial z} \right| F_{\rm m}(Ri_{\rm b}) , \qquad (9.1)$$

with Blackadar's (1962) expression for the mixing length *l*:

$$l = kz / (1 + kz/\lambda)$$
. (9.2)

The asymptotic mixing length  $\lambda$  is an adjustable constant: we use 100 m, the same value as Louis (1979), rather than the Coriolis dependent expression given by Blackadar. The expression for the Richardson number has already been given by (5.4). Following Louis (1979) the expressions used for  $F_m$  and  $F_h$  are similar to those used in Section 5 for the surface fluxes. The only changes required are to (5.16) and (5.17) for the unstable parameters  $c_m$  and  $c_h$ , to introduce a dependence on *l* rather than  $z_0$  and  $z_T$ :

$$c_{\rm m}^{2} = (c_{\rm m}^{*} l^{2} b_{\rm m})^{2} \{ [(1 + \Delta z/z)^{1/3} - 1] / \Delta z \}^{3} / z .$$
(9.3)

The constants  $b_m$  and  $c_m^*$  keep the same values as in Section 5. A similar expression to (9.3) is used for  $c_h$ . The diffusion coefficients  $K_m$  and  $K_h$  are required at the vertical half-levels. Thus z corresponds to a half-level value and  $\Delta z$  is the

corresponding distance between the surrounding full-levels; analogously to Section 5, mid-longitude values calculated from the hydrostatic equation are used for z and  $\Delta z$  for each latitude row.

With the constants specified as in Section 5, it may be noted that  $K_m$  will equal  $K_h$  for neutral or stable stratification, but for unstable stratification  $F_m < F_h$  and so  $K_m < K_h$ .

#### Shallow convection

The shallow cumulus convection scheme of Geleyn (1987) is used. There is a minor modification to his procedure in that potential temperature is used rather than static energy. The effect is to modify the vertical mixing as described above by replacing  $\partial \theta / \partial z$  in (5.4) for Ri<sub>b</sub> by

$$\frac{\partial \theta}{\partial z} + \frac{L}{c_{p}} \min \left\{ 0, \frac{\partial q}{\partial z} - \frac{\partial q_{sat}}{\partial z} \right\}$$
(9.4)

where  $q_{sat}$  is the saturated mixing ratio. As for the usual vertical mixing calculations this provides  $K_m$  and  $K_h$  at all model half-levels between the bottom and top level. Modification of the fields at the first level due to surface fluxes is obtained by incorporation of those fluxes (in the case of T and q), as previously calculated. An alternative shallow convection scheme following the ideas of Tiedtke (1987) is also available in the model.

### Time integration of vertical mixing

At this stage model fields are available for timesteps  $\tau$ -1 and  $\tau$  and a first approximation to fields at the new timestep is available, at  $\tau$ +1<sup>\*</sup> say, but where the vertical mixing and gravity wave processes have not yet been carried out. An implicit split calculation is performed for q, T, u and v over the double (leapfrog) timestep interval. T and q are updated at this latest time interval. However, effective time tendencies are deduced for u and v; these are then used to produce time tendencies for divergence and vorticity which are combined with those from the gravity wave drag parameterization and added in later while proceeding to  $\tau$ +1 values. This procedure avoids an extra grid to spectral transform for the winds. The description here is given for an implicit split calculation over a time interval  $2\Delta t$ .

The following equations are presented for  $\theta$ , in the special form defined by (5.7). The equations for q, u and v follow in a very similar manner. Surface values are denoted by the subscript s. Note that  $u_s = v_s = 0$ . The split equations to be solved for vertical mixing are

$$\frac{\partial \theta}{\partial t} = -\frac{1}{\rho} \frac{\partial}{\partial z} \left( \rho \overline{\theta' w'} \right)$$
(9.5)

where the fluxes  $\overline{\theta'w'}$  are given in terms of the above  $K_h$  as

$$\overline{\theta'w'} = -K_{\rm h} \frac{\partial \theta}{\partial z} \,. \tag{9.6}$$

It is convenient to define the vertical discretization operator ( $\Delta$ ) as follows:

$$\Delta \phi_{k} = \phi_{k+0.5} - \phi_{k-0.5} \quad \text{for } k = 1, \ 1.5, \ 2, \ 2.5, \ \dots \tag{9.7}$$

where the k = 0.5 level is to be taken as synonymous with the surface. Note from the hydrostatic equation that

$$\frac{g\rho}{p_s} = \frac{g\sigma}{RT} = -\frac{\Delta\sigma}{\Delta z}.$$
(9.8)

Performing the vertical differencing and substituting  $\rho$  and  $\overline{\theta'w'}$  yields

$$\left(\frac{\partial\theta}{\partial t}\right)_{k} = \left( \left( K_{h} \frac{\Delta\sigma}{\Delta z} \frac{\Delta\theta}{\Delta z} \right)_{k+0.5} - \left( \frac{g\sigma}{RT} \overline{\theta'w'} \right)_{k-0.5} \right) / \Delta\sigma_{k} \quad k = 1, 2, 3,..$$
(9.9)

We choose to write the flux at the surface as

$$\overline{\theta'w'} = \varepsilon \left(\overline{\theta'w'}\right)_{s} + (1-\varepsilon) C_{h} \left| \underline{v}_{1} \right| (\theta_{s} - \theta_{1})$$
(9.10)

where either  $\varepsilon = 1$  for surface heat flux passed through from the surface flux routine (normal option for heat and moisture fluxes) or  $\varepsilon = 0$  for only the net transfer coefficient pre calculated (normal option for momentum fluxes).  $(\overline{\theta'w'})_s$  is an alternative notation for H<sub>0</sub>/( $\rho c_p$ ). Equation (9.9) may then be rewritten as

$$\left(\frac{\partial \theta}{\partial t}\right)_{k} = \left( \left(G \ \Delta \theta\right)_{k+0.5} - \left(G \ \Delta \theta\right)_{k-0.5} \right) / \Delta \sigma_{k} \qquad k = 2, 3,..$$
(9.11)

and

$$\left(\frac{\partial \theta}{\partial t}\right)_{1} = \left( \left(G \ \Delta \theta\right)_{1.5} - \left(G \ \Delta \theta\right)_{0.5} - \frac{\varepsilon g}{RT_{s}} (\overline{\theta' w'})_{s} \right) / \Delta \sigma_{1}$$
(9.12)

where

$$G_{k+0.5} = \left( K_h \frac{\Delta \sigma}{(\Delta z)^2} \right)_{k+0.5} \text{ for } k = 1, 2, 3,....$$
(9.13)

and

$$G_{0.5} = (1-\varepsilon) \frac{g}{RT_0} C_h |\underline{v}_1| . \qquad (9.14)$$

A zero flux condition is enforced at the top of the model by specifying  $K_h = 0$  there. These equations are amenable to an implicit tridiagonal solution. For this they are conveniently rewritten as

$$2\Delta t \left(\frac{\partial \theta}{\partial t}\right)_{k} = -A_{k}\theta_{k-1} + (A_{k} + C_{k})\theta_{k} - C_{k}\theta_{k+1} - \varepsilon D_{k}$$
(9.15)

where

$$A_k = -2\Delta t \frac{G_{k-0.5}}{\Delta \sigma_k}$$
 for k = 1, 2, 3,... (9.16)

$$C_k = -2\Delta t \frac{G_{k+0.5}}{\Delta \sigma_k}$$
 for  $k = 1, 2, 3,...$  (9.17)

$$D_k = 0$$
 for  $k = 2, 3,...$  (9.18)

$$D_{1} = 2\Delta t \frac{g}{RT_{s}} \frac{(\theta'w')_{s}}{\Delta \sigma_{1}} . \qquad (9.19)$$

The final equations, in a form suitable for tridiagonal implicit solution, are

$$A_{k}\theta_{k-1}^{\tau+1} + (1 - A_{k} - C_{k}) \theta_{k}^{\tau+1} + C_{k}\theta_{k+1}^{\tau+1} = \theta_{k}^{\tau+1^{*}} - \varepsilon D_{k}.$$
(9.20)

Once  $\theta^{\tau_{+1}}$  has been evaluated,  $T^{\tau_{+1}}$  is obtained from (5.7). The above time differencing may be easily replaced by a Crank-Nicholson formula, but this was found to be quite noisy in practice.

## Time integration of vertical fluxes of moisture

The equations for updating moisture are very similar to those above. The changes are that only the  $\varepsilon = 1$  option is available, and  $\theta$  is replaced by q in all equations. The surface moisture flux  $(\overline{q'w'})_s$  is just an alternative notation for  $E/\rho$ .

## **10.** Gravity wave drag (gwdrag)

The inclusion of gravity wave drag parameterization is beneficial to the climate simulations of atmospheric models (Boer et al. 1984; Palmer et al. 1986; McFarlane 1987). This forcing arises when gravity waves are excited at the surface by stable air flowing over irregular terrain; the waves propagate vertically and exert an implicit drag on the large scale flow. The current version of CSIRO9 uses the gravity wave drag formulation of Chouinard et al. (1986). This drag is dependent on the sub-grid-scale variations in surface topography, and is parameterized by means of a "launching height" ( $h_e$ ) which is defined to be twice the standard deviation of the surface heights. Following the method used in the parameterization of Palmer et al. (1986), the maximum value of this launching height is limited to 800 m in order to prevent two-grid noise near steep mountains. The standard deviations of the sub-grid-scale topography (the variations within a grid square) are shown in Figure 6. This data was kindly supplied by the U.K. Meteorological Office.

The scheme will be described only briefly. For a full explanation, refer to Chouinard et al. (1986); note that the equations in that paper have some obvious errors. The gravity wave drag is applied in stable atmospheric conditions only. The momentum flux at the surface (for gravity wave drag) is given by

$$(\tau_s)_{\text{ewd}} = -\alpha h_e \rho_s N_s V_s$$
(10.1)

where  $\alpha$  is a preset constant (0.01), the subscript s refers to surface values, and N is the Brunt-Väisälä frequency

$$N = \left(\frac{g}{\theta} \frac{d\theta}{dz}\right)^{1/2}.$$
 (10.2)





The Froude number is defined by

$$Fr = \frac{N}{U} h_e \left( \frac{\rho_s}{\rho} \frac{N_s}{N} \frac{U_s}{U} \right)^{1/2}$$
(10.3)

where U is the projection of the atmospheric velocity on the surface velocity given by

$$\mathbb{U} = (\underline{\mathbf{V}}, \underline{\mathbf{V}}_{\mathbf{s}}) / |\underline{\mathbf{V}}_{\mathbf{s}}| \quad . \tag{10.4}$$

The frictional change caused by the gravity wave drag is then given by

$$\frac{\partial \underline{\mathbf{V}}}{\partial t} = -\lambda \left( \underline{\mathbf{V}}_{s} / |\underline{\mathbf{V}}_{s}| \right) \widetilde{\mathbf{U}}^{2}$$
(10.5)

where

 $\tilde{U}^2 = U^2 \max \{1 - \delta Fr_c^2 / Fr^2, 0\}$  (10.6)

This applies in the region  $\sigma_s < \sigma < \sigma_c$  where  $\sigma_c$  is a "critical" level which is defined as the level at which the wind turning is such that the gravity waves break, and all drag is absorbed at or below that level. Now U, as defined above, is positive in the region between the reference level  $\sigma_s$  and  $\sigma_c$ , but is negative or zero above that level. Fr<sub>c</sub><sup>2</sup> is preset to 0.5. The quantity  $\delta$  is zero at the top level of the model, and unity below. The parameter  $\lambda$  (which has units m<sup>-1</sup>) is determined by requiring that

$$\int_{z_{c}}^{z_{s}} \rho \left( \frac{\partial \underline{V}}{\partial t} \right)_{gwd} dz = - (\underline{\tau}_{s})_{gwd} .$$
(10.7)

Converting to  $\sigma$  coordinates gives

$$\frac{p_s}{g} \int_{\sigma_c}^{\sigma_s} \left( \frac{\partial V}{\partial t} \right)_{gwd} d\sigma = + \left( \underline{\tau}_s \right)_{gwd}$$
(10.8)

where  $\sigma_s = 1$ . This yields

$$\frac{\sigma_s}{g} \int_{\sigma_c} -\lambda \left( \underline{V}_s / |\underline{V}_s| \right) \widetilde{U}^2 \, d\sigma = -\alpha \, h_e \, \rho_s \, N_s \, \underline{V}_s \,, \qquad (10.9)$$

whence

$$\lambda = (g/p_s) \alpha h_e \rho_s N_s |\underline{V}_s| / \int_{\sigma_c}^{\sigma_s} \widetilde{U}^2 d\sigma . \qquad (10.10)$$

In order that the above scheme be applied to a flux spectral model, the frictional terms are computed as  $\hat{F} \cos(\phi)/a$ , which for gravity wave drag becomes

$$p_{s} \frac{\cos\phi}{a} \left(\frac{\partial \underline{V}}{\partial t}\right)_{gwd} = -\left\{\frac{\lambda p_{s}}{|\underline{V}_{s}|} \frac{\underline{V}_{s}\cos\phi}{a}\right\} \widetilde{U}^{2} .$$
(10.11)

Note that the surface velocity component has the requisite spectral weighting, and that the term  $\{ \}$  is independent of height.

The method of implementing the gravity wave drag is to compute the U values, which are prescribed to be zero above the critical level. The  $\theta$  and  $d\theta/dz$  are used to obtain N and N<sub>s</sub> values. From the Froude numbers we get

$$Fr_{c}^{2}/Fr^{2} = Fr_{c}^{2} / \left(\frac{N^{2}h_{e}^{2}}{U^{2}}\frac{\rho_{s}N_{s}U_{s}}{\rho N U}\right) = \left(\frac{Fr_{c}^{2}T_{s}}{N_{s}U_{s}h_{e}^{2}}\right) \left(\frac{U^{3}\sigma^{1-\kappa}}{N \theta}\right)$$
(10.12)

where the potential temperature  $\theta$  is given by (5.7) and  $\kappa = R/c_p$ . Next the component  $\tilde{U}^2$  is computed for all levels, and then the vertical integral of this is computed by

$$\int_{\sigma_{c}}^{\sigma_{c}} \widetilde{\mathbb{U}}^{2} \, \mathrm{d}\sigma = \int_{0}^{1} \widetilde{\mathbb{U}}^{2} \, \mathrm{d}\sigma = \sum_{k}^{2} \widetilde{\mathbb{U}}^{2}_{k} \, \Delta\sigma_{k}$$
(10.13)

because  $U^2 = 0$  for  $\sigma_c > \sigma > 0$ . This integral is the denominator for  $\lambda$  in (10.10). If there is no generation of gravity wave drag (due to zero  $h_e$  value), then this integral is zero. The formula for  $\lambda$  value is defined to be zero (factor  $h_e$ ) in this case.

## Time differencing considerations

The changes to the wind fields due to gravity wave drag are applied similarly to the changes for vertical mixing. That is, new winds are obtained proceeding from timestep  $\tau+1^*$  to an actual  $\tau+1$ . This requires an implicit solution of (10.5). Using an effective time interval  $2\Delta t$ , tendencies are evaluated and actually added in

during the calculation of the  $\tau+2^*$  vorticity and divergence.

## 11. Radiation parameterization (radfs)

The amount of cloud cover is predicted by the model. The radiation code was developed at GFDL (Fels and Schwarzkopf 1975, 1981; Fels 1985; Schwarzkopf and Fels 1991) and allows for annual and diurnal cycles. CSIRO9 uses the recently released vectorized version. The spectrum of radiation is separated into solar (short) and terrestrial (long) bands which are treated independently. At each grid-point the code calculates the net radiative heating rate for each atmospheric layer, the radiative energy absorbed by the ground and various short and long-wave flux diagnostics, including those determining cloud radiative forcing. These quantities depend on the incoming solar flux at the top of the atmosphere, the atmospheric and surface temperatures, the surface albedo, the cloud layer densities and radiative properties, and the concentrations of water vapour, ozone and carbon dioxide.

The incoming solar flux at a grid-point depends on the position of the earth relative to the sun for the day, with orbital values for the year 1979 used repeatedly, and the zenith angle of the sun above the horizon. The solar constant is 1367 W m<sup>-2</sup>. Ozone concentrations are interpolated from the Dopplink (1974) climatology specified as a function of latitude, pressure and season. Carbon dioxide concentration is assumed constant at 330 ppmv. The other inputs to the radiation code, including the cloud amounts (but not cloud properties), depend on the atmospheric and surface state.

For both the long and short-wave bands the atmosphere is assumed to consist of homogeneous, plane parallel layers with interfaces at the half-levels. The radiative fluxes for both upward and downward directions perpendicular to the layers are calculated for each interface including the ground and the top of the atmosphere. The cooling rate of a layer is the net flux divergence divided by its heat capacity.

#### Short-wave code

The short-wave code is based on the Lacis and Hansen (1974) (hereafter L&H) approach. A complete calculation of short-wave radiation must consider for each wavelength both the direct solar beam and the diffuse component due to Rayleigh scattering by the air molecules and scattering by the clouds and earth surface. In the model, approximations are made so that only the perpendicular components of the diffuse radiation need be calculated. Furthermore, the short-wave spectrum is divided into nine bands within which the radiative properties are taken as uniform.

The code first calculates the optical depths for each layer and band, as described below. The transmission and reflection rates or functions for each band are then calculated, using the "adding method" described by L&H. Allowance is made for the effects of the cloud layers and the surface. The relative fluxes at each interface are then determined. These are summed, with the appropriate strengths, to give the net fluxes and hence the net heating rates and diagnostics. The heating rates for layers within a "thick" cloud are assumed to be constant.

Ozone is assumed to affect only band 1 which covers the ultraviolet (approximately 0.1  $\mu$ m to 0.4  $\mu$ m) and visible (0.4  $\mu$ m to 0.7  $\mu$ m) wavelengths; band 1 contains half the total incoming flux. A weak absorption of the band by H<sub>2</sub>O is included but none by CO<sub>2</sub>. For band 1 the approximation is made that the atmosphere acts as an absorbing layer on top of a reflecting layer, which is the topmost cloud or the ground in the case of clear skies. Since Rayleigh scattering is ignored for

the other bands the calculation of transmission functions of all bands for a cloud-free layer depends only on the optical path of the layer.

The optical path across a layer is the mass of absorbing gas in the layer magnified by a factor. For the downward path for all bands the factor above the top cloud or surface layer is given by (12) of L&H and accounts for the slant angle and the refraction of the incoming beam. The surface is assumed to be a Lambert reflector so that the upward radiation, except for band 1, is uniformly diffuse, requiring a factor of 5/3. For band 1 L&H find that the combined effects of Rayleigh scattering and reflection are best accounted for with a factor of 1.9. Below a non-zero cloud layer (regardless of cloud amount) these same factors are used for radiation in both directions.

For ozone the absorption of band 1 over a path is given as a function of the optical path by twice the value from (10) of L&H. The doubling is because (10) approximates the absorption of the total short-wave flux, whereas band 1 accounts for only half. Similarly the absorption by  $CO_2$  in each of the 8 near-infrared bands is given by twice the net absorption given by Sasamori et al. (1972). The only property varying across the infrared bands is the absorption by  $H_2O$ . For each band the absorption for  $CO_2$  is given by  $\{1 - \exp(-k_a y_p)\}$  where  $y_p$  is the optical path and  $k_a$  the constant absorption coefficient for the band. It is assumed that the absorption by two gases is given as the product of the two individual values.

The absorptivity and reflectivity of clouds depend only on cloud level and band, with the same values for the near-infrared bands. Both are taken to be proportional to cloud amount, hence the notion of "random overlap" of clouds within a grid square does not apply. The absorptivity of band 1 is assumed zero. The other values are given elsewhere.

The albedo of the earth surface  $\alpha_s$  is the same for each band. It depends on surface type:

 $\alpha_s = \alpha_{land}$  for land = 0.65 for sea-ice = 0.05 / (0.15 + cos  $\xi_a$ ) for sea.

For snow covered land:

$$\alpha_{\rm s} = \alpha_{\rm land} + (\alpha_{\rm sn} - \alpha_{\rm land}) \sqrt{0.1 \, d_{\rm snow}} \qquad \qquad {\rm if } d_{\rm snow} < 10 \, {\rm cm} \\ = \alpha_{\rm sn} \qquad \qquad {\rm if } d_{\rm snow} \ge 10 \, {\rm cm}$$

where  $\alpha_{tand}$  is from a prescribed data set (Posey and Clapp 1964),  $d_{snow}$  is the snow depth and  $\xi_a$  is the zenith angle. The snow albedo  $\alpha_{sn}$  is 0.8, or 0.5 if melting. The net surface reflectivity for band 1 is given by combining the earth surface value with the reflectivity due to Rayleigh scattering using equation (15) of L&H. This scattering depends on the solar zenith angle via (18) of L&H. Although (18) of L&H was derived for clear-sky conditions it is also used under cloud in the code.

#### Long-wave code

The atmosphere is itself a source of longwave radiation, not just an absorber and scatterer as it is for shortwave. The long wave code is therefore considerably more complicated. CSIRO9 uses the longwave radiation parameterization developed by Fels and Schwarzkopf at GFDL. The development of this code is described in several papers (Fels and Schwarzkopf 1975, 1981; Schwarzkopf and Fels 1985, 1991). CSIRO9 uses the most recent version of the code, as described in the 1991 paper. The frequency bands used by the code have changed substantially from the earlier versions (e.g. as used by Hart et al. 1990) and calculated heating rates agree better with those from line by line models. This version of the code has also been written for efficient vectorization.

The longwave radiation code covers the frequency range 0-2200 cm<sup>-1</sup> (wavelengths longer than 4.5  $\mu$ m). The processes included are absorption by the vibrational and rotational lines of water vapour, carbon dioxide and ozone, and the water vapour continuum absorption. The frequency ranges for each of these processes is shown in Table 1.

| Bands (cm <sup>-1</sup> ) | Absorber                    |
|---------------------------|-----------------------------|
| 0-400                     | H <sub>2</sub> O            |
| 400-560                   | $H_2O$ , continuum          |
| 560-800                   | $H_2O$ , $CO_2$ , continuum |
| 800-990                   | $H_2O$ , continuum          |
| 990-1070                  | $H_2^{-}O$ , continuum      |
| 1070-1200                 | continuum                   |
| 1200-2200                 | H <sub>2</sub> O            |
|                           |                             |

 
 Table 1. Bands used in the model for longwave absorption, following Schwarzkopf and Fels (1991).

Some weak absorption bands of  $O_3$  and  $CO_2$  are neglected here. For more details of the band structure used in the radiation code and of the particular approximation used (the "Simplified Exchange Approximation") see Schwarzkopf and Fels (1991). The remainder of this section describes details of the implementation particular to CSIRO9 only.

 $\rm CO_2$  is approximately uniformly distributed in the atmosphere and so detailed precomputed transmission functions can be used. Transmission coefficients for  $\rm CO_2$ concentrations of 330 and 660 ppmv of  $\rm CO_2$  were supplied by GFDL for a high resolution vertical grid. For use in the model, the high resolution coefficients were interpolated to the CSIRO9 vertical levels. These transmissions are calculated for two surface pressures and three temperature profiles. The interpolation method of Fels and Schwarzkopf (1981) allows calculation of accurate transmission coefficients for the actual temperatures and pressure of each model column. It is also possible to derive transmission coefficients for any desired concentration of  $\rm CO_2$  (Schwarzkopf and Fels 1985) via interpolation, though the standard CSIRO9 runs have used 330 ppmv.

The distributions of both ozone and water vapour vary in both space and time (though the variation of ozone is prescribed as noted earlier) and this precomputation is not possible. For these gases a random band model is used. This calculation includes the temperature variation of the absorption.

## Clouds

The radiation code allows for any number of cloud layers, and the cloud top and bottom for each layer can be specified separately. This allows thick clouds that fill more than one model layer or single layer clouds (by setting the top and bottom to be equal). However, cloud top and base must each correspond to a model level and cannot be set to an arbitrary pressure. The current CSIRO9 cloud scheme allows thick low cloud but the middle and high clouds are restricted to single-levels. Separate cloud layers are assumed to be randomly overlapped.

For single-level clouds, the cloud top and bottom temperatures are both equal to the temperature at the model level. In a multi-level cloud the heating rate is calculated from the fluxes at the cloud top and bottom and is constant through the depth of the cloud. The code allows for specification of cloud emissivity but presently this is set to 1 for all cloud types.

### Modifications to the radiation code for CSIRO9

A full radiation calculation is done every 2 hours (4 model timesteps), with the atmospheric heating rates held constant between these times. There is consequently a jump in the heating rate whenever a full radiation calculation is done, primarily due to the diurnal variation of solar radiation at the top of the atmosphere. This causes no problems in the free atmosphere but the surface energy balance is rather more sensitive. There are two additional steps taken at the surface to smooth the diurnal cycle of net radiation. The downward longwave flux at the surface is held constant over the 2 hours but the upward longwave flux (the  $T_s^4$  term) varies with the surface temperature each step. The solar radiation incident at the surface is smoothed by interpolating the variation of the solar zenith angle. This can be done analytically, assuming that the sun angle is the only quantity varying and that it has no effect on the time-integrated radiation.

The original code assumed that the model half-levels (or layer interfaces) were midway between the levels. In CSIRO9 the reverse is true and the calculation of the temperature and pressure at the half-levels was changed appropriately.

In order to obtain cloud forcing diagnostics, the longwave code was modified to do a clear sky calculation along with the usual calculation. This gives the so-called method II cloud radiative forcing (Cess and Potter 1987) at little extra cost (approximately 2%). The effect of clouds is not as easily separated in the shortwave code and so this code must be repeated with the cloud set to zero for the diagnostics. However, the cloud free calculation takes only about half the time of a typical cloudy calculation. Overall the diagnostic cloud forcing calculation adds about 20% to the cost of the radiation calculation.

## 12. Cloud prediction (radfs)

In CSIRO9, clouds are diagnosed from the current state of the atmosphere but are not an interactive part of the hydrological cycle. Clouds influence the model atmosphere only through their effect on the flux of radiation. The cloud diagnostic scheme is required to evaluate a cloud amount at each grid-point - nominally the fraction of the grid square "covered" by cloud - in each of three layers, and to determine the model levels of each layer. The low layer may have differing top and bottom levels. Cloud amounts and levels are based on relative humidity (RH), stability in the lower atmosphere and convective activity. A random overlap of the cloud layers is assumed in the calculation of the diagnostic total cloud amount.

The cloud optical properties required by the GFDL radiation code are emissivity (*emcld*), visible band reflectivity (*coca*), infrared band reflectivity (*cwca*) and infrared band absorptivity (*cwcb*), where array names are given in brackets. These are constants given in Table 2. They are derived from the GFDL code, except that the reflectivities are reduced by the factor 0.968 in order to produce a close balance, averaged over a typical year, of net radiation at the top of the atmosphere.

|                         | Low   | Middle | High  |
|-------------------------|-------|--------|-------|
|                         | cloud | cloud  | cloud |
| sigma levels            | 2 & 3 | 4 & 5  | 6 & 7 |
| RH <sub>c</sub>         | 72%   | 45%    | 77%   |
| C <sub>max</sub>        | 70%   | 53%    | 50%   |
| convective humidity     | 95%   | 92%    | 90%   |
| convective cloud amount | 54.6% | 41.7%  | 25.4% |
| visible reflectivity    | 0.638 | 0.523  | 0.203 |
| infrared reflectivity   | 0.484 | 0.445  | 0.184 |
| infrared absorptivity   | 0.30  | 0.20   | 0.04  |
| emissivity              | 1.0   | 1.0    |       |

Table 2. Parameters used in the calculation of diagnostic clouds.

The cloud scheme is based on the Rikus (1990) adaptation of Slingo's (1987) scheme for the GCM of the Bureau of Meteorology Research Centre (BMRC). Humidity-dependent cloud is determined in a way similar to the scheme used by Geleyn (1981) in an early ECMWF model. For each cloud layer a humidity-dependent cloud amount  $Cl_a$  is determined from the function:

$$Cl_a = 0$$
 for  $RH \le RH_c$  (12.1)

$$Cl_a = C_{max} RH (RH - RH_c) / (1 - RH_c) \text{ for } RH_c < RH < 1.0$$
 (12.2)

$$Cl_a = C_{max}$$
 for  $RH \ge 1.0$  (12.3)

where  $RH_c$  and  $C_{max}$  depend on the layer as in Table 2 and RH is the maximum of the two humidities on the sigma levels assigned to that layer (Table 2). The functions are illustrated in Figure 7. The sigma level of the cloud is that of this maximum humidity. For middle and high cloud, this is taken to be the upper level (5 and 7 respectively) in the case of two equal humidities or if both levels are saturated. In the case of low cloud, the top and bottom levels are different (levels 2 and 3) if the smaller humidity is less than 0.25% lower than the maximum or if both levels are saturated. The adoption of a  $C_{max}$  value less than 100% follows Saito and Baba (1988).

If convective activity occurred in the convection scheme during the previous timestep an effective minimum humidity (for determining cloud) is assigned to levels within a convective tower, starting at the level above the base. This results in set cloud amounts (Table 2) for convective cloud. In accord with the humidity-dependent cloud specification, the convective cloud can only exist at levels 3, 5 and/or 7.

The model of stability-dependent cloud is based on Slingo (1987). Provided the relative humidity at the bottom level is at least 60%, the low-cloud amount is set to the maximum (i.e.  $C_{max}$  for the low-cloud type) of the humidity-dependent (or convective) cloud amount and the stability-dependent quantity

1 - 33.33 (
$$T_2 - T_3$$
) /{  $p_s$  ( $\sigma_2 - \sigma_3$ ) } . (12.4)



Figure 7. Specification of cloud fraction as a function of relative humidity (%) for each of the three cloud layers.

Thus if the lower layers are sufficiently stable, low-cloud can be increased. Moreover, if the humidity-dependent low-cloud is less than 2%, then it is considered confined to level 1, unless the stability-dependent low-cloud is greater than 2% in which case it is confined to level 2.

## 13. Rainfall and cumulus convection (rainda and conv or hkuo)

First, any super-saturated layers are adjusted to be saturated, with the excess moisture removed as large-scale rainfall. The model atmosphere is adjusted for any dry instability. This uses the dry static energy (S) defined by

$$\mathbf{S} = \mathbf{c}_{\mathbf{p}}\mathbf{T} + \boldsymbol{\phi} \ . \tag{13.1}$$

The atmosphere is deemed dry unstable if  $\partial S/\partial z < 0$ . The geopotential ( $\phi$ ) is obtained from the hydrostatic equation as described in Section 15. The atmosphere is adjusted to just above neutral conditions. The cumulus convection scheme requires that the atmosphere be statically stable for dry air.

The CSIRO9 cumulus convection scheme models the release of precipitation and the redistribution of moisture and momentum which occurs within the sub-grid-scale cumulus towers. The scheme is a modification of the Arakawa (1972) "soft" moist adjustment scheme and it is described fully in Appendix C. The current CSIRO9 model also has an alternative Kuo scheme for penetrating convection (subroutine *hkuo*). The two schemes give similar global rainfall patterns, but the Hadley circulation was found to be somewhat too weak when the Kuo parameterization was used.

Cumulus towers are diagnosed to occur over a grid square when a layer (other than the lowest layer) is moist unstable with respect to one or more layers above. It is assumed that there is a constant convective mass flux  $(M_p)$  between the base

and top layers of the convective column. This flux causes a redistribution of heat in the column such that the moist instability at each level (as measured by the difference between the moist static energy at cloud base and the saturation value at each level) decays with an e-folding time of 1 hour. The flux also produces a redistribution of moisture in the column. Convection can only be initiated if the cloud base relative humidity is greater than 75%.

The generation of a convective mass flux allows for the inclusion of convective mixing of momentum (subroutines *conv* and *cvmix*). The mass flux is confined to the levels from cloud base (k = kb) to cloud top (k = kt), and is constant between these levels. The mixing of momentum is assumed to transfer momentum upward directly from cloud base to cloud top (the rising parcels of moist air), and to transfer momentum downward through an adjacent-level mixing process simulating the surrounding large scale descent from cloud top to cloud base. The scheme is applied as

$$\frac{\partial \underline{\mathbf{V}}_{\mathbf{k}}}{\partial t} = - M_{\mathbf{p}} \left( \underline{\mathbf{V}}_{\mathbf{k}} - \underline{\mathbf{V}}_{\mathbf{k}+1} \right) / \Delta \sigma_{\mathbf{k}}$$
(13.2)

for all convective levels except for k = kt where

$$\frac{\partial \underline{\mathbf{V}}_{kt}}{\partial t} = - \mathbf{M}_{p} \left( \underline{\mathbf{V}}_{kt} - \underline{\mathbf{V}}_{kb} \right) / \Delta \sigma_{kt} .$$
(13.3)

Momentum is conserved, whilst kinetic energy is reduced. This mixing is not applied directly to the velocity fields, but is computed as part of the stress tendencies.

### 14. Frictional heating

In order for the climate model to conserve energy, the dissipation of kinetic energy in the model must be fully accounted for. This energy loss is converted to a heating source term, and is applied as an adjustment to the temperature during the evaluation of the physical processes. The kinetic energy (KE) change is computed from

$$\frac{\partial KE}{\partial t} = u \left(\frac{\partial u}{\partial t}\right)_{fr} + v \left(\frac{\partial v}{\partial t}\right)_{fr}$$
(14.1)

where the subscript denotes the combined dissipation effects, which include horizontal diffusion, vertical mixing (including convective mixing), surface drag, and gravity wave drag.

As mentioned earlier, the model retains the time tendencies for the vorticity and divergence equations due to the inclusion of spectral horizontal diffusion (see Section 17 for horizontal diffusion details). From these we can calculate the equivalent tendencies for U and V. This is because it is straightforward in spectral models to derive the spectral components of the U and V fields from the vorticity  $\xi$  and velocity potential  $\chi$ . Thus the same method can be used to derive the diffusive tendencies of U and V from the diffusive tendencies of  $\xi$  and  $\chi$ . These spectral diffusion components are stored, and transformed during the subsequent timestep onto the "physics" grid, and then added to the other frictional stresses mentioned above. The complete frictional dissipation of energy can be calculated and added as part of the thermodynamic heating.

## 15. Non-linear dynamics and energy conservation (dynm)

Following the Physics transform loop, the Dynamics loop is used to spectrally synthesize the non-linear advection tendencies (see model flow diagram in Figure 2). The grid-point values transformed from spectral space are: the vorticity  $(\hat{\xi})$ , the divergence  $(\hat{D})$ , the temperature  $(\hat{T})$ , and the gradients of surface pressure. The horizontal diffusion of moisture (see Section 17) requires that the spectral fields of  $\nabla^2 q$  and  $\nabla^2 p_s$  (which were created during the Physics loop) be transformed onto the grid. The physically adjusted mixing ratio (q) is already available in grid form, and the values of  $\hat{U}$ ,  $\hat{\nabla}$ ,  $p_s$  have been held in grid form from the Physics loop. The standard spectral methods for evaluating the gradients of products are used.

It is essential that the principle of conservation of energy be adhered to in climate-length integrations, and it can be shown that the flux form of the spectral equations formally conserves both energy and mass. When applying these equations in the discrete form, care must be taken to ensure that total energy is conserved exactly. It will be shown that there are certain requirements on how half-level values are obtained from full-level values. It will also be shown how to formulate the energy conversion terms in the temperature and divergence equations for exact energy conservation.

In formulating the velocities at half-levels, we consider the kinetic energy equation which is formed at a level  $\sigma_k$  by multiplying the corresponding tendency equations for  $U_k$ ,  $V_k$  by these velocities. We require that the global mean, vertical integral of the discrete form of the total energy equation be conservative for frictionless, adiabatic flow. In the flux form of the model equations, the standard vertical finite difference form is given, for example, by

$$\left\{\frac{\partial(\mathbf{p}_{s}\dot{\sigma}\mathbf{U})}{\partial\sigma}\right\}_{k} = \left((\mathbf{p}_{s}\dot{\sigma}\mathbf{U})_{k=0.5} - (\mathbf{p}_{s}\dot{\sigma}\mathbf{U})_{k=0.5}\right) / \Delta\sigma_{k}$$
(15.1)

where

~

•

$$\mathbf{p}_{\mathbf{s}} \, \dot{\boldsymbol{\sigma}} = \boldsymbol{\sigma} \, [\hat{\mathbf{D}}]^1 - [\hat{\mathbf{D}}]^{\boldsymbol{\sigma}} \, . \tag{15.2}$$

It can be shown that for energy to be conserved, the *half* level values of  $U_{k\pm 0.5}$ ,  $V_{k\pm 0.5}$  must be taken as average of the adjacent full-level values regardless of the thickness of the  $\sigma$  levels.

On the other hand, the temperature and moisture equations have no such restriction on half-level values used in the  $\sigma$  vertical motion terms. These T, q half-level values are derived by  $\sigma$  level interpolation.

As mentioned above, another important area for energy conservation concerns the "energy conversion" terms between the kinetic energy equation and the thermodynamic (heat) equation. These terms are respectively

KE terms: {
$$\nabla .(\underline{V}p_s\phi) - \phi \stackrel{\wedge}{D} + RT \underline{V}.\nabla p_s$$
} (15.3)

Heat terms: -RTω/σ

$$\omega = \mathbf{p}_{s} \boldsymbol{\sigma} + \boldsymbol{\sigma} \left( \underline{\mathbf{V}} \cdot \nabla \mathbf{p}_{s} - [\hat{\mathbf{D}}]^{1} \right) . \tag{15.5}$$

(15.4)

where

Here  $\omega$  values are at full-levels whilst  $\dot{\sigma}$  are at half-levels. We have used the notation

$$[\hat{\mathbf{D}}]^{\boldsymbol{\mu}} = \int_{0}^{\boldsymbol{\mu}} \hat{\mathbf{D}} \, \mathrm{d}\boldsymbol{\sigma}$$

Performing the vertical integral part of the total energy conservation equation gives rise to the requirement that

$$\sigma_{k} = (\sigma_{k-0.5} + \sigma_{k+0.5}) / 2 \tag{15.6}$$

when using

$$\Delta \sigma_{k} = \sigma_{k-0.5} - \sigma_{k+0.5} . \tag{15.7}$$

Considering the above energy conversion components, the terms involving  $RTV.\nabla p_s$  will cancel exactly if the same form for  $RT_k$  is used for each. This is the first requirement for consistency in this part of the formulation. It is to be noted that the first term in the KE expression has a global mean of zero. This leaves the  $-\phi \hat{D}$  and the RT  $[\hat{D}]^{\sigma}/\sigma$  components. Now from the hydrostatic equation we have  $\partial \phi/\partial \sigma = -RT/\sigma$ . Thus for deriving an energetically consistent set we first replace RT  $[\hat{D}]^{\sigma}/\sigma$  by  $-[\hat{D}]^{\sigma}\partial \phi/\partial \sigma$ . Both terms now only involve  $\phi$ .

The vertical integral of the sum of these components (in the total energy equation) should yield  $-\phi_s [\hat{D}]^1$ . This, when globally averaged, is the so called mountain torque term. In order that the vertical integral of the discrete form does yield

$$-\phi_{s}\sum \Delta\sigma_{k} \hat{D}_{k} = -\phi_{s} [\hat{D}]^{1}$$
(15.8)

the following discrete forms are required:

$$\left(\frac{\partial \phi}{\partial \sigma}\right)_{k} = (\phi_{k-0.5} - \phi_{k+0.5}) / \Delta \sigma_{k}$$
(15.9)

and

$$\phi_{k} = (\phi_{k-0.5} + \phi_{k+0.5}) / 2$$
(15.10)

. . .. . . .

where  $\phi_{0.5} = \phi_s$  (the surface geopotential height).

Now the hydrostatic equation  $\partial \phi / \partial \sigma = -RT/\sigma$  is used to derive the geopotential heights from the full-level values of  $T_k$ . In the finite difference forms given above, if the geopotential heights were simply derived at half-levels by using

$$\left(\frac{\partial \phi}{\partial \sigma}\right)_{k} = (\phi_{k-0.5} - \phi_{k+0.5}) / \Delta \sigma_{k} = -RT_{k} / \sigma_{k}$$
(15.11)

and then averaged to get full-level values, energy conservation would be achieved. However, the  $\phi_k$  so derived are found to be not sufficiently accurate since (15.11) assumes temperatures are constant within each layer. A better representation of the vertical temperature profile is given by

32
$$T = \alpha + \beta \ln \sigma . \tag{15.12}$$

This expression is piecewise linear in terms of  $\ln \sigma$  between the full-levels. Near the surface, the temperature profile is extrapolated in this form from the lowest two levels.

The method by which the above energy conservation requirements are met for such a temperature profile is now given. The hydrostatic equation in the form

$$\frac{\partial \phi}{\partial \ln \sigma} = -RT = -R (\alpha + \beta \ln \sigma)$$
(15.13)

is integrated from the surface up, as a function of  $\ln \sigma$ , using the piecewise loglinear expression (15.12) to give the half-level values of  $\phi$ . For the top level, the actual integrated value of  $\phi$  is used. For all lower levels these half-level values are then averaged to derive the necessary full-level values by means of

$$\phi_{\mathbf{k}} = (\phi_{\mathbf{k}-0.5} + \phi_{\mathbf{k}+0.5}) / 2 \tag{15.14}$$

Note that these  $\phi_k$  values must be used if energy conservation is to be guaranteed for heights derived from a ln  $\sigma$  temperature profile. Whilst the heights are very similar to the full-level values that would be obtained directly by piecewise integration, in general they are not identical.

A further improvement to the geopotential height calculation is made by adding a correction for virtual temperature effects. In the energy conversion terms above involving  $\phi$ , we now replace T by T<sub>v</sub> where

$$T_v = T (0.622 + q) / (0.622 + 0.622q)$$
 (15.15)

This change to the calculation of the geopotential height has a noticeable effect on the surface pressures especially in the tropics. When this form of the geopotential is used, the temperature equation and the momentum equation have to be formulated in a manner that ensures cancellation of the energy conversion terms when deriving the total energy equation (see above).

During the Dynamics loop, some grid-point forcing terms computed during the Physics loop are added to the non-linear dynamical terms for spectral synthesis. This method avoids having additional synthesis during the Physics loop. The terms so added are the atmospheric stresses which have been computed as tendencies for the divergence and vorticity equations. For stability, these terms are backward in time, and are obtained from storage arrays by rotation of indices. The spectral fields evaluated during the Dynamics loop are the non-linear part of the time tendencies for temperature, vorticity, and divergence. The kinetic energy based term  $\hat{E}$ , as defined by Gordon (1981, 1993), is also evaluated spectrally so that  $\nabla^2 \hat{E}$  can be added as part of the linear tendencies (*linear*).

# 16. Time integration and temporal smoothing (semii and assel)

This section outlines the time integration of the main atmospheric variables. These are the spectral temperature, divergence, vorticity and surface pressure fields, and the grid-point moisture field. The other variables associated with the surface processes which require time integration have already been described in previous sections. A leapfrog time integration scheme is used for the main atmospheric fields. Thus two time levels of the main prognostic variables are retained. The divergence, temperature, and surface pressure equations are coupled linearly by gravity wave generation terms. A semi-implicit treatment of the gravity wave terms is used, to enable the model to utilize long timesteps (30 minutes at resolution R21). It is facilitated by the fact that the  $\nabla^2$  terms in the divergence equation have a simple solution when converted to spectral form (e.g.  $\nabla^2 p_s$  becomes  $-l(l+1)p_s l^m/a^2$  in spectral form). Full details of the derivation of the set of coupled equations are given by Gordon (1981, 1993). The general method for such a treatment can be found in Bourke (1974). The remaining prognostic variables of vorticity (spectral) and moisture (grid-point) do not entail a semi-implicit time algorithm.

Because of the use of a leapfrog time integration scheme, the solution will tend to become de-coupled at odd and even timesteps. To help prevent this, a weak time filter of the Robert (Asselin) type is applied to temperature, vorticity, divergence, moisture, and surface pressure. The form of this filter for a variable,  $\mu$ , is

 $\mu^{\tau'} = (1 - 2 F) \mu^{\tau} + F (\mu^{\tau+1} + \mu^{\tau-1})$ (16.1)

where  $\mu^{\tau'}$  is now the smoothed value at timestep  $\tau$ , performed after  $\tau+1$  fields have been evaluated. The value of F is set at 0.05.

In the case of the temperature and moisture fields this time filter is applied in two stages. This is due to the fact that the predicted values of moisture and temperature at timestep  $\tau+1$  are not yet adjusted for the effects of rainfall, latent heat release etc. (whereas the  $\tau$  and  $\tau-1$  components are fully adjusted). Thus a partial correction is first made to these fields following time integration by excluding the  $\tau+1$  component. This missing part is added later during the next timestep following all physical adjustments to these fields (see routine *assel* in the flow diagram in Figure 2).

## 17. Horizontal diffusion (diffn)

Horizontal diffusion is necessary in climate models to prevent an unrealistic build-up of amplitude of the highest wavenumber coefficients. It is included as a crude representation of the effects of sub-grid-scale motions. Horizontal diffusion is applied to the temperature, vorticity, divergence, and moisture fields. The surface pressure  $(p_s)$  is not diffused. The diffusion is applied as part of a split time integration scheme; it is applied as an adjustment following the main time integration (see below). For the temperature, vorticity and divergence, the diffusion is applied directly in a simplified spectral form. For the grid moisture, the relevant expression for the diffusion requires that certain terms be evaluated via a spectral transform process, and the diffusion is applied during the next timestep following the physics transform loop and at the start of the Dynamics loop.

For the temperature  $\hat{T}$ , the horizontal diffusion is represented by

$$\frac{\partial \hat{\mathbf{T}}}{\partial t} = - \mathbf{K}_{\mathrm{H}} \nabla \mathbf{v}_{\mathrm{p}} \nabla_{\mathrm{p}} \mathbf{T}$$
(17.1)

which is in a flux form which maintains conservation of heat. The gradient operator should be evaluated on constant pressure surfaces, but may be expanded as

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$$\nabla_{\rm p} T = \nabla T - \frac{\sigma}{p_{\rm s}} \frac{\partial T}{\partial \sigma} \nabla p_{\rm s}$$
(17.2)

where all horizontal derivatives without the p subscript are evaluated on  $\sigma$  surfaces. Note that  $\stackrel{A}{T} = p_s T'$  (where  $T = T_0 + T'$ , and  $T_0$  is isothermal). Equation (17.1) is then approximated by

$$\frac{\partial \hat{T}}{\partial t} = - + K_{\rm H} \left\{ p_{\rm s} \nabla^2 T - \sigma \frac{\partial T}{\partial \sigma} \nabla^2 p_{\rm s} \right\}.$$
(17.3)

We also approximate  $\ p_s \nabla^2 T$  by {  $\nabla^2(p_s T)$  -  $T \ \nabla^2 p_s$  } :

$$\frac{\partial \hat{T}}{\partial t} = - + K_{\rm H} \left\{ \nabla^2(p_{\rm s}T) - \left( T + \sigma \frac{\partial T}{\partial \sigma} \right) \nabla^2 p_{\rm s} \right\}.$$
(17.4)

For ease of spectral computation and to maintain conservation, the term  $(T + \sigma \partial T/\partial \sigma)$  is replaced by  $(\overline{T} + \sigma \partial \overline{T}/\partial \sigma)$  where  $\overline{T}$  is the global mean  $\sigma$ -level temperature, giving

$$\frac{\partial \hat{T}}{\partial t} = - + K_{\rm H} \left\{ \nabla^2 \hat{T} - \left( \overline{T} + \sigma \frac{\partial \overline{T}}{\partial \sigma} - T_0 \right) \nabla^2 p_s \right\}.$$
(17.5)

This expression for diffusion can be applied directly in spectral form because of the simple  $\nabla^2$  conversion (see previous section).

The temperature diffusion is evaluated in two stages as a split time scheme. First, the component involving  $p_s$  is added as an adjustment to the current value of  $\hat{T}$ . Secondly, the  $\nabla^2 \hat{T}$  term is applied as a forward implicit adjustment for stability. This diffusion is only applied to a selected part of the spectrum. For the CSIRO9 model this is the upper half of the rhomboid. The diffusion coefficient is  $K_{\rm H} = 10^6 \, {\rm m}^2 \, {\rm s}^{-1}$ .

The diffusion of vorticity and divergence assumes the standard form (see for example, Bourke 1974) and is applied directly to the pressure weighted stream function and velocity potential values on  $\sigma$  surfaces without attempting to correct for any  $p_s$  weighting effects. Again, the diffusion affects only the upper half of the rhomboid, and is also evaluated as an implicit forward adjustment for stability. The implied time tendencies for the stream function and velocity potential are evaluated and retained, and are used later to obtain the equivalent tendencies for the U, V components. These latter values are used during the physical adjustments as a source of frictional heating to the atmosphere.

Diffusion of moisture was incorporated into the model because it was effective in ameliorating occasional unrealistically dry columns. Because moisture is carried on the grid, the treatment is different to the spectral diffusion parameterizations above. The expression for moisture diffusion (where  $q = p_s q$ ) is given by

$$\frac{\partial \hat{\mathbf{q}}}{\partial t} = - + \mathbf{K}_{\mathrm{H}} \left\{ \mathbf{p}_{\mathrm{s}} \nabla^2 \mathbf{q} - \boldsymbol{\sigma} \frac{\partial \mathbf{q}}{\partial \boldsymbol{\sigma}} \nabla^2 \mathbf{p}_{\mathrm{s}} \right\}.$$
(17.6)

This form of moisture diffusion requires that both  $\nabla^2$  components be available in grid-point form. To achieve this, the spectral field of q is synthesized during the Physics loop ( $p_s$  is already available in spectral form). Both  $\nabla^2$  terms can then be transformed to grid-point form via the equivalent spectral formulae

- 
$$l(l+1) q_{l}^{m} / a^{2}$$
 and -  $l(l+1) p_{s_{l}}^{m} / a^{2}$ 

during the Dynamics loop. The total expression for moisture diffusion is then evaluated and the diffusion is then applied as an adjustment to the grid moisture field before the non-linear dynamical tendencies are computed.

In the case of moisture, the terms  $\nabla^2 q$ ,  $\nabla^2 p_s$  are created using the entire rhomboid. The diffusion coefficient is half that used for the other terms as that was found sufficient to moisten most of the dry columns. It is to be noted that the above form for the horizontal diffusion is non-conservative (unlike, for example, the spectral temperature diffusion), so the global mean change implied at each level is corrected each timestep.

### **18. Model Climatology**

A 10-year run of the model described in the previous sections was performed during 1991. The SSTs were prescribed to follow values linearly interpolated between mean values for the start of each month from the MIT-UKMO climatology for 1951-80 (Bottomley et al. 1990). The seasonal climatology from this simulation is presented here as a brief overview, and also to allow comparison with future versions of the model. The coverage is limited, as it is anticipated that the model performance, particularly for the Australian region, will be analysed in more detail in later papers. A brief comparison with observations available to the authors and published results from other models is given. The accuracy of the observations is uncertain, so statistical significance of the apparent errors is difficult to assess. All the model data used here were calculated from monthly mean fields saved during the run. The dynamical quantities u, v, T and q were interpolated to preset pressure levels each 6 model hours during the run and the monthly means were calculated from those fields. The pressure values chosen were the sigma levels times 1000 hPa. Values at pressures beneath the surface were left undefined. For the other fields the monthly means were evaluated from values determined at each timestep through the run.

## Global means

Some global, annual mean quantities are given in Tables 3 and 4, together with values for each month of the year. The names given to the quantities are those on model outputs. Where standard symbols are available they are mentioned in the text. The modelled global means exhibit an annual cycle as a result of several effects. The changing earth to sun distance produces a variation in the mean incoming solar radiation (from 353 W m<sup>-2</sup> in January to 331 W m<sup>-2</sup> in July). Much of this change is offset by the cycle in reflected short wave (SW) radiation (*solrf* in Table 3). The imposed SSTs ( $T_{sea}$ ) are an important determinant of the annual cycle. The mean SST peaks in May-June. The variation in the imposed ozone distribution also contributes to the annual cycle. The net flux through the top of the atmosphere (*asbal*) equals the net solar flux minus the outgoing long wave (LW) radiation (*rt*).

| month | T <sub>atm</sub> | T <sub>surf</sub> | T <sub>land</sub> | T <sub>sea</sub> | TC | LC | MC | HC | rt                | solrf             | CF <sub>LW</sub> | CF <sub>sw</sub>  |
|-------|------------------|-------------------|-------------------|------------------|----|----|----|----|-------------------|-------------------|------------------|-------------------|
|       | К                | К                 | K                 | К                | %  | %  | %  | %  | W m <sup>-2</sup> | W m <sup>-2</sup> | W m-2            | W m <sup>-2</sup> |
| 1     | 248.9            | 286.0             | 276.5             | 290.0            | 55 | 34 | 25 | 17 | 234.5             | 111.3             | 29.9             | -51.1             |
| 2     | 248.8            | 286.1             | 276.7             | 290.0            | 55 | 34 | 25 | 18 | 234.4             | 109.3             | 29.9             | -51.0             |
| 3     | 248.9            | 286.4             | 277.9             | 290.0            | 55 | 34 | 25 | 18 | 234.3             | 108.4             | 30.3             | -49.4             |
| 4     | 249.4            | 287.3             | 280.5             | 290.2            | 55 | 34 | 25 | 18 | 235.0             | 107.8             | 31.3             | -47.4             |
| 5     | 250.1            | 288.6             | 284.3             | 290.4            | 54 | 32 | 25 | 18 | 237.1             | 105.0             | 31.9             | -46.4             |
| 6     | 250.9            | 289.7             | 288.2             | 290.4            | 53 | 31 | 25 | 18 | 240.3             | 99.3              | 31.6             | -46.3             |
| 7     | 251.2            | 290.3             | 290.3             | 290.3            | 53 | 30 | 25 | 18 | 241.6             | 96.7              | 31.3             | -46.6             |
| 8     | 250.7            | 290.1             | 289.6             | 290.3            | 53 | 30 | 25 | 18 | 240.9             | 96.8              | 31.4             | -48.5             |
| 9     | 249.8            | 289.2             | 286.6             | 290.3            | 53 | 31 | 25 | 18 | 238.7             | 99.5              | 31.4             | -50.0             |
| 10    | 249.2            | 288.1             | 283.1             | 290.3            | 54 | 32 | 25 | 18 | 236.8             | 104.4             | 31.1             | -49.7             |
| 11    | 248.9            | 287.1             | 279.8             | 290.2            | 55 | 34 | 26 | 18 | 235.3             | 110.0             | 30.8             | -49.5             |
| 12    | 248.9            | 286.4             | 277.5             | 290.1            | 55 | 34 | 25 | 17 | 234.7             | 112.3             | 30.3             | -50.1             |
| year  | 249.7            | 288.0             | 282.6             | 290.2            | 54 | 33 | 25 | 18 | 237.0             | 105.1             | 30.9             | -48.8             |

| month | asbal             | atbal             | sfbal             | ocbal             | hflux             | atwt | snw  | sic  | evap | rain | runf | gwet |
|-------|-------------------|-------------------|-------------------|-------------------|-------------------|------|------|------|------|------|------|------|
|       | W m <sup>-2</sup> | mm   | cm   | cm   | mm/d | mm/d | mm   |      |
| 1     | 7.4               | -0.4              | 8.4               | 10.4              | 16.9              | 22.1 | 20.9 | 10.2 | 2.9  | 2.9  | 10.7 | 0.68 |
| 2     | 6.7               | -0.6              | 8.2               | 10.2              | 17.3              | 22.3 | 23.8 | 10.4 | 2.9  | 2.9  | 10.1 | 0.68 |
| 3     | 2.8               | 0.0               | 3.9               | 3.6               | 17.7              | 22.5 | 26.1 | 11.0 | 3.0  | 2.9  | 12.7 | 0.69 |
| 4     | -3.2              | 1.5               | -3.9              | -8.7              | 18.0              | 23.1 | 25.7 | 11.5 | 3.0  | 3.0  | 14.0 | 0.69 |
| 5     | -7.7              | 2.5               | -9.5              | -18.9             | 19.2              | 23.8 | 18.5 | 11.7 | 3.1  | 3.1  | 20.1 | 0.69 |
| 6     | -8.2              | 2.7               | -10.3             | -21.5             | 21.6              | 24.8 | 16.0 | 11.3 | 3.1  | 3.1  | 12.7 | 0.66 |
| 7     | -7.3              | 0.8               | -6.7              | -17.0             | 23.0              | 25.4 | 17.3 | 10.1 | 3.1  | 3.0  | 11.1 | 0.62 |
| 8     | -4.2              | -2.3              | 0.3               | -4.4              | 22.1              | 24.8 | 19.1 | 9.2  | 2.9  | 2.9  | 11.0 | 0.60 |
| 9     | 0.0               | -3.9              | 5.9               | 7.2               | 19.3              | 23.4 | 20.8 | 9.3  | 2.8  | 2.9  | 10.5 | 0.60 |
| 10    | 2.8               | -2.9              | 6.8               | 10.3              | 17.3              | 22.4 | 22.2 | 9.6  | 2.9  | 2.9  | 9.9  | 0.62 |
| 11    | 4.1               | -1.5              | 6.3               | 9.2               | 16.3              | 21.9 | 21.6 | 10.0 | 2.9  | 2.9  | 9.3  | 0.65 |
| 12    | 5.8               | -0.5              | 6.8               | 8.5               | 16.5              | 21.9 | 19.5 | 10.2 | 2.9  | 2.9  | 10.8 | 0.67 |
| year  | -0.1              | -0.4              | 1.4               | -0.9              | 18.8              | 23.2 | 21.0 | 10.4 | 3.0  | 3.0  | 10.7 | 0.65 |

 Table 3. Global mean quantities generated by the model for individual months averaged over 10 years. Field names are explained in the text.

The mean temperatures of land  $(T_{land})$  and global atmosphere  $(T_{atm})$  peak in July. The net heating of the air (*atbal*) is consistent with the  $T_{atm}$  cycle. The net surface heating (*sfbal*, G for land grid-points) and the ocean heating component (*ocbal*) are not consistent with surface temperature changes due to the SSTs being imposed (heat storage by the deep ocean is implied). The sensible heat flux from the surface (*hflux*), H<sub>0</sub>, peaks with the mean surface temperature ( $T_{surf}$ ). Annual mean  $T_{surf}$  is within a degree of the observed value given by Ramanathan et al. (1989). Annual mean reflected SW and outgoing LW are also within 1 W m<sup>-2</sup> of observed values given by Ramanathan et al. (1989).

The mean precipitable water column (*atwt*) evidently varies with the mean air temperature. Of the cloud layers low (LC), middle (MC) and high (HC) only LC has a recognizable annual cycle in the global mean. The annual mean total cloud cover (TC) is within the range of observed estimates (Schlesinger and Zhao 1989). The radiative effect of the clouds can be assessed through cloud forcing diagnostics. Annual mean of the LW and SW components ( $CF_{LW}$ ,  $CF_{SW}$ ) are within 1 W m<sup>-2</sup> of observed values given by Ramanathan et al. (1989). Precipitation (*rain*), evaporation (LE or *evap*) and runoff (*runf*) all increase in the first half of the year. Annual mean

precipitation is larger than the 2.7 mm day<sup>-1</sup> of Jaeger (1976) but smaller than the mean of 3.1 mm day<sup>-1</sup> of Legates and Willmott (1990). Ground wetness (*gwet*),  $\alpha$ , mean snow (*snw*) and sea-ice (*sic*) depths are also given. Snow depths are in cm, which are equivalent numerically to mm of water. Hemispheric snow and sea-ice amounts vary greatly, as shown in Table 4.

|       |       | low  |       |      |       |      |       |      |
|-------|-------|------|-------|------|-------|------|-------|------|
|       | SH    |      | NI    | H    | SI    | H    |       | NH   |
| month | depth | area | depth | area | depth | area | depth | area |
|       | cm    | %    | cm    | %    | ст    | %    | сш    | %    |
| 1     | 12.0  | 20.1 | 32.0  | 20.0 | 6.4   | 12.2 | 6.8   | 8.2  |
| 2     | 10.8  | 18.6 | 32.4  | 26.4 | 5.2   | 11.2 | 7.2   | 9.6  |
| 3     | 10.4  | 19.6 | 31.2  | 30.9 | 4.8   | 11.1 | 7.4   | 10.8 |
| 4     | 11.0  | 21.8 | 29.1  | 30.9 | 5.0   | 11.4 | 7.4   | 11.5 |
| 5     | 12.6  | 24.5 | 24.4  | 21.3 | 6.1   | 11.8 | 7.4   | 11.5 |
| 6     | 13.6  | 27.8 | 15.2  | 5.5  | 7.0   | 12.5 | 7.3   | 10.1 |
| 7     | 15.2  | 31.5 | 7.2   | 1.9  | 8.4   | 13.3 | 6.5   | 6.8  |
| 8     | 15.7  | 35.2 | 8.5   | 1.3  | 8.6   | 14.1 | 5.1   | 4.3  |
| 9     | 15.1  | 38.3 | 17.8  | 1.8  | 8.7   | 14.7 | 4.7   | 3.8  |
| 10    | 14.8  | 39.7 | 24.7  | 3.8  | 8.7   | 14.8 | 4.8   | 4.3  |
| 11    | 14.5  | 37.0 | 27.9  | 7.6  | 8.7   | 14.6 | 5.1   | 5.4  |
| 12    | 13.8  | 28.0 | 30.9  | 13.2 | 8.2   | 13.7 | 5.9   | 6.6  |
| year  | 13.3  | 28.5 | 23.4  | 13.7 | 7.2   | 13.0 | 6.3   | 7.7  |

 Table 4. Snow and sea-ice monthly mean depths in cm averaged over the hemisphere, and areas as percentage of the hemispheric area.

### Zonal means of the atmospheric fields.

Zonal means of the atmospheric variables zonal wind (u), meridional wind (v), temperature (T) and relative humidity (RH) for the seasons December-February (DJF) and June-August (JJA) are shown in Figures 8 to 15 (these Figures are at the end of the report as Appendix F). These are compared with the observed climatology based on ECMWF analyses. The fields for 1983-88 (from Hoskins et al. 1989) are used for u, v and T. For RH the ECMWF-TOGA fields for 1985-89 were used. In DJF, both SH and NH peak zonal winds (jets) are several degrees equatorward of the observed peaks at most levels. The northern jet is around 5 m s<sup>-1</sup> too weak at 200 hPa. The easterlies are a little weak in mid-levels. In JJA the split in the southern jet, which many models fail to simulate (Boer et al. 1991), is a little excessive. The northern jet is again weaker than the climatology, by 8 m s<sup>-1</sup> at 200 hPa.

From Figures 10 and 11 we see that the basic structure of the zonal mean Hadley and Ferrel cells are simulated. However, the upper tropospheric winds peak at too low an altitude (one level down), suggesting a deficiency in the convection scheme. The modelled zonal mean temperatures (Figures 12 and 13) suffer from a cold bias of 2 to 5 K in the troposphere. Stratospheric errors are to be expected given the low vertical resolution aloft. Near both poles, the troposphere is too warm in the summer and too cool in winter, while in the lower stratosphere the reverse errors occur. Similar deficiencies are found in most other models (Boer et al. 1991).

The model's mean relative humidity (Figures 14 and 15) is broadly like the ECMWF means. It should be noted that the latter field is rather uncertain, with a drop of around 20% in the analysed field in the mid-level tropics having occurred in May 1985 (Trenberth and Olsen 1988). Note that given the cold bias in the mid troposphere, the relative and specific humidities cannot both be accurate. There is

a clear dryness in the bottom level of the model.

### Global projections

The mean sea-level pressures (MSLP) are compared with those from the ECMWF-TOGA climatology for January and July in Figures 16, 17 and 18. The global projections and zonal means of both fields are given. The model gives a generally satisfactory simulation of MSLP over most of the globe, compared with other models (Boer et al. 1991). A clear deficiency is the weakness of the Aleutian low in January. In addition, the Aleutian and Icelandic lows appear to be displaced southward by the excessive pressure in the far north. The modelled sub-Antarctic trough appears to be more than 10 hPa too weak in January, but again inaccuracies in the observations may contribute to the difference.

The peaks in the zonal wind field at 500 hPa (Figures 19 and 20) are quite successfully simulated (even to within the likely uncertainty of the observations), particularly in DJF. The wind is over 5 m s<sup>-1</sup> too westerly in the tropical eastern Pacific in DJF and 5 m s<sup>-1</sup> too easterly in the Indian Ocean in JJA. The northern summer jets are 5°S and 20°W of the observed positions.

The mean surface temperature fields for January and July are shown in Figures 21, 22 and 23 and compared with the ECMWF-TOGA values. Note that differences of a few degrees may occur simply because the latter are averages of the 0 UTC and 12 UTC fields. The SSTs are naturally very similar. The modelled land temperatures tend to be too high in summer, particularly over the northern continents. Much of the north is too cold in winter.

Precipitation is compared with the Jaeger (1976) climatology in Figures 24, 25 and 26. The basic patterns are represented in the modelled fields. The subtropical oceans appear to have too much rainfall as do the drier continents in winter. The seasonal shift of the monsoon is quite well modelled except that the heavy rainfall bands are shifted towards the central Pacific in JJA. Note that even the larger differences over the oceans are in many cases comparable to those between the Jaeger (1976) and Legates and Willmott (1990) fields. The absence of El Niño in the model SSTs may bias the rainfall climatology, and so diminish the validity of the comparison.

The total cloud cover shown in Figure 27 can be compared with observations compiled by Warren et al. (1986, 1988). Zonal means are compared with Nimbus-7 satellite observations in Figure 28. While the zonal mean values appear realistic, except near  $60^{\circ}N$  and  $90^{\circ}S$ , the regional distributions include some significant errors. Convective cloud in the Pacific tends to be shifted eastward along with the Marine stratocumulus in the eastern subtropical oceans rainfall. is deficient. Cloud cover over the northern land in summer is also too low. Cloud cover in winter at high latitudes is typically too great. The zonal means of the three cloud layers are shown in Figure 29. Amongst other considerations, the cloud scheme was tuned so that the layer values were like those constructed from the Warren et al. (1986, 1988) climatology of cloud types (Ian Smith, personal communication). The distribution of real clouds is rather uncertain. Of course, they do not form in such layers.

Zonal means of the SW and LW components of the cloud forcing and the net forcing are shown in Figure 30. These are typically within around 10 W m<sup>-2</sup> of single month fields from the ERBE data, shown by Harrison et al. (1990), except at high latitudes in winter. The outgoing LW radiation field shown in Figure 31 may be compared to that from Harrison et al. (1990). The net SW radiation at the surface is shown in Figure 32. The pattern of values is comparable to those derived by Darnell et al. (1992), except where cloud cover errors occur. Zonal means of the outgoing LW and the net SW at the surface are shown in Figure 33. The modelled global and annual mean of 165 W m<sup>-2</sup> is between the observed value of 151 W m<sup>-2</sup> from Darnell et al. (1992) and that of 169 W m<sup>-2</sup> given by Ramanathan et al. (1989).

For more realistic climate simulations, in particular modelling of global warming under steadily increasing  $CO_2$  concentrations, a coupled ocean-atmosphere GCM is desirable. Two fields important to atmospheric interaction with the ocean are shown in Figure 34. These are the annual mean heat flux into the ocean and the annual mean surface stress on the ocean. In common with most models, differences between the modelled heat flux and estimates from Esbensen and Kushnir (1981) are considerable; the model supplies around 30 W m<sup>-2</sup> too little heat to the tropical oceans and warms rather than cools the ocean at some high latitude locations.

### Australian region

The fields shown here illustrate both some of the success and some of the difficulty of simulating features at a regional scale. The mean daily maximum and minimum of surface air (screen) temperature for DJF and JJA, as calculated in Appendix B, are shown in Figures 35 and 36, together with observed values (provided by the Australian Bureau of Meteorology). The modelled values are within a few degrees of the observed, except where high topography is not resolved.

The mean upper soil layer moisture field for the four seasons is shown in Figure 37. The patterns are similar to those in the mean rainfall. There is some reduction to the north in the ratio of moisture to rainfall consistent with higher temperatures causing more rapid evaporation. Comparable observed values are not available, but judging by the actual vegetation distribution, the pattern of likely errors appears to follow from errors in rainfall. In particular, the peaks in soil wetness in the south in JJA are evidently misplaced. The centre of the continent is drier than the model predicts.

### Future Developments

In general the model appears to simulate the global climate with a degree of skill similar to that of most comparable models (e.g. Boer et al. 1991). The Division of Atmospheric Research climate modelling group is currently using the mixed-layer ocean version of the model in its greenhouse simulations.

Nevertheless, improvements in the model's performance are being sought. A vegetation and canopy scheme, as developed by Kowalczyk et al. (1991), is now being incorporated into the model. Together with a more realistic soil moisture scheme, this may help reduce surface temperatures over northern land in July. A semi-Lagrangian moisture transport scheme is being implemented and improvements in the cloud scheme are also planned. A replacement for the current simple sea-ice scheme is in preparation. A new tracer scheme which comprises semi-Lagrangian advection of trace gases and vertical transport by mixing and deep convection is also being incorporated into the model. Subject to computing resources, increases in resolution should enable more realistic representation of many physical processes and also more accurate regional simulations. It is now generally accepted that a promising route to increased computer resources for climate modelling is via parallel computers. With this in mind the model code has recently been adapted (see e.g. Rotstayn and Dix 1992) so that it can be run efficiently on a parallel shared memory Silicon Graphics computer, as an alternative to the vector-processor-based Cray Y-MP.

### Acknowledgements

We are pleased to acknowledge the encouragement and exhortations of our program leader Barrie Hunt. Valuable programming support has been provided by Wenju Cai and the figures were drafted by Sean Higgins. Helpful comments on the manuscript were provided by Jack Katzfey, Chris Mitchell and Graeme Pearman. We thank ECMWF for supplying the Global Atmospheric Data Archive and Dr. D. Schwarzkopf of GFDL for kindly supplying the radiation code. The computations were performed on the CSIRO Cray Y-MP computer at the Joint Supercomputing Facility (Melbourne, Australia). The project forms part of the CSIRO Climate Change Research Program and has been funded in part by the Department of the Environment, Sport and Territories.

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# APPENDIX A: List of symbols and abbreviations used in the text

# List of symbols

The superscript ^ indicates a weighting by surface pressure. The subscript s denotes a surface value.

| a                                | radius of the earth (6370 km)  |
|----------------------------------|--|
| c <sub>p</sub>                   | specific heat capacity of dry air (1004.64 J kg <sup>-1</sup> K <sup>-1</sup> )  |
| D                                | divergence $(\nabla^2 \chi)$   |
| Ε                                | surface evaporation  |
| F <sub>h</sub> , F <sub>m</sub>  | stability functions for heat and momentum  |
| Fr                               | Froude number  |
| g                                | acceleration due to gravity (9.81 m s <sup>-2</sup> )                            |
| G                                | total heat flux into the ground  |
| h                                | thickness of soil layer  |
| h <sub>e</sub><br>H <sub>O</sub> | launching height for gravity wave drag<br>surface sensible heat flux             |
| k                                | von Karman constant (0.4)  |
| K <sub>h</sub> , K <sub>m</sub>  | vertical diffusion coefficients for heat and momentum                            |
| K <sub>H</sub>                   | horizontal diffusion coefficient   |
| l                                | total (zonal + meridional) wavenumber  |
| L                                | latent heat of evaporation of water (2.51 x 10 <sup>6</sup> J kg <sup>-1</sup> ) |
| m                                | meridional wavenumber  |
| M <sub>p</sub>                   | convective mass flux   |
| Ν                                | Brunt-Väisälä frequency  |
| p                                | pressure   |
| p <sub>s</sub>                   | surface pressure   |
| $\mathbf{P}_{l}^{m}$             | associated Legendre polynomial of order $m$ and degree $l$                       |
| P <sub>r</sub>                   | precipitation rate   |
| q                                | mixing ratio for water vapour  |
| <b>q</b> <sub>sat</sub>          | saturation mixing ratio for water vapour   |
| R                                | specific gas constant for dry air (287 J kg <sup>-1</sup> K <sup>-1</sup> )      |
| RH                               | relative humidity  |
| Ri                               | Richardson number  |
| Т                                | temperature  |
| T <sub>0</sub>                   | isothermal mean temperature (290 K)  |
| Τ′                               | $T - T_0$  |
| T <sub>s</sub>                   | surface temperature  |
| T <sub>s2</sub>                  | temperature of subsoil layer 2   |
| T <sub>s3</sub>                  | temperature of subsoil layer 3 (lowest layer)                                    |
| T <sub>v</sub>                   | virtual temperature  |
| U, u                             | zonal velocity   |

| U                      | projection of velocity upon the surface velocity                         |
|------------------------|--|
| <b>V</b> , v           | meridional velocity  |
| <u>v</u>               | horizontal velocity vector (U,V)   |
| Wg                     | surface soil moisture  |
| w <sub>b</sub>         | deep soil moisture   |
| z <sub>0</sub>         | roughness length   |
| α                      | soil wetness factor; albedo; constant used for gravity wave drag         |
| θ                      | potential temperature  |
| ξ                      | vorticity $(\nabla^2 \psi)$  |
| ρ                      | density of air, soil or snow   |
| σ                      | $p/p_s$ ; or Stefan-Boltzmann coefficient (5.67 $\times$ 10-8 W m-2 K-4) |
| σ                      | dơ/dt  |
| τ                      | timestep number  |
| $\underline{\tau}_{s}$ | surface stress vector  |
| χ                      | velocity potential   |
| φ                      | geopotential height or latitude  |
| φ <sub>s</sub>         | surface geopotential height  |
| Ψ                      | stream function  |
| ω                      | vertical (pressure) velocity (dp/dt); angular frequency                  |
|                        |  |

# List of abbreviations

Note that the text also contains in italics some subroutine names (see Appendix E) and Fortran variable names.

| AGCM  | atmospheric general circulation model                        |
|-------|--|
| BMRC  | Bureau of Meteorology Research Centre (Australia)            |
| CSIRO | Commonwealth Scientific and Industrial Research Organisation |
|       | (Australia)  |
| DJF   | December, January, February                                  |
| ECMWF | European Centre for Medium-Range Weather Forecasts (U.K.)    |
| ERBE  | Earth Radiation Budget Experiment                            |
| FFT   | Fast Fourier Transform                                       |
| GCM   | general circulation model                                    |
| GFDL  | Geophysical Fluid Dynamics Laboratory (U.S.A.)               |
| JJA   | June, July, August   |
| LW    | long wave  |
| MAM   | March, April, May  |
| MIT   | Massachusetts Institute of Technology (U.S.A.)               |
| MLO   | mixed layer ocean  |
| NCAR  | National Center for Atmospheric Research (U.S.A.)            |
| NH    | northern hemisphere  |
| RHS   | right hand side  |
| R21   | rhomboidal truncation at 21 waves                            |
| SH    | southern hemisphere  |
| SON   | September, October, November                                 |
| SST   | sea surface temperature                                      |
| SW    | short wave   |
| TOGA  | Tropical Ocean Global Atmosphere                             |
| UKMO  | United Kingdom Meteorological Office                         |

## **APPENDIX B: Screen temperatures** (*hsflux*)

The screen temperature  $T_{scr}$  is the quantity most readily available in the observational literature; it usually refers to a level about 2 m above the ground. As a diagnostic quantity, it provides an important check on the performance of the model on a regional scale. However most GCMs produce only temperatures at the soil surface  $(T_s)$  and at the lowest atmospheric level  $(T_1)$ ;  $T_{scr}$  is *not* the equivalent of a simple linear combination of these these two temperatures. Such proxy "surface air temperatures" may not compare well with the observations, particularly when a diurnal cycle is included in the model.

To allow a better comparison with observations an explicit calculation of the screen-level temperature is included in the CSIRO9 model. The surface flux scheme used in the model implicitly specifies vertical profiles of temperature and wind which can be inverted to get a temperature compatible with the model physics. The full description of the method used to derive  $T_{scr}$  is included here because it is not a trivial task to do this correctly.

The equations for screen temperature follow on from those of the Monin-Obukhov theory described in Section 5. The screen temperature height of 2 m is assumed to lie within the constant flux layer, i.e. between the surface and the first model level.

We note that the original Monin-Obukhov approach (e.g. Businger et al. 1971; Dyer and Bradley 1982) used structure functions which expressed the vertical profiles of the temperature and wind, rather than of the mixing coefficients as used here. For example, the unstable profile of Dyer and Bradley is

$$\frac{\partial \theta}{\partial z} = - \theta_* \left(1 - \frac{14z}{L}\right)^{1/2} / k_{\rm H} z . \tag{A.1}$$

The temperature at any level could be easily calculated from this expression. However, the calculation of fluxes with this form requires a double iteration, which is avoided by the Louis (1979) procedure. Although Louis utilized the Businger structure functions, with the modified F functions used here the structure functions are not explicitly known. An alternative method must be used to calculate the screen temperature.

In the model calculation of the surface fluxes the bulk Richardson number of the layer is calculated using surface and level 1 values in (5.4). Then  $F_m$  and  $F_h$  are used to calculate the fluxes. To calculate the screen temperature consistently this process must be inverted.  $u_*$  and  $\theta_*$  are calculated from the fluxes and the system of (5.4-5.6) solved to give |V| and  $\Delta\theta$  at the screen height. Note that  $\theta$  is defined from (5.7), using  $p_s$  rather than  $p_{1000}$ . Note also that the value of  $Ri_b$  appropriate for the layer between the surface and the screen height is not the same as the value over the whole layer. Unlike the Monin-Obukhov length, it is not a proper constant. Solving (5.5) and (5.6) gives

$$|\underline{V}| = u_* / (C_{DN} F_m)^{1/2}$$
 (A.2)

$$\theta = \theta_* (C_{\rm DN} F_{\rm m})^{1/2} / (C_{\rm HN} F_{\rm h})$$
 (A.3)

Substituting in (A.2) and (A.3) into (5.4) gives

$$Ri_{b} = gz \; \theta_{*} \; (C_{DN} \; F_{m})^{3/2} \; / \; (\theta \; u_{*}^{2} \; C_{HN} \; F_{h}) \; . \tag{A.4}$$

### Stable Case

In the stable case  $F_m = F_h$  and (A.4) reduces to a simple quadratic equation for  $Ri_b$ 

$$b'Ri_b^2 + Ri_b - gz \theta_* C_{DN}^{3/2} / (\theta u_*^2 C_{HN}) = 0$$
 (A.5)

This equation has two real roots, one positive and one negative. The positive root is the physical solution, because for a stable atmosphere  $Ri_b > 0$ . After solving for  $Ri_b$ , (A.3) can be used to calculate  $\Delta\theta$ . There is a slight inconsistency here in that the model calculates  $Ri_b$  using  $\theta$  at the first level.  $Ri_b$  in (A.5) should use  $\theta$  at the screen level. However the error introduced by using the surface temperature is very small; the resulting error in  $\Delta\theta$  is of the order of  $\Delta\theta/\theta$ . The screen temperature  $T_{scr}$  is then given by

$$T_{scr} = T_s + \Delta \theta . \tag{A.6}$$

### Unstable Case

In the unstable case, the more complicated form of  $F_m$  and  $F_h$  means that (A.4) can not be solved directly. The system (A.2-A.5) is solved iteratively. A first guess value of  $Ri_b$  is used to calculate u and  $\Delta\theta$ , which are then used to update  $Ri_b$ . In practice this process converges reliably within about five iterations from a starting value of  $Ri_b = 0$  (neutral stability).

A rearrangement of the equation allows a faster Newton-Raphson method to be used. Using (A.4), define  $G(Ri_b)$  as

$$G(Ri_b) = Ri_b^2 F_h^2 - c^2 F_m^3 = 0$$
 (A.7)

where

$$c = g \ z \ \theta_* \ C_{DN}^{3/2} \ / \ (\theta \ u_*^2 \ C_{HN}) \ . \tag{A.8}$$

Substituting a new variable  $y = (-Ri_b)^{1/2}$  gives

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$$G(Ri_b) = y^4 F_h^2 - c^2 F_m^3$$
 (A.9)

$$\frac{dG}{dy} = 4y^3 F_h^2 + 2y^4 F_h \frac{dF_h}{dy} - 3c^2 F_m^2 \frac{dF_m}{dy}.$$
(A.10)

From (5.14), for the unstable case,

$$F_{m} = \{ 1 + b_{m} y^{2} / (1 + c_{m} y) \}$$
(A.11)

and

$$\frac{dF_m}{dy} = b_m y (2 - c_m y) / (1 + c_m y)^2$$
(A.12)

and similarly for  $dF_h/dy$ . With this calculation of the derivative, the standard Newton-Raphson method can be used to solve (A.7).

As G(0) < 0 and  $\lim_{y\to\infty} G(y) = \infty$  there is at least one positive solution. Graphs of the function for reasonable values of the parameters show that there is only one positive solution, but the function has a shallow minimum for y > 0. For the

iteration to converge to the correct solution, the starting value must be greater than this minimum. In practice a value of  $0.4 \text{ c}^{1/2}$  is suitable. Three iterations gives satisfactory convergence in practice.

# **APPENDIX C: Cumulus convection details**

The basis of this parameterization is a "soft" moist adjustment scheme which generates a mass flux. This scheme has evolved from the early Arakawa methods as used in the UCLA global model (Arakawa 1972). The scheme gives a vertical distribution of heating/cooling and moistening/drying which ensures that the model vertical profile tends towards the moist adiabat under convective conditions.

Defining moist static energy as  $H_k = c_pT + \phi + Lq = S + Lq$  and  $H_{sat_k}$  as the saturated quantity at level k, then the atmospheric model is moist unstable at a level k=kb for rising parcels of air if  $H_{kb} > H_{sat_{kb+1}}$ . The instability in the vertical ceases at a level kt when  $H_{kb} \leq H_{sat_{kt+1}}$ .

We define the cloud base at level kb+0.5 and the cloud top at level kt+0.5. Moist instability parameters U (for levels above cloud base) are defined by

$$U_k = (H_{kb} - H_{sat_k})/c_p > 0$$
 for  $kb+1 \le k \le kt$ . (C.1)

The temperature of the parcel at cloud top is given by

$$T_{clkt} - T_{kt} = U_{kt}/(1 + \gamma_{kt})$$
(C.2)

where

$$\gamma_{kt} = \frac{L}{c_p} \frac{\partial q_{sat}(T_{kt})}{\partial T} .$$
 (C.3)

. ... ...

For the model levels k = kb to kt, the highly reduced convective equations for temperature and moisture are given by

$$\frac{\partial S_{k}}{\partial t} = M_{p} \left( \frac{\partial S}{\partial \sigma} \right)_{k}$$
(C.4)

$$\frac{\partial q_{k}}{\partial t} = M_{p} \left( \frac{\partial q}{\partial \sigma} \right)_{k} . \tag{C.5}$$

Here  $M_p$  is the convective mass flux. Note that the RHS terms are a subset of the full equations governing convection. These equations involve terms for updrafts, downdrafts, entrainment, detrainment etc. Thus the single RHS terms are a gross simplification, but their use enables a framework for convection to be derived which is *only* dependent on the large scale model parameters. To utilize the above, we take initially a simple finite difference representation of the vertical derivatives in order to derive an energy conserving system. For k=kb

$$\frac{\partial S_{kb}}{\partial t} = M_p \left( S_{kb+0.5} - S_{kb-0.5} \right) / \Delta \sigma_{kb}$$
$$\cong M_p \left( S_{kb+0.5} - S_{kb} \right) / \Delta \sigma_{kb} . \tag{C.6}$$

Here  $S_{kb-0.5}$  has been replaced by  $S_{kb}$  to account for cloud base effects with the convection starting at a full model level. For higher levels

$$\frac{\partial S_k}{\partial t} = M_p \left( S_{k+0.5} - S_{k-0.5} \right) / \Delta \sigma_k \quad \text{for } kb+1 \le k \le kt-1 \quad (C.7)$$

and

$$\frac{\partial S_{kt}}{\partial t} = M_p \left\{ (S_{kt} - S_{kt-0.5}) + (S_{cl,kt} - S_{kt}) \right\} / \Delta \sigma_{kt} . \tag{C.8}$$

At cloud top there is the additional term involving  $S_{cl,kt}$  (i.e. the value of S at the cloud temperature  $T_{cl}$  at level k=kt) which is incorporated to account for the excess temperature of the cloud parcel over the environment at the upper convective level.

In the above, the S values at half levels have been included only in order to illustrate how total energy conservation is achieved. The RHS of these equations will later be replaced by a parameterized form which, whilst maintaining the same total heating (and associated drying), reduces the moist instabilities such that the model atmosphere tends towards a uniform moist lapse rate. Combining all k level equations gives a total change equation

$$\frac{\partial}{\partial t} \sum_{kb}^{kt} \Delta \sigma_k S_k = M_p \left\{ \left( S_{cl,kt} - S_{kt} \right) + \left( S_{kt} - S_{kb} \right) \right\}.$$
(C.9)

There is a net heating since  $S_{cl,kl} > S_{kl}$  and  $S_{kl} > S_{kb}$  (dry stable atmosphere).

In a similar manner we derive the total moisture change

$$\frac{\partial}{\partial t} \sum_{kb}^{kt} \Delta \sigma_k q_k = M_p \{ (q_{cl,kt} - q_{kt}) + (q_{kt} - q_{kb}) \}.$$
(C.10)

Note that the system is energy conserving as it can be shown that

$$\frac{L}{c_{p}} \frac{\partial}{\partial t} \sum_{kb}^{kt} \Delta \sigma_{k} q_{k} = -\frac{\partial}{\partial t} \sum_{kb}^{kt} \Delta \sigma_{k} T_{k}$$
(C.11)

(where  $\frac{\partial S_k}{\partial t} \approx c_p \frac{\partial T_k}{\partial t}$  has been assumed - i.e. changes in  $\phi$  ignored).

The individual temperature changes for a convective level k are now rewritten as

$$\frac{\partial T_{k}}{\partial t} = \frac{M_{p}}{c_{p}} \frac{dS_{k}}{\Delta \sigma_{k}}$$
(C.12)

with

$$(dS_{kt} + dS_{kt-1} + .... + dS_{kb})/c_{p} = \{ (S_{cl,kt} - S_{kt}) + (S_{kt} - S_{kb}) \}/c_{p}$$
$$= U_{kl}/(1 + \gamma_{kl}) + (S_{kt} - S_{kb})/c_{p} \text{ denoted by HPC.}$$
(C.13)

Note that the temperature changes are written in terms of " $dS_k$ " instead of  $\Delta S_k$  (the latter would be based on the simple finite difference forms given in (C.6) to (C.8) bove). The derivation of the  $dS_k$  (and similarly  $dq_k$  for the equivalent moisture aequations) now follows.

The moisture change at level k is given by

$$\frac{\partial q_k}{\partial t} = -M_p \frac{dq_k}{\Delta \sigma_k}$$
(C.14)

with

$$(dq_{kt} + dq_{kt-1} + \dots + dq_{kb}) = (q_{kb} - q_{cl,kt}) .$$
(C.15)

In the case of moisture changes, we know from observations (Yanai et al. 1973) that there is mainly drying of the moist levels during convection. As a *first approximation*, bearing in mind that formally deriving moisture changes using discrete model levels from the full convection equations is difficult with q varying approximately in the vertical as  $p^3$  (i.e.  $\sigma^3$ ), we set

$$dq_k = \alpha (q_k - \varepsilon q_{sat_k}) \Delta \sigma_k \quad \text{for } kb \le k \le kt, \quad 0 < \varepsilon < 1 .$$
 (C.16)

In this expression,  $\varepsilon$  can be used to ensure that not only is there drying for the moist levels but there is moistening for dry levels that the convection may pass through. If  $\varepsilon = 0$ , then there will be drying at all levels.

The current model uses  $\varepsilon = \min(0.6, 0.9\varepsilon^*)$  where 0.6 gives a 60% RH cut-off for net moistening/drying. Note that this preset cut-off must not exceed the value at which the solution becomes indeterminate which is defined by

$$\epsilon *$$
 =  $\sum q_k \Delta \sigma_k$  /  $\sum q_{sat_k} \Delta \sigma_k$  .

It then follows that

$$\alpha = (q_{kb} - q_{cl,kl}) / \left\{ \sum q_k \Delta \sigma_k - \epsilon \sum q_{sat_k} \Delta \sigma_k \right\}.$$
(C.17)

Having achieved a suitable partitioning of the moisture changes in the vertical, the next stage deals with the temperature changes. This involves finding a convective mass flux  $M_p$ . Here we adopt Arakawa's assumption that the convection leads to an exponential decay of the instability which is modelled through the instability parameters  $U_{kb+1}$  through to  $U_{kt}$ . Thus we use

$$\frac{\partial U_k}{\partial t} = -\frac{U_k}{\tau_r} \qquad \text{for } kb+1 \le k \le kt$$
 (C.18)

and  $\tau_r$  is a convective relaxation time (set currently at 1 hour for the CSIRO9 model which uses half hour timesteps). Note that the U<sub>k</sub> values for different levels k are *not* equal. We may eliminate  $\tau_r$  from pairs of equations (using the kb+1 equation repeatedly):

$$U_{k} \frac{\partial U_{kb+1}}{\partial t} = U_{kb+1} \frac{\partial U_{k}}{\partial t} \qquad \text{for } kb+2 \le k \le kt .$$
 (C.19)

This method will be shown to allow for the elimination of the unknown  $M_p$ , and give a system of equations that have a solution if an assumption is made about the amount of cloud base heating.

We use the above expressions to first determine a suitable form for the heating profile. The values to be derived for  $dS_{kb}$ ,  $dS_{kb+1}$ ,  $dS_{kb+2}$ , ...,  $dS_{kt}$  reduce the moist instability parameters at each level according to their relative magnitudes.

The moist instability parameter for level kb+1 is given by  $U_{kb+1}$  and the change due to the convection is given by

$$\frac{\partial U_{kb+1}}{\partial t} = \frac{\partial}{\partial t} \left\{ H_{kb} - H_{sat_{kb+1}} \right\} / c_p$$
$$= \frac{\partial}{\partial t} \left\{ L(q_{kb} - q_{sat_{kb+1}}) - (S_{kb+1} - S_{kb}) \right\} / c_p .$$
(C.20)

From the definition of dry static energy  $S = c_p T + \phi$ , and using the hydrostatic equation in the form  $\partial \phi / \partial \ln(\sigma) = -RT$ , we obtain

$$(S_{kb+1} - S_{kb})/c_p = (1 + \delta_{kb+1}) T_{kb+1} - (1 - \delta_{kb+1}) T_{kb}$$
(C.21)

where a  $ln(\sigma)$  profile for T has been used, giving

$$\delta_{kb+1} = -0.5 \ R \ln(\sigma_{kb+1}/\sigma_{kb}) \ /c_p \ .$$
 (C.22)

Substituting (C.21) into (C.20) yields

$$\frac{\partial U_{kb+1}}{\partial t} = \frac{L}{c_p} \frac{\partial q_{kb}}{\partial t} + (1 - \delta_{kb+1}) \frac{\partial T_{kb}}{\partial t} - (1 + \gamma_{kb+1} + \delta_{kb+1}) \frac{\partial T_{kb+1}}{\partial t}$$
(C.23)

where the following expansion for  $\partial q_{sat}/\partial t$  has been used, defining  $\gamma$ 

$$\frac{\partial q_{\text{sat}_k}}{\partial t} = \frac{\partial q_{\text{sat}_k}}{\partial T} \frac{\partial T_k}{\partial t} = \frac{c_p}{L} \gamma_k \frac{\partial T_k}{\partial t} . \qquad (C.24)$$

Similarly, using

$$(S_{kb+2} - S_{kb})/c_{p} = (S_{kb+2} - S_{kb+1})/c_{p} + (S_{kb+1} - S_{kb})/c_{p}$$
  
= (1 +  $\delta_{kb+2}$ ) T<sub>kb+2</sub> + ( $\delta_{kb+1}$  +  $\delta_{kb+2}$ ) T<sub>kb+1</sub>  
= (1 +  $\delta_{kb+2}$ ) T<sub>kb+2</sub> + ( $\delta_{kb+1}$  +  $\delta_{kb+2}$ ) T<sub>kb+1</sub> - (1 -  $\delta_{kb+1}$ ) T<sub>kb</sub> (C.25)

implies that

$$\frac{\partial U_{k^{b+2}}}{\partial t} = \left\{ \begin{array}{c} \frac{L}{c_{p}} \frac{\partial q_{kb}}{\partial t} + (1 - \delta_{kb+1}) \frac{\partial T_{kb}}{\partial t} \end{array} \right\} \\ - (\delta_{kb+1} + \delta_{kb+2}) \frac{\partial T_{kb+2}}{\partial t} - (1 + \gamma_{kb+2} + \delta_{kb+2}) \frac{\partial T_{kb+2}}{\partial t} \end{array}$$
(C.26)

and so on for all levels up to  $\boldsymbol{U}_{kt}$  . We have already set

$$\frac{\partial q_{kb}}{\partial t} = -M_p \frac{dq_{kb}}{\Delta \sigma_{kb}} = M_p DQ_{kb} \text{ (by definition)}$$
(C.27)

where  $dq_{kb}$  is known from (C.16) and (C.17). Now put

$$\frac{\partial T_k}{\partial t} = \frac{M_p}{c_p} \frac{dS_k}{\Delta \sigma_k} = M_p DS_k \text{ for } kb \le k \le kt .$$
(C.28)

We also require from (C.13) that

$$(dS_{kt} + dS_{kt-1} + ... + dS_{kb})/c_p = HPC$$
. (C.29)

In the above set of equations (C.19, C.23, C.26, C.27, C.28 and C.29) it will be found that the number of unknowns  $(DS_{kt}, DS_{kt-1},..., DS_{kb})$  is one greater than the number of equations. A closure is thus required. This is obtained by using observational evidence which suggests that the heating at cloud base is generally small (much smaller than mid to upper levels). Thus as an initial guess we set  $DS_{kb} = 0$ . This is in line with other convection schemes such as Kuo.

Combining the above equations, we can now rewrite (C.19) as a set of equations in the unknowns  $DS_{kt}$ ,  $DS_{kt-1}$ , ...,  $DS_{kb+1}$ . These equations can be shown to be solvable by simple elimination and back substitution (no matrix inversion is required). They are of the form

Fn( 
$$DS_{kb+1}$$
,  $DS_{kb+2}$ ) =  $A_1$   
Fn(  $DS_{kb+1}$ ,  $DS_{kb+2}$ ,  $DS_{kb+3}$ ) =  $A_2$  (C.30)  
 $\vdots$   
Fn(  $DS_{kb+1}$ ,  $DS_{kb+2}$ ,.....,  $DS_{kt}$ ) =  $A_{N-1}$ 

where kb+N = kt. From (C.26) and (C.27) we also have

$$\Delta \sigma_{kb+1} DS_{kb+1} + \Delta \sigma_{kb+2} DS_{kb+2} + \dots + \Delta \sigma_{kt} DS_{kt} = HPC$$
  
Fn( DS<sub>kb+1</sub>, DS<sub>kb+2</sub>,...., DS<sub>kt</sub> ) = A<sub>N</sub> . (C.31)

In the above the  $A_1$ ,  $A_2$ , ...,  $A_N$  are known, and we may solve as described. We now need the mass flux  $M_p$ . For this we use

$$\frac{\partial \mathbf{U}_{\mathbf{k}\mathbf{b}+1}}{\partial \mathbf{t}} = -\frac{\mathbf{U}_{\mathbf{k}\mathbf{b}+1}}{\tau_{\mathbf{t}}}$$

and

i.e.

$$\frac{\partial U_{kb+1}}{\partial t} = -\frac{L}{c_p} M_p DQ_{kb} + (1 - \delta_{kb+1}) M_p DS_{kb} - (1 + \gamma_{kb+1} + \delta_{kb+1}) M_p DS_{kb+1}$$
(C.32)

which gives  $M_p$  upon eliminating  $\partial U_{kb+1}/\partial t$ . The temperature and moisture changes due to convection may now be written as

$$T'_{k} = T_{k} + 2\Delta t M_{p} DS_{k}$$
 for  $kb \le k \le kt$  (C.33)

$$q'_k = q_k + 2\Delta t M_p DQ_k$$
 for  $kb \le k \le kt$ . (C.34)

In conclusion, there are several useful features of this convection parameterization:

- a) By the use of an exponential decay of moist instability between the cloud base and each level above it up to cloud top, the model atmosphere tends towards a uniform moist adiabatic profile.
- b) The moisture changes are formulated for drying of the most saturated levels with an optional amount of moistening of the driest levels.
- c) The rate of decay of the moist instability can be controlled by an adjustable convective relaxation time.
- d) A convective mass flux is computed, and can be used to compute momentum mixing by convection.
- c) Unlike Kuo type schemes, this parameterization does not need to know about moisture convergence. However, in practice, it is found that for convection to be initiated and sustained, there has to be convergence. The convection is also not limited by the RH value at cloud base (except in that physically, convection will not commence unless the RH at cloud base is high, but not necessarily saturated).

# **APPENDIX D: Particle trajectory facility** (tracer)

The CSIRO9 model has the facility to keep track of the position of particles (tracers) released at a selected points in the model atmosphere. This may be used, for example, to show inter-hemispheric (cross equatorial) flow patterns, and to track tropospheric/stratospheric interactions. Another example of the use of this has been to estimate possible final positions of smoke particles following their release due to the burning of the Kuwait oil installations following the 1991 Gulf war.

The particles can be released at any  $\sigma$  height, and at any position within a grid square by means of latitude/longitude coordinates. The particles can be released at every timestep, and then their route is computed using the model velocities. The horizontal velocities (u,v) are centered on the Gaussian grid, whilst the vertical velocities  $d\sigma/dt$  are at the half levels. Dispersion due to wind variations at smaller scales, both horizontally and vertically, is absent. The separation with time of particles which are initially close is therefore likely to be underestimated. A form of subgrid scale vertical motion which may be particularly important is that within convective towers. The modelled vertical winds respond only indirectly to the convective latent heating. There is thus an underestimation of the deflection upwards of air trajectories in the regions of cumulus activity. The modelled vertical deflections are considerable nevertheless. There is no modelling of particle removal processes such as washout of particles by rainfall or gravitational fall out.

The calculation of trajectories is made during the running of the model, thereby avoiding the need to store wind fields for all points for the duration of the trajectory run. Given the vector position  $\underline{x}^{\tau}$  (where  $\underline{x}$  denotes longitude, latitude, and  $\sigma$ ) at time  $\tau$ , the position at the next timestep is computed (by means of forward stepping) as

$$\underline{\mathbf{x}}^{\tau+1} = \underline{\mathbf{x}}^{\tau} + \Delta t \ \underline{\mathbf{W}}^{\tau} \tag{D.1}$$

where  $\underline{W}^{\tau}$  is the velocity obtained from linear interpolation of the modelled velocities at the 8 vertices of the grid box containing  $\underline{x}^{\tau}$ , and  $\Delta t$  is the model timestep. In this calculation the winds at the earth's surface are taken to be those of the first model level (which is about 180 m above the ground).

# **APPENDIX E:** List of subroutine names

The subroutines are listed here in the order in which they are called. The indentation of each routine indicates its position in the calling tree. The routines from *realim* through *timet* inclusive are within the timestep loop. A few optional routines of a diagnostic nature are not shown.

| csiro9             | Model control and timestep loop.                                       |
|--------------------|--|
| inital             | Initialize model parameters.   |
| openfl             | Open files.  |
| initax             | Further initialization.  |
| filerd             | Read in model initial condition.                                       |
| gauleg             | Driver routine to set up Gaussian weights and Legendre polynomials.    |
| gaussy             | Calculate Gaussian weights and latitudes.                              |
| ordlev             | Generating routine for Gaussian latitudes.                             |
| londre             | Set up Legendre polynomials.   |
| 7erost             | Zero statistics arrays.  |
| realim             | Convert complex spectral arrays to real and imaginary arrays.          |
| wharm              | Form spectral velocity and frictional components.                      |
| avsol              | Calculate zenith angle at each point.                                  |
| zernoi             | Routine to manipulate various arrays before physics and dynamics.      |
| nhvs               | Physics loop over latitude: re-create spectral fields at end.          |
| aneofix            | Remove negative mixing ratios by borrowing from other points.          |
| ntog               | Convert required spectral fields to grid (Legendre transform and FFT). |
| mfto               | Interface to Cray (inverse) FFTs.                                      |
| radin              | Interface to physics routines: also collect statistics.                |
| surfset            | Set surface type, albedo, roughness lengths etc.                       |
| asoam              | Calculate saturation mixing ratio.                                     |
| neorm              | Calculate relative humidity.   |
| heflur             | Calculate surface fluxes of heat, momentum and moisture.               |
| radfs              | Interface to Fels-Schwarzkopf radiation code (every 4th timestep).     |
| hconst             | Define numerical constants at start of run.                            |
| table              | Precalculate properties of radiation bands at start of run.            |
| solar              | Calculate sun position   |
| zonith             | Calculate mean zenith angle for the radiation timester.                |
| 2 chin<br>o 3 set  | Internolate seasonal ozone data to particular time.                    |
| rasat              | Initialization for asset interpolation at start of run.                |
| aloud              | Set cloud properties for radiation code                                |
| cioua              | Diagnose cloud amounts and levels                                      |
| ciaaia<br>sur80    | Solar radiation calculation  |
| swi03              | Set up cloud amounts and overlap factors for longwave calculation.     |
| 1,                 | Longwave radiation driving routine                                     |
| 1W/00<br>fat98     | Main I W calculation routine includes emissivity calculations.         |
| J\$100<br>a1a288   | Colculate I W exchange terms between atmospheric layers                |
| e1e200<br>a2u99    | Calculate nearby layer transmissivities for water vanor                |
| e5700              | Calculate "cool to space" heating rates                                |
| spuoo<br>surfuna   | Undate soil temperature and moisture using the surface fluxes.         |
| surjupa<br>hvortmr | Apply vertical mixing and shallow convection.                          |
| trim               | Tridiagonal solver for vertical mixing scheme.                         |
| awdraa             | Calculate gravity wave drag  |
| rainda             | Calculate large-scale rainfall   |
| ranaa              | Calculate cumulus convection and convective rainfall.                  |
| comir              | Calculate vertical mixing of momentum due to convection.               |
| c vnix<br>surfunh  | Undate soil moisture and snow cover due to rainfall.                   |
| sur jupo<br>mfftm  | Interface to Cray (forward) FFTs called by <i>phys</i> after physics.  |
| dynm               | Dynamics loop over latitude: re-create spectral fields at cnd.         |
| dton               | Convert required spectral fields to grid (Legendre transform and FFT). |
| uiug               | Interface to Crav (inverse) FFTs                                       |
| nyjig<br>adiff     | Annly moisture diffusion   |
| quijj              | Compute non-linear terms in grid space                                 |
| aynmni             | Compute non-initial terms in grid space.                               |
| aynmst             | Store pressure lever una every six nours.                              |

| mfftm   | Interface to Cray (inverse) FFTs.                      |
|---------|--|
| energy  | Calculate energy diagnostics at preset intervals.      |
| linear  | Add the linear spectral tendencies.                    |
| assel   | Apply the Robert (Asselin) time filter.                |
| matset  | Set up matrices for semi-implicit time integration.    |
| mtxmtx  | Matrix multiplication routine.                         |
| semii 👘 | Perform the semi-implicit time integration.            |
| diffn   | Incorporate the forward implicit horizontal diffusion. |
| timet   | Update day count and update SSTs at end of month.      |
| filewr  | Write restart file to disk at end of run.              |
| filest  | Save time averaged global fields at preset intervals.  |
| collst  | Array manipulation routine.                            |

# **APPENDIX F: Figures of model climatology**

Figures of the model climatology, as described in Section 18, are provided in this Appendix.

| Figure | 8.  | Zonal mean of zonal wind averaged over DJF for (a) model            |
|--------|-----|---|
|        |     | and (b) model minus ECMWF 6 y data. Units are m s <sup>-1</sup> .   |
|        |     | Shading range is given in small boxes.                              |
| Figure | 9.  | Zonal mean of zonal wind averaged over JJA for (a) model            |
|        |     | and (b) model minus ECMWF data. Units are m s <sup>-1</sup> .       |
| Figure | 10. | Zonal mean of meridional wind averaged over DJF for                 |
|        |     | (a) model and (b) ECMWF data. Units are m s <sup>-1</sup> .         |
| Figure | 11. | Zonal mean of meridional wind averaged over JJA for                 |
|        |     | (a) model and (b) ECMWF data. Units are m s <sup>-1</sup> .         |
| Figure | 12. | Zonal mean of temperature in K averaged over DJF for                |
|        |     | (a) model and (b) model minus ECMWF data.                           |
| Figure | 13. | Zonal mean of temperature in K averaged over JJA for                |
|        |     | (a) model and (b) model minus ECMWF data.                           |
| Figure | 14. | Zonal mean of relative humidity in percent averaged over DJF for    |
|        |     | (a) model and (b) ECMWF-TOGA 5 y data.                              |
| Figure | 15. | Zonal mean of relative humidity in percent averaged over JJA for    |
|        |     | (a) model and (b) ECMWF-TOGA data.                                  |
| Figure | 16. | Mean sea-level pressure for January for (a) model,                  |
|        |     | (b) ECMWF-TOGA data. Units are hPa.                                 |
| Figure | 17. | Mean sea-level pressure for July for (a) model,                     |
|        |     | (b) ECMWF-TOGA data. Units are hPa.                                 |
| Figure | 18. | Zonal mean of mean sea-level pressure from model and ECMWF-TOGA     |
|        |     | data for (a) January and (b) July.                                  |
| Figure | 19. | Zonal wind at 500 hPa averaged over DJF for (a) model and (b) ECMWF |
|        |     | data. Units are m s <sup>-1</sup> .                                 |

- Figure 20. Zonal wind at 500 hPa averaged over JJA for (a) model and (b) ECMWF data. Units are m s<sup>-1</sup>.
- Figure 21. Mean surface temperature for January in °C for (a) model, (b) model minus ECMWF-TOGA data and (c) zonal means. Note in (a) interval is 5°C below 15°C, interval is 3°C above 15°C. In (b) light dots shade regions 2°C to 6°C, dense dots indicate above 6°C. Light hatching indicates -6°C to -2°C and dense hatching below -6°C.
- Figure 22. Mean surface temperature for July in °C for (a) model, (b) model minus ECMWF-TOGA and (c) zonal means. Contours and shading as in Fig. 21.
- Figure 23. Zonal mean surface temperature in °C from model and ECMWF-TOGA data for (a) January and (b) July.
- Figure 24. Mean precipitation in mm/day averaged over DJF for (a) model, (b) Jaeger data. Contours are 1, 2, 5 and 10 mm/day.
- Figure 25. Mean precipitation in mm/day averaged over JJA for (a) model, (b) Jaeger data. Contours are 1, 2, 5 and 10 mm/day.
- Figure 26. Zonal mean precipitation in mm/day from model and Jaeger data for (a) DJF and (b) JJA.
- Figure 27. Mean total cloud cover from model in percent for (a) DJF and (b) JJA.
- Figure 28. Zonal mean total cloud cover from model and Nimbus-7 observations for (a) January and (b) July.
- Figure 29. Zonal mean cloud cover for each layer from model in percent for (a) DJF and (b) JJA.
- Figure 30. Zonal mean cloud forcing (LW, SW and net) from model for (a) DJF and (b) JJA. Units are W m<sup>-2</sup>.
- Figure 31. Mean outgoing long wave radiation from model for (a) DJF, (b) JJA. Units are W m<sup>-2</sup>.
- Figure 32. Mean net solar radiation at ground from model for (a) DJF, (b) JJA. Units are W m<sup>-2</sup>.
- Figure 33. Zonal means over DJF and JJA of (a) outgoing long wave radiation from model and (b) net solar radiation at ground from model. Units are W m<sup>-2</sup>.
- Figure 34. Annual mean (a) net heat flux into surface from model in W m<sup>-2</sup> and (b) mean stress on surface from model in N m<sup>-2</sup>.
- Figure 35. Mean daily extreme surface air temperatures for DJF in °C: (a) minimum from model, (b) maximum from model, (c) observed minimum and (d) observed maximum. Contour levels are 10, 14, 18, 22, 26, 30, 34, 38 °C.
- Figure 36. Mean daily extreme surface air temperatures for JJA in °C: (a) minimum from model, (b) maximum from model, (c) observed minimum and (d) observed maximum. Contour levels are 0, 4, 8, 12, 16, 20, 24, 28 °C.
- Figure 37. Mean soil wetness (as a volume fraction  $\leq 0.36$ ) from model for (a) DJF, (b) MAM, (c) JJA and (d) SON.



Figure 8. Zonal mean of zonal wind averaged over DJF for (a) model and (b) model minus ECMWF 6 y data. Units are m s<sup>-1</sup>. Shading range is given in small boxes.



Figure 9. Zonal mean of zonal wind averaged over JJA for (a) model and (b) model minus ECMWF data. Units are m s<sup>-1</sup>.



Figure 10. Zonal mean of meridional wind averaged over DJF for (a) model and (b) ECMWF data. Units are m s<sup>-1</sup>.



Figure 11. Zonal mean of meridional wind averaged over JJA for (a) model and (b) ECMWF data. Units are m s<sup>-1</sup>.







CONTOUR FROM -5 TO 8 BY 1





Figure 14. Zonal mean of relative humidity in percent averaged over DJF for (a) model and (b) ECMWF-TOGA 5 y data.



Figure 15. Zonal mean of relative humidity in percent averaged over JJA for (a) model and (b) ECMWF-TOGA data.



Figure 16. Mean sea-level pressure for January for (a) model, (b) ECMWF-TOGA data. Units are hPa.



Figure 17. Mean sea-level pressure for July for (a) model, (b) ECMWF-TOGA data. Units are hPa.



Figure 18. Zonal mean of mean sea-level pressure from model and ECMWF-TOGA data for (a) January and (b) July.



Figure 19. Zonal wind at 500 hPa averaged over DJF for (a) model and (b) ECMWF data. Units are m s<sup>-1</sup>.


Figure 20. Zonal wind at 500 hPa averaged over JJA for (a) model and (b) ECMWF data. Units are m s<sup>-1</sup>.



Figure 21. Mean surface temperature for January in °C for (a) model, (b) model minus ECMWF-TOGA data and (c) zonal means. Note in (a) interval is 5°C below 15°C, interval is 3°C above 15°C. In (b) light dots shade regions 2°C to 6°C, dense dots indicate above 6°C. Light hatching indicates -6°C to -2°C and dense hatching below -6°C.



Figure 22. Mean surface temperature for July in °C for (a) model, (b) model minus ECMWF-TOGA and (c) zonal means. Contours and shading as in Fig. 21.



Figure 23. Zonal mean surface temperature in °C from model and ECMWF-TOGA data for (a) January and (b) July.



Figure 24. Mean precipitation in mm/day averaged over DJF for (a) model, (b) Jaeger data. Contours are 1, 2, 5 and 10 mm/day.



Figure 25. Mean precipitation in mm/day averaged over JJA for (a) model, (b) Jaeger data. Contours are 1, 2, 5 and 10 mm/day.



Figure 26. Zonal mean precipitation in mm/day from model and Jaeger data for (a) DJF and (b) JJA.



Figure 27. Mean total cloud cover from model in percent for (a) DJF and (b) JJA.



Figure 28. Zonal mean total cloud cover from model and Nimbus-7 observations for (a) January and (b) July.



Figure 29. Zonal mean cloud cover for each layer from model in percent for (a) DJF and (b) JJA.



Figure 30. Zonal mean cloud forcing (LW, SW and net) from model for (a) DJF and (b) JJA. Units are W m<sup>-2</sup>.



Figure 31. Mean outgoing long wave radiation from model for (a) DJF, (b) JJA. Units are W  $m^{-2}$ .



Figure 32. Mean net solar radiation at ground from model for (a) DJF, (b) JJA. Units are W  $m^{-2}$ .



Figure 33. Zonal means over DJF and JJA of (a) outgoing long wave radiation from model and (b) net solar radiation at ground from model. Units are W m<sup>-2</sup>.



Figure 34. Annual mean (a) net heat flux into surface from model in W m<sup>-2</sup> and (b) mean stress on surface from model in N m<sup>-2</sup>.



Figure 35. Mean daily extreme surface air temperatures for DJF in °C: (a) minimum from model, (b) maximum from model, (c) observed minimum and (d) observed maximum. Contour levels are 10, 14, 18, 22, 26, 30, 34, 38 °C.





Figure 36. Mean daily extreme surface air temperatures for JJA in °C: (a) minimum from model, (b) maximum from model, (c) observed minimum and (d) observed maximum. Contour levels are 0, 4, 8, 12, 16, 20, 24, 28 °C.





Figure 37. Mean soil wetness (as a volume fraction  $\leq 0.36$ ) from model for (a) DJF, (b) MAM, (c) JJA and (d) SON.

