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A moisture transport and precipitation parameterization for energy balance climate models

Chu, Shaoping, M.S.

Rice University, 1992
RICE UNIVERSITY

A MOISTURE TRANSPORT AND PRECIPITATION PARAMETERIZATION FOR ENERGY BALANCE CLIMATE MODELS

BY

SHAOPING CHU

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE MASTER OF SCIENCE

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Houston, Texas
October, 1991
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by

Shaoping Chu

ABSTRACT

The spatial distribution of water in all its three phases is an important factor in determining the climate. The interactions among temperature, water vapor, infrared emission and solar radiation form a series of feedback mechanisms, which play a very important role in the climate system. In order to trace moisture flow through the climate system and examine its impact on climate, a parameterization for the computation of moisture transport and precipitation is developed, one that will eventually be incorporated into a coupled energy balance climate-thermodynamic sea ice model (the CCSI model). This parameterization is tested by comparing computed energy transports and precipitation rates with available observations and by evaluating its sensitivity to variations in the values of specified parameters. The results of these studies indicate that the moisture parameterization is somewhat sensitive to variations in wind speed, surface air temperature and moisture flux, while it is relatively insensitive to changes in relative humidity. In general this parameterization does a good job in simulating the seasonal cycle and latitudinal distribution of the wind speed, moisture transport and precipitation when compared to the observed data and general circulation model (GCM) results.
ACKNOWLEDGMENTS

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I. Introduction

The climate is a statistical representation of the very complex interactive physical system made up of the Earth, ocean, atmosphere, biosphere and cryosphere (ice and snow). The causes of climate change are both natural and anthropogenic. In the past, variations in climate resulted from natural causes, but now human activities may be reaching the level where they will have a measurable impact on the climate.

It is theorized that climatic perturbations can occur in response to external changes in solar radiation and to internal changes in albedo, sea surface temperature, cloud cover, atmospheric composition, to mention a few. With so many possible influences on the climate, it is clear that mathematical modeling is a very good tool for examining the impact of each on the climate system. Numerous such models have been developed, ranging from simple zero-dimensional energy balance models (EBM) to complex three-dimensional general circulation models (GCM) (for general reviews, see Schneider and Dickinson 1974, Shine and Henderson-Sellers 1983, M Sankar-Rao 1986).

Early research led to the development of simple, highly empirical, one-dimensional models (e.g., Budyko, 1969, 1972; Sellers, 1969, 1973) along with attempts to more directly access the effects of various atmospheric modification (e.g., Manabe, 1971; Rasool and Schneider, 1971; Mitchell, 1971). In 1969, Budyko and Seller proposed two simple energy balance models. The models considered only annual averaged conditions and were based on the thermodynamic energy equation for the Earth-atmosphere system. Their studies showed that the sensitivity of the climate might be greatly enhanced by the temperature-albedo feedback mechanism. Both authors pointed out that, according to their models, a decrease of the solar constant by only a few percent would be sufficient to initiate
another ice age. The possibility of such extreme sensitivity to external controls, and similarly to internal parameters possibly influenced by man's activities, has led to a great spurt of interest in understanding the Earth's climate. Budyko-Sellers type model was further investigated by a number of workers. North and Coakley (1979) expanded the Budyko-Sellers mean annual energy balance climate model to include the seasonal cycle. Their results revealed an asymmetry introduced by the seasonal cycle on the climate system. During the winter, albedo is higher than during the summer. This results in absorbed solar radiation being lower during the winter and higher during the summer. The seasonal oscillation in albedo causes a warming because the hemisphere has reduced snow coverage in the summer and thus absorbs well, while in the winter, when snow advances, the incident solar radiation is reduced so the high reflectivity matters less. Therefore, the seasonal model has a warmer hemisphere than might have been obtained with a mean annual snow line.

Recently a great deal of effort in climate modeling has been devoted to parameterizing the complicated feedback mechanisms which link the surface state to atmospheric processes. In current models, surface albedo and the water related parameters appear to be the most frequently used in describing radiation and hydrologic properties, respectively. Alan Robock (1982) developed a parameterization of snow and ice area and albedo as functions of land and sea surface air temperature, incorporated it into a seasonal energy-balance model, and tested the model sensitivity to external forcings of climate change, such as solar constant variations and changes in the atmospheric carbon dioxide amount. His studies showed that ice and snow feedbacks produce enhanced global sensitivity, the effect is same for doubling CO\textsubscript{2} or for increasing or decreasing the solar constant. Gutman, Ohring and Joseph (1984) pointed out that since the surface state is closely related to climate, surface parameters should be included in a description of the
long-term interaction between surface characteristics (soil, vegetation and moisture) and the atmosphere. By using a zonal climate model including such a biogeophysical feedback mechanism, their studies showed that precipitation can be influenced by biofeedback, and the incorporation of biofeedback is desirable when simulating long-term impacts of surface alterations.

It is known that sea ice plays a very important role in the climate change. When sea ice is present, it acts like an insulator between the relatively warm ocean and the cold winter atmosphere, severely diminishing the transfer of energy and moisture to the atmosphere. Sea ice and snow also have high albedos compare to that of open ocean, which causes a large reduction in absorption of shortwave radiation. In order to study the interactions between the atmosphere, ocean and sea ice, a coupled energy balance climate-thermodynamic sea ice model (i.e. coupled climate sea ice model, CCSI model) has been developed (Ledley, 1988 a&b).

Previous research with this model investigated the impact of the feedback mechanisms between the atmosphere, ocean and sea ice on the surface energy exchanges, and thus on both polar and global climate (Ledley, 1985 a&b, 1988 a&b, 1990, 1991 a&b). It showed that snow and ice cover strongly influence the radiation field and the interactions between temperature, water vapor, infrared radiation and solar radiation. These interactions form a series of feedback mechanisms, which play a very important role in the climate system (Ledley, 1985 a&b, 1988 a&b, 1990, 1991 a&b). The studies also showed that the impact of snow and its thermal properties on sea ice is a very important factor in shaping climate. The moisture for snow comes from the ocean, and ice sheets are made up of snowfall in long term glacial-interglacial cycles. So, it is important to trace moisture flow through the climate system, and to determine the response of precipitation rates and
atmospheric moisture and the resulting response of climate to man-induced perturbations in the climate system.

The object of the work presented here is to develop and test a parameterization of moisture transport and precipitation. This parameterization will eventually be incorporated into the CCSI model. The expanded CCSI model will be used to examine the direct effect of environmental changes on the energy and moisture exchanges between the different parts of the climate system, and their subsequent effect on the climate.

The next chapter describes this parameterization in detail. In chapter III the sensitivity of the parameterization to prescribed parameters and a comparison between simulated and observed moisture transports and precipitation are described. In chapter IV the results are discussed and summarized and the future plans for this parameterization are outlined.
II. PARAMETERIZATION OF HYDROLOGIC CYCLE

General Description

The parameterization of the hydrologic cycle has been developed to be incorporated into the CCSI model. The energy balance climate model used in the CCSI model encompasses four parts of the climate system: the atmosphere over land, the atmosphere over ocean, a mixed layer ocean, and a ground layer. Energy fluxes are computed at the top of the atmosphere, at the atmosphere-surface interface, and between latitude zones over land and sea, and are specified in the ocean. This energy balance climate model is coupled to the three layer thermodynamic sea ice model (Ledley, 1985a), which includes conduction within the ice and snow, penetration of solar radiation into the ice, surface energy balances, leads and sea-ice transport. The current version of the model employs a 10° latitude grid, with land-sea resolution in each zone distributed in accordance with current land-sea distribution. The land masses have been offset relative to one another, in order to simulate a meridional transport across each 10° latitude circle from sea to land, sea to sea, land to land, and land to sea, similar to that which actually exists. A detailed description of the CCSI model may be found in Ledley (1988b, 1991a).

A typical grid area for the parameterization of the hydrologic cycle is shown in Figure 1 with all the moisture fluxes across the proper boundaries. Computations are performed over both land and sea in each zone. The results are then weighted by the relative fraction of land and sea in a zone. Therefore in zones where either the land or sea fraction is small, the computed changes for the small area have a similarly small effect on the climate.
Figure 1. A typical land-sea grid area.

E = atmospheric eddy moisture flux
M = atmospheric moisture flux due to mean winds
$\Delta xS$ = atmospheric zonal moisture flux
SR = solar radiation
IR = terrestrial radiation
The parameterization of moisture transport will follow that of Sellers (1973), in which transport by mean meridional and eddy meridional motions are computed separately. Transport by mean meridional motions is a function of the mean meridional wind velocity, which is a function of surface air temperature gradient; the surface specific humidity; surface air pressure; surface vapor pressure; and precipitable water (see Appendix D for definition). The transport by eddy meridional motions is a function of the surface specific humidity gradient; an eddy diffusivity coefficient, which is a function of surface air temperature gradient; surface air pressure; surface vapor pressure; and precipitable water. The zonal moisture transport is a function of the zonal surface specific humidity gradient; mean zonal wind velocity; surface air pressure; surface vapor pressure; precipitable water; vertical zonal wind gradient with respect to pressure, which is a function of surface temperature gradient; and a term which weights the transport from land to water or vice versa by the effective width of the latitude belt that is land. This will be described further later.

The moisture convergence in the air over land or sea is computed in the same way as the meridional and zonal energy convergence are computed in the CCSI model (Ledley, 1988 a & b). The moisture convergence is computed from the net moisture transports into the air over land or sea divided by the area of land or sea.

The precipitation rate in each latitude zone will be determined using a parameterization of zonal precipitation rates developed by Schneider and Thompson (1978). The parameterization includes contributions from the moisture convergence discussed above; evaporation-precipitation recycling within a single latitude zone, which is a function of the saturation vapor pressure; and a baroclinicity term which is a function of the temperature gradient and the saturation vapor pressure.
Theory and Numerical Method

1. Thermodynamic Energy Equation

The basic equation used to develop the parameterization of the hydrologic cycle is the time-averaged thermodynamic energy equation, averaged vertically through the depth of the atmosphere (Sellers, 1973).

\[
\bar{R} = \frac{1}{g} \int_{0}^{d} \frac{d}{dt} (Lq + c_p T + gh)dp + \int_{0}^{d} \frac{dT_E}{dt} dz
\]  \hspace{1cm} (1)

Where the bar represents a time average,

\[
\bar{x} = \frac{1}{t} \int_{0}^{t} x dt
\]

\( t = \) time interval, sec

\( R = \) net available radiation for a column from the top of the atmosphere to a depth \( d \) in the soil or water, below which the vertical energy flux is negligible = net solar radiation minus net terrestrial radiation, W m\(^{-2}\)

\( g = \) acceleration of gravity \( \equiv 9.8 \) m sec\(^{-2}\)

\( p_0 = \) surface pressure = 1000 mb = 10\(^5\) Pa

\( L = \) latent heat of condensation = 2.5 \times 10\(^6\) J kg\(^{-1}\)

\( q = \) specific humidity

\( c_p = \) specific heat of air at constant pressure = 1004 J kg\(^{-1}\) (°K\(^{-1}\))

\( T = \) air temperature, °K

\( h = \) height above sea level, m

\( C = \) heat capacity of soil or water times the density, J m\(^{-3}\) (°K\(^{-1}\))

\( T_E = \) soil or water temperature, °K

"0" subscript refers to surface values
The three energy terms in the first integral on the right-hand side of (1) represent the latent heat (Lq), sensible heat (c_pT) and potential energy (gh). The latent heat term in this equation, as parameterized by Sellers (1973) will be used to compute moisture transport. The derivation of this parameterization follows.

The variable $s$ is used to represent any of the three terms Lq, c_pT or gh. Each variable in (1) can be expressed as the sum of a mean value and a deviation: $s = \bar{s} + s'$. Employing this notation to each of the total derivatives under the first integral on the right-hand side of (1), the continuity equation in pressure coordinates, and noting that x, y, and p are the eastward, northward and vertical coordinates

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial \omega}{\partial p} = 0$$

we get:

$$\frac{d\bar{s}}{dt} = \frac{\partial \bar{s}}{\partial t} + u\frac{\partial \bar{s}}{\partial x} + v\frac{\partial \bar{s}}{\partial y} + \omega\frac{\partial \bar{s}}{\partial p}$$

$$= \frac{\partial \bar{s}}{\partial t} + \frac{\partial u\bar{s}}{\partial x} + \frac{\partial v\bar{s}}{\partial y} + \frac{\partial \omega\bar{s}}{\partial p} - \bar{s} \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial \omega}{\partial p}\right)$$

$$= \frac{\partial \bar{s}}{\partial t} + \frac{\partial}{\partial x}(\bar{u}\bar{s} + u's') + \frac{\partial}{\partial y}(\bar{v}\bar{s} + v's') + \frac{\partial}{\partial p}(\bar{\omega}\bar{s} + \omega's')$$

In considering the grid shape (10° latitude x 360° wide zone which is divided into one land grid box and one sea grid box), it is clear that dx is much greater than dy, so $\bar{u}\bar{s}'/dx$ is relatively small compared to the other terms and therefore is neglected.

$$\Rightarrow \quad \frac{d\bar{s}}{dt} = \frac{\partial \bar{s}}{\partial t} + \frac{\partial}{\partial x}(\bar{u}\bar{s} + u's') + \frac{\partial}{\partial p}(\bar{\omega}\bar{s} + \omega's')$$

(2)
where: \( s = \) latent heat \((L_q)\), sensible heat \((c_p T)\) or potential energy \((gh)\)

\(u, v\) and \(\omega\) = east-west, north-south and vertical wind velocity components (the vertical wind velocity is in pressure units, Pa s\(^{-1}\))

The eddy north-south transport term, \(\overline{\nu' s'}\), can be parameterized using a standard eddy diffusivity parameterization

\[
\overline{\nu' s'} = -K_S \frac{\partial s}{\partial y}
\]

where \(K_S\) is the eddy diffusivity coefficient.

Integrating (2) over the depth of the atmosphere we get:

\[
\frac{1}{g} \int_0^\infty \frac{ds}{dt} \, dp = \frac{1}{g} \int_0^\infty \frac{\partial s}{\partial t} \, dp + \frac{1}{g} \int_0^\infty \frac{\partial}{\partial x} \overline{\nu s} \, dp + \frac{1}{g} \int_0^\infty \frac{\partial}{\partial y} \left( \overline{\nu s} - K_S \frac{\partial s}{\partial y} \right) \, dp + \frac{1}{g} \int_0^\infty \frac{\partial}{\partial p} (\overline{\omega s'} + \overline{\omega' s'}) \, dp
\]

A B C D

Here term A is the heat storage; terms B, C are the east-west, and north-south, heat (latent heat, sensible heat, potential energy) transports respectively; and term D is the vertical heat transport. Term C is divided into two parts representing the mean meridional heat transport and the eddy heat transport.

In order to use these equations to determine energy transport, it is necessary to parameterize them in terms of the surface values. This parameterization follows Sellers (1973). Note, "\(0\)" subscript refers to surface values:

\[
G_S = \frac{1}{g} \int_0^\infty \frac{\partial s}{\partial t} \, dp = \frac{1}{g} a_{CS} \frac{\partial s_0}{\partial t} p_0
\] (3)

\[
\Delta x S = \frac{1}{g} \int_0^\infty \frac{\partial}{\partial x} \overline{\nu s} \, dp = \frac{1}{g} \int_0^\infty \frac{\partial s}{\partial x} \, dp = \frac{1}{g} a_{xS} \overline{\nu_0 s_0} \frac{\partial s_0}{\partial x} p_0 \Delta L'
\] (4)
\[ \Delta y S = \frac{1}{g} \int_{0}^{\rho_0} \frac{\partial}{\partial y} \left( \bar{v} \bar{s} - K_s \frac{\partial \bar{s}}{\partial y} \right) dp = \frac{1}{g} \int_{0}^{\rho_0} \left( \bar{v} \bar{s} - K_s \frac{\partial \bar{s}}{\partial y} \right) dp \]  \tag{5}

\[ M_S = \frac{1}{g} \int_{0}^{\rho_0} \bar{v} \bar{s} \ dp = \frac{1}{g} a_{MS} \bar{v}_0 \bar{s}_0 \rho_0 \]  \tag{6}

\[ E_S = - \frac{1}{g} \int_{0}^{\rho_0} K_s \frac{\partial \bar{s}}{\partial y} dp = - \frac{1}{g} a_{ES} K_s \frac{\partial \bar{s}_0}{\partial y} \rho_0 \]  \tag{7}

\[ S = E_S + M_S \]  \tag{8}

\[ \frac{1}{g} \int_{0}^{\rho_0} \frac{\partial}{\partial p} \left( \bar{\omega} \bar{s} + \bar{\omega} \bar{s} \right) dp = 0 \]  \tag{9}

In equation (4), \( A_{L'} = A_L / A_{LM} \), where \( A_L \) is the fraction of a given latitude belt occupied by land and \( A_{LM} \) is the fraction of the belt occupied by the largest single land mass. This factor varies with the number of land-sea boundaries in a zonal direction in each latitude belt, and is the effective width of the belt. It accounts for the larger zonal land-sea transport if there are a large number of zonal land-sea boundaries. The value of \( A_{L'} \) given by Robock (1977) is always between 1 and 2.33 and is listed in Table 1.

The coefficients in (3), (4), (6), (7) relate the vertical integrals of temperature, wind velocity and humidity to their surface values. They are discussed in the next section.
Table 1. Parameter values (A_L' from Robock 1977; a and b from Sellers 1973)

<table>
<thead>
<tr>
<th>Latitude belt</th>
<th>A_L'</th>
<th>Latitude circle</th>
<th>a (10^-5 m^-1)</th>
<th>b</th>
</tr>
</thead>
<tbody>
<tr>
<td>80-90N</td>
<td>1.00</td>
<td>80N</td>
<td>0.62</td>
<td>1.16</td>
</tr>
<tr>
<td>70-80N</td>
<td>1.75</td>
<td>70N</td>
<td>1.94</td>
<td>1.10</td>
</tr>
<tr>
<td>60-70N</td>
<td>1.53</td>
<td>60N</td>
<td>2.63</td>
<td>0.96</td>
</tr>
<tr>
<td>50-60N</td>
<td>1.54</td>
<td>50N</td>
<td>2.26</td>
<td>0.86</td>
</tr>
<tr>
<td>40-50N</td>
<td>1.36</td>
<td>40N</td>
<td>1.99</td>
<td>0.95</td>
</tr>
<tr>
<td>30-40N</td>
<td>1.67</td>
<td>30N</td>
<td>1.73</td>
<td>1.10</td>
</tr>
<tr>
<td>20-30N</td>
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<td>20N</td>
<td>1.42</td>
<td>1.19</td>
</tr>
<tr>
<td>10-20N</td>
<td>1.29</td>
<td>10N</td>
<td>1.10</td>
<td>1.17</td>
</tr>
<tr>
<td>0-10N</td>
<td>1.60</td>
<td>0</td>
<td>0.99</td>
<td>1.16</td>
</tr>
<tr>
<td>0-10S</td>
<td>1.75</td>
<td>10S</td>
<td>0.94</td>
<td>1.17</td>
</tr>
<tr>
<td>10-20S</td>
<td>2.33</td>
<td>20S</td>
<td>0.99</td>
<td>1.19</td>
</tr>
<tr>
<td>20-30S</td>
<td>2.00</td>
<td>30S</td>
<td>0.73</td>
<td>1.10</td>
</tr>
<tr>
<td>30-40S</td>
<td>2.00</td>
<td>40S</td>
<td>0.36</td>
<td>0.95</td>
</tr>
<tr>
<td>40-50S</td>
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<td>50S</td>
<td>0.31</td>
<td>0.86</td>
</tr>
<tr>
<td>50-60S</td>
<td>1.00</td>
<td>60S</td>
<td>0.31</td>
<td>0.96</td>
</tr>
<tr>
<td>60-70S</td>
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<td>1.16</td>
</tr>
<tr>
<td>80-90S</td>
<td>1.00</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* Note, "a" and "b" discussed later.
2. Vertical Profiles

In order to evaluate the a coefficients in (3), (4), (6) and (7), vertical profiles of temperature, wind speed, and humidity were assumed, in a manner similar to Saltzman and Vernekar (1971):

\[ \bar{T} = T_0 - (p_0 - p) \frac{\partial \bar{T}}{\partial p} \quad (10) \]

\[ \bar{u} = u_0 - (p_0 - p) \frac{\partial \bar{u}}{\partial p} \quad (11) \]

\[ \bar{v} = v_0 \left( \frac{2p}{p_0} - p_0 \right) \quad (12) \]

\[ \bar{q} = q_0 (p/p_0)^{a_1} \quad (13) \]

It is assumed that (i) \( \partial \bar{T}/\partial p = 0.12 \, ^\circ \text{K} \, \text{mb}^{-1} \) (0.0012 \( ^\circ \text{K} \, \text{Pa}^{-1} \)), a value similar to that of Saltzman and Vernekar (1971), and (ii) except at the equator

\[ \frac{\partial \bar{u}}{\partial p} = R_d \frac{\partial \bar{T}_0}{f \partial y} \quad (14) \]

where \( p_a \) is 500 mb (5 x 10^4 Pa), \( f \) is the Coriolis parameter, \( R_d \) is the gas constant for dry air which is 287.04 J kg\(^{-1}\) (\( ^\circ \text{K} \))\(^{-1}\). At the equator it is assumed that \( \partial \bar{u}/\partial p = 0 \).

In (13), the exponent \( a_1 \) may be determined by integrating (13) from \( p_0 \) to the top of the atmosphere (see Appendix A for derivation). This gives

\[ 1 + a_1 = \frac{0.622 \bar{w}_0}{g \bar{w}_a} \quad (15) \]

The term \( \bar{w}_a \) is the precipitable water in an atmospheric column in cm, and \( \bar{w}_0 \) is the surface vapor pressure in dyne cm\(^{-2}\) (Equation 15 is a parameterization that was derived using CGS units). \( \bar{w}_0 \) is computed using the saturation vapor pressure as follows,
\( \bar{e}_0 = p_0 \bar{q}_0 / 0.622 = \bar{r} \bar{e}_s \), where \( r \) is the relative humidity at the surface, specified to 0.74 everywhere, and \( \bar{e}_s \) is the saturation vapor pressure at the temperature \( \bar{T}_0 \).

The a coefficients may now be determined by introducing (10) - (13) into (3), (4), (6) and (7) and performing the energy transport integration. For the parameterization of the hydrologic cycle only the latent heat term, \( s = Lq \), is of interest, so only the latent heat related a coefficients are shown. The derivations of these coefficients are given in Appendix B. The coefficients are

\[
\begin{align*}
a_{GV} & = \frac{1}{1 + a_1} \\
a_{xV} & = \frac{1}{1 + a_1} \left( 1 - \frac{1}{2 + a_1} \frac{p_0}{\bar{u}_0} \frac{\partial \bar{u}}{\partial p} \right) \\
a_{MV} & = \frac{a_1}{(1 + a_1) (2 + a_1)} \\
a_{EV} & = \frac{1}{1 + a_1}
\end{align*}
\]

3. Dynamical Fluxes

(1) Mean wind fluxes

To evaluate (4) and (6), it is necessary to calculate the surface values of the time average horizontal wind components, \( \bar{u}_0 \) and \( \bar{v}_0 \).

By assuming that the real forces acting on the atmosphere are the Coriolis force, pressure gradient force, gravitation force, and friction force (Holton, 1979), Newton's second law can be written as follows:

\[
\frac{dU}{dt} = -2\Omega \times U - \frac{1}{\rho} \nabla p + g + F_r
\]
where $F_r$ is the friction force. This equation can be expanded into its three coordinate directions, and then through scale analysis some of the terms can be neglected (Holton, 1979). The remaining terms form a simplified set of the momentum equations. The two equations for the x and y coordinates are the horizontal momentum equations:

$$fv - \frac{1}{\rho} \frac{\partial p}{\partial x} + F_{rx} = 0$$

(21)

$$fu + \frac{1}{\rho} \frac{\partial p}{\partial y} - F_{ry} = 0$$

(22)

which states that there is a balance between the Coriolis force, pressure gradient force and frictional force. The frictional terms can be written as:

$$F_{rx} = \alpha_0 \frac{\partial \tau_{0x}}{\partial z}$$

(23)

$$F_{ry} = \alpha_0 \frac{\partial \tau_{0y}}{\partial z}$$

(24)

where $\alpha_0$ is the surface specific volume and is equal to $1/\rho$, and $\tau_{0x}$ and $\tau_{0y}$ are, respectively, the eastward and northward components of the frictional stress, assumed to vanish at the top of the friction layer.

Since in this parameterization, the grid shape is such that $dx$ is much larger than $dy$ the $\partial p/\partial x$ term is relatively small and can thus be neglected. Considering this, insert (23) and (24) into (21) and (22), and employ the time average. The result of this are the equations of motion applicable to this problem

$$\bar{fv} + \alpha_0 \frac{\partial \tau_{0x}}{\partial z} = 0$$

(25)
\[ fu_0 - \alpha_0 \frac{\partial \tau_{0y}}{\partial z} + \alpha_0 \frac{\partial p_0}{\partial y} = 0 \]  

(26)

From micrometeorological theory, assuming a logarithmic wind profile near the ground (Sellers, 1973), it follows that

\[ -\alpha_0 \frac{\partial \tau_{0x}}{\partial z} = a |\overline{u_0}| \bar{u}_0 \]  

(27)

\[ -\alpha_0 \frac{\partial \tau_{0y}}{\partial z} = a |\overline{v_0}| \bar{v}_0 \]  

(28)

where

\[ a = \frac{g}{\alpha_0 \Delta p} k^2 (\ln \frac{z}{z_0})^{-2} \]  

(29)

\( \Delta p \) is the pressure thickness of the friction layer, assumed to equal 100 mb (10^4 Pa); \( k \) is the von Karman constant, assumed to equal 0.4; and \( z_0 \) is the roughness length (this term is a measure of the roughness of the surface, such as considering the vegetation, buildings, etc. on the land, waves on the ocean), assumed to equal 1 m over land and 0.0001 m over water. \( z \) is the height of surface values and is assumed to equal 10 m. Then for each latitude circle, a weighted average of \( a \) is calculated, depending on the portions of the circle that are land-land, land-sea, or sea-sea boundaries (see Table 1 for values).

The pressure gradient term in (26) is the small difference between two large terms of the same sign. That is, from the equation of state \( p_0 \alpha_0 = R_d T_0 \),

\[ \frac{\partial \overline{p_0}}{\partial y} = R_d \frac{\partial \overline{T_0}}{\partial y} - \frac{\partial \overline{\alpha_0}}{\partial y} = R_d \frac{\partial \overline{T_0}}{\partial y} - bR_d \frac{\partial \overline{T_0}}{\partial y} = R_d (1 - b) \frac{\partial \overline{T_0}}{\partial y} \]  

(30)

in which

\[ b = \frac{\overline{T_0}}{\alpha_0} \frac{\partial \overline{\alpha_0}}{\partial \overline{T_0}} \]  

(31)

Values of \( b \) were determined by Sellers (1973) partly from the latitudinal variation of the zonally averaged sea-level pressure given by Mintz (1968) and are listed in Table 1.
Equations (25) and (26) now become:

\[ f\bar{v}_0 - a \bar{u}_d \bar{u}_0 = 0 \]  \hspace{1cm} (32)

\[ f\bar{u}_0 + a \bar{v}_d \bar{v}_0 + R_d (1 - b) \frac{\partial T_0}{\partial y} = 0 \]  \hspace{1cm} (33)

Solving for \( \bar{u}_0 \) and \( \bar{v}_0 \) gives

\[ \bar{u}_0 = 0, \quad \bar{v}_0 = \left( \frac{R_d (b-1) \frac{\partial T_0}{\partial y}}{a} \right)^{0.5} \]  \hspace{1cm} (34)

\[ \bar{u}_0 = \frac{R_d (b-1) \frac{\partial T_0}{\partial y}}{f (1 + \left| \frac{a u_0^3}{f} \right|)}, \quad \bar{v}_0 = \frac{a}{f} \bar{u}_d \bar{u}_0 \]  \hspace{1cm} \text{elsewhere.}  \hspace{1cm} (35)

Equation (35) for \( \bar{u}_0 \) is solved using the half-interval iteration method (Carnahan, 1976), since \( \bar{u}_0 \) appears on both sides of the equation. In order to increase the stability of the computations, the values of \( \bar{u}_0 \) and \( \bar{v}_0 \) obtained are smoothed latitudinally using a 3-point (1-2-1) binomial smoothing (Sellers, 1973).

(2) Moisture transport

The latent heat transport by mean meridional motions, \( M_v \), eddy meridional motions, \( E_v \), and mean zonal motions \( \Delta_x S \) can be written from equations (6), (7), (4) respectively by substituting for the a coefficients, equations (17) - (19). These equations are as follows

\[ M_v = -\frac{a_1}{g(1 + a_1)(2 + a_1)} \bar{v}_0 L \bar{q}_0 \nu_0 \]  \hspace{1cm} (36)

\[ E_v = -\frac{1}{g(1 + a_1)} K_v \frac{\partial L \bar{q}_0}{\partial y} \nu_0 \]  \hspace{1cm} (37)

\[ \Delta_x S = \frac{1}{g(1 + a_1)} \left( 1 - \frac{1}{(2 + a_1) \nu_0 \frac{d\nu_0}{dp}} \right) \nu_0 \frac{\partial L \bar{q}_0}{\partial x} \nu_0 A \nu' \]  \hspace{1cm} (38)
Following evidence presented by Sellers (1973), the eddy diffusivity coefficients, $K_v$ for latent heat are made proportional to the temperature gradient:

$$K_v = 2.5 \times 10^5 \Delta T_0 \text{ m}^2 \text{ sec}^{-1}$$

(39)

where $\Delta T_0$ is the surface air temperature difference between successive $10^\circ$ latitude belts. For each grid box there are four possible gradients: land-land, sea-sea, sea-land, and land-sea and so there are four values of $K_v$. These values are binomial smoothed in the same manner as the velocity components and then used in equation (37).

4. Precipitation

The parameterization of zonal precipitation rate is from Schneider and Thompson (1978). They developed various regression equations of varying complexity. The equation used in this study decomposes the precipitation rate into contributions by latent heat convergence in a zone, evaporation-precipitation recycling within a single latitude zone, and baroclinic eddies (see Appendix D for definition).

Precipitation rate = precipitation due to latent heat convergence in a zone + precipitation due to evaporation-precipitation recycling within a zone + precipitation due to baroclinic eddies

The latent heat convergence into a latitude zone is determined by

$$L_H = \frac{F_2 \cos \phi_2 - F_1 \cos \phi_1}{a (\sin \phi_2 - \sin \phi_1)}$$

(40)

where $F_1, F_2$ are latent heat transports (i.e. $M_v + E_v$) across the adjacent latitude circles $\phi_1, \phi_2$ (energy per unit length per unit time). $\phi_2$ is southern most for positive-northward transports. $a$ is the radius of the Earth. $L_H$ (energy per unit area per unit time) can be
converted to water equivalent assuming 1 gram of precipitation = 2470 J of latent heat release, or for $L_H$ in W m$^{-2}$

$$p_L = 0.0350 L_H \text{ (mm d}^{-1}\text{)}$$ (41)

The derivation of this is given in Appendix C.

Schneider and Thompson (1978) developed seven regression equations of varying complexity. Table 2 shows these equations and how well the parameterized precipitation rate correlated with observation globally, and in each hemisphere. The first criterion in choosing the regression equation in this parameterization is a high correlation coefficient, and secondly consistency between the two hemispheres. Equations 5G, 6G and 7G fit these criteria. The increase in the correlation coefficient for the increase in the complexity in the equation from 5G to 7G does not seem significant. Therefore equation 5G was chosen for this study.

$$\text{precipitation rate} = c_1 + c_2 p_L + c_3 e_s + c_4 |\Delta T| e_s \text{ (mm d}^{-1}\text{)}$$ (42)

where $c_1 = 0.01$

$c_2 = 1.06$

$c_3 = 0.109$

$c_4 = 0.0138$

$p_L$ = precipitation rate resulting from latent heat convergence, mm day$^{-1}$

$T$ = surface air temperature, °K

$e_s$ = saturation vapor pressure, mb

$\Delta T$ = temperature gradient, °K/1000 km

$|\Delta T|$ computed by $|T_{+1} - T_{-1}| / dy$ (centered difference),
$T_{+1}, T_{-1} =$ temperature at latitudes 20° apart bordering the point at which the gradient is determined

d is length over which the gradient is determined in units of 1000 km

$|\Delta T|$ can be written as $S |T_{+1} - T_{-1}|,$

where $S = 1/dy = 0.4496 \, (20^\circ \text{ latitude}/1000 \, \text{km})$
<table>
<thead>
<tr>
<th>Model Number</th>
<th>Regression Function</th>
<th>$R^2$</th>
<th>Annual** RMS Error Northern Hemisphere</th>
<th>Annual** RMS Error Southern Hemisphere</th>
<th>Seasonal*** RMS Error Northern Hemisphere</th>
<th>$c_1$</th>
<th>$c_2$</th>
<th>$c_3$</th>
<th>$c_4$</th>
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<td>1G</td>
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<td>0.861</td>
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$R^2$ refers to the correlation coefficient squared between the individual regression function results and the data from which it was derived. For "G," 17 global points are used and for "N," 9 northern hemisphere points plus the equator are used. Annual means are used also.

** These RMS precipitation rate errors (in mm day$^{-1}$) are not area weighted averages. They are calculated using latitudes 80°-10° inclusive.

*** These seasonal RMS precipitation rate errors (in mm day$^{-1}$) are not area weighted averages.

Note: G indicates an equation derived from global data while N indicates the equation was derived from Northern Hemisphere data.
III. RESULTS

This chapter will first evaluate the sensitivity of the parameterization of the hydrologic cycle to variations in the values of specified parameters, and then test the parameterization by comparing computed results with available observations and GCM results. Here northward and eastward are defined as positive.

A. Sensitivity of the Parameterization of the Hydrologic Cycle

Since the parameterization of the hydrological cycle involves making many simplifying assumptions, a number of sensitivity studies are performed to help identify the strengths and weaknesses of the parameterization. These studies include tests of the parameterization sensitivity to surface air temperature, zonal wind, mean meridional moisture flux, and relative humidity.

Sensitivity to Surface Air Temperature

The hydrologic cycle parameterization is dependent on the surface air temperature in all latitude belts over land and ocean. Eventually the surface air temperature will be computed in the CCSI model; however in these sensitivity studies two sets of surface air temperatures are chosen to test the parameterization. These are 1) surface air temperatures computed by CCSI model (Ledley, 1988 a&b) in which the precipitation rate was specified to 0.4 m yr\(^{-1}\) (1.268 x 10\(^{-8}\) ms\(^{-1}\)) (referred to as CCSI model) and 2) observations from Schutz and Gates (1974) (referred to as S & G data) (Figures 2 - 3).

Figures 2 - 3 show surface air temperatures are higher in the tropics region and lower toward polar region. In Northern Hemisphere, during winter DJF, the temperatures are the lowest of the year, and in the zone closest to the north pole reach the lowest point of
Figure 2. Seasonally averaged surface air temperature over land in °K data from Schutz and Gates (1974) (solid line) and computed by CCSI model with precipitation rate specified to 1.268 x 10^-8 ms^-1 (dashed line) for: (a) December, January and February (referred to as DJF), (b) March, April and May (MAM), (c) June, July and August (JJA), (d) September, October and November (SON).
Figure 3. Seasonally averaged surface air temperature over ocean in °K data from Schutz and Gates (1974) (solid line) and computed by CCSI model with precipitation rate specified to $1.268 \times 10^{-8}$ m s$^{-1}$ (dashed line) for: (a) DJF, (b) MAM, (c) JJA, (d) SON.
about 240°K over both land and ocean. During the summer JJA, temperatures are the highest of the year, and in the zone closest to the north pole reach the highest point of about 270°K for both land and ocean. The other two seasons (spring MAM, fall SON) are between these two cases. In the Southern Hemisphere, the situation is reversed, during summer DJF, the temperatures are the highest of the year, with the south pole region reaching the highest point of about 245°K over land and 260°K over ocean. During the winter JJA, temperatures are the lowest of the year, with south pole region reaching the lowest temperature of about 220°K over land and 240°K over ocean. The other two seasons (spring SON, fall MAM) are between these two cases. Also note that around north pole zone land and ocean air temperatures are almost same, while around south pole zone ocean surface air temperatures are always warmer by about 20°K than land surface air temperatures during the year. This can be understood in part by examining Figure 4 (sea fraction distribution). Around the north pole, the land fraction is very small, and surface air temperatures are strongly affected by the ocean which has a higher heat capacity than land. Around the south pole, the land fraction is much larger than the sea fraction, so land surface air temperatures are almost not affected by the ocean. In addition, the land surface at the south pole is at a much higher elevation than the land in the poleward most zone in the Northern Hemisphere. Therefore the surface air temperatures are colder over the land near the south pole than over the land near the north pole.

Using these temperatures the parameterization computes zonal winds, meridional winds, mean meridional moisture fluxes, eddy meridional moisture fluxes, and precipitation rates over land and ocean. In the following paragraphs the effect of variations in temperature on some of these variables will be discussed.
Figure 4. Sea fraction latitude distribution from CCSI model (Ledley, 1988a&b).
Figures 5 - 6 give the seasonally and zonally averaged zonal wind and meridional wind respectively for the four seasons. These two figures show the general atmospheric circulation pattern. In the region 0°-30°, the wind is easterly (westward, Figure 5), and equatorward (Figure 6). In the region 30°-60°, the wind is westerly (eastward, Figure 5), and also blows toward the poles (Figure 6). In the region 60°-90°, the wind is easterly (Figure 5), and equatorward again. The results from the two sets of input temperatures (CCSI model and S & G data) coincide fairly well in latitudinal and seasonal distribution, however there are several discrepancies. These can be explained by differences in the temperature gradient, upon which the wind speeds are most dependent. The following are the equations calculating zonal and meridional wind from Chapter II.

\[ \bar{u}_0 = 0, \quad \bar{v}_0 = \left( \frac{R_d (b - 1) \frac{\partial T_0}{\partial y}}{a} \right)^{0.5} \text{ at equator,} \]  

(34)

\[ \bar{u}_0 = \frac{R_d (b - 1) \frac{\partial T_0}{\partial y}}{f (1 + \frac{|\bar{u}_0|}{f})}, \quad \bar{v}_0 = \frac{a}{f} \bar{u}_0 \bar{u}_0 \text{ elsewhere.} \]  

(35)

Except at the equator, zonal wind is dependent mainly on the meridional temperature gradient; the larger the temperature gradient, the larger the zonal wind will be. This will result in a larger meridional wind. This can be seen clearly by comparing Figure 5, 6 and 7 as discussed below.

Figure 7 gives the seasonally and zonally averaged surface air temperature difference between two adjacent latitudes (north boundary value minus south boundary value) for four seasons. Since temperature is highest at equator, the temperature difference, \(\Delta T\), is positive in Southern Hemisphere and negative in Northern Hemisphere. Around 70°S the temperature difference is very large, because in that region the land fraction
Figure 5. Seasonally and zonally averaged zonal wind in m s\(^{-1}\) computed using surface air temperature computed by CCSI model with precipitation rate specified to 1.268 × 10\(^{-8}\) m s\(^{-1}\) (dashed line) and using data from Schutz and Gates (1974) (Solid line) for: (a) DJF, (b) MAM, (c) JJA, (d) SON.
Figure 6. Seasonally and zonally averaged meridional wind in ms\(^{-1}\) computed using surface air temperature computed by CCSI model with precipitation rate specified to 1.268 x 10\(^{-8}\) ms\(^{-1}\) (dashed line) and using data from Schutz and Gates (1974) (Solid line) for: (a) DJF, (b) MAM, (c) JJA, (d) SON.
Figure 7. Seasonally and zonally averaged temperature difference between two adjacent latitudes in °K for surface air temperature 1) data from Schutz and Gates (1974) (solid line), 2) computed by CCSI model with precipitation rate specified to $1.268 \times 10^{-8}$ m s$^{-1}$ (dashed line). (a) DJF, (b) MAM, (c) JJA, and (d) SON.
increases dramatically moving poleward. Thus, the temperature changes from being strongly affected by ocean with its high thermal inertia to being strongly affected by land with its low thermal inertia.

In DJF, the meridional temperature differences (referred to as ΔT) from the CCSI model's surface air temperatures are larger around 70°S than for the S & G temperatures, Figures 5a and 6a show a correspondingly larger zonal wind speeds, u, and meridional wind speeds, v. Between about 40°S-10°S and 30°N-60°N, ΔT from the CCSI model temperatures are are smaller than for the S & G data, and thus u and v are also smaller. Between 10°S-20°N and 60°N-80°N, ΔT from the CCSI model temperatures are larger than for the S & G data, and thus u and v are also larger. For both sets of temperature data, the meridional temperature differences across 20°N are larger than across 20°S, this causes a relatively larger wind speed at 20°N than 20°S.

In MAM (Figures 5b and 6b), ΔT from the CCSI model temperatures are smaller than for the S & G data between 40°S and 10°S, and so are u and v. At the equator, the meridional wind speed is determined directly from the temperature difference, which is -0.53°K between 5°N and 5°S in the CCSI model, however, using S & G data the temperature difference between 5°N and 5°S is 0.26°K. Therefore the meridional wind speed is different at equator for the two sets of input temperatures, (Figure 6b). The magnitudes of the temperature difference, ΔT, values computed from the CCSI model are larger than those computed from S & G data between 0° and 30°N and at 50°N, and the same is true of the u and v magnitudes.

In JJA (Figures 5c and 6c), the major differences in two sets zonal and meridional wind occur at 20°S and 0°-30°N. At 20°S, ΔT from the CCSI model is smaller than from S
& G data, and so are u and v. At 20°N, \( \Delta T \) from the CCSI model is larger than from the S & G data, and so are u and v. For both sets of input temperatures, the meridional temperature differences across 20°N are smaller than across 20°S, this causes a relatively smaller wind speed at 20°N than 20°S.

In SON (Figures 5d and 6d), \( \Delta T \) from the CCSI model are larger compared those from the S & G data at 20°N, and so u and v are also larger.

Considering the zonal and meridional wind have been binomial smoothed, the above analysis explains the two set wind results quite well.

In magnitude, the greatest difference in \( \Delta T \) between two adjacent latitudes, 1.4°K, at 10°N in MAM results in zonal wind difference of 0.64 m s\(^{-1}\), and meridional wind difference of 1 m s\(^{-1}\).

Figure 8 show the precipitation rates seasonally and zonally averaged over land and ocean for four seasons for the input temperatures from the CCSI model (dashed line) and from S & G data (solid line). The two sets precipitation rates fit fairly well in latitudinal and seasonal distribution, however there are several discrepancies. Major differences occur between 0°-45°N during DJF; in Northern Hemisphere during MAM; 75°S, 15°S and 20°N-60°N during JJA; 75°S, 15°S, tropics and 20°N-60°N during SON. The differences are associated with the differences in the surface air temperatures and temperature gradients in the same regions and seasons. These differences produce changes in latent heat convergence, evaporation-precipitation recycling in a zone, and baroclinic eddies, resulting in differences in precipitation rates.

In magnitude, the greatest temperature difference, 3.87°K over land, and 3.32°K over ocean at 15°N in MAM results in a difference in precipitation rate of 1.9 x 10\(^{-8}\) m s\(^{-1}\)
(0.6 m yr\(^{-1}\)); the greatest difference in \(\Delta T\) between two adjacent latitudes, 1.4\(^\circ\)K, at 10\(^\circ\)N in MAM results in a difference in precipitation rate of 1.39 \(\times\) 10\(^{-8}\) m s\(^{-1}\) (0.44 m yr\(^{-1}\)).

In summary, the precipitation rate is somewhat sensitive to surface air temperature changes, but the sensitivity is well within the natural variability of precipitation rate.
Figure 8. Precipitation rate seasonally and zonally averaged over land and ocean in ms$^{-1}$ computed using surface air temperature computed by CCSI model with precipitation rate specified to 1.268 x 10$^{-8}$ ms$^{-1}$ (dashed line) and using data from Schutz and Gates (1974) (Solid line) for: (a) DJF, (b) MAM, (c) JJA, (d) SON.
Sensitivity to Surface Air Temperature Seasonal Amplitude and Mean Value

In this experiment set, two group experiments are performed, group one with surface air temperature seasonal amplitude changed, group two with surface air temperature mean value changed. The run with surface air temperature data from Schutz and Gates (1974) is hereafter referred to as the control case.

a. Sensitivity to surface air temperature seasonal amplitude

In this group two experiments are performed, one reducing the seasonal amplitude of control case surface air temperature by 0.5 times, and second enhancing the seasonal amplitude of control case surface air temperature by 1.5 times.

The seasonal cycle of these two surface air temperatures and control case temperature over land and over ocean are shown in Figures 9 and 10 respectively for 85°S, 5°N, 45°N, and 85°N. We can see the differences between control case and seasonal amplitude changed cases decrease toward the equator, and for all latitudes, the largest differences occur during DJF and JJA, with almost no changes during MAM and SON.

Figures 11 - 12 give the surface air temperatures versus latitude for four seasons over land and over ocean respectively, for the control and modified amplitude cases. The surface air temperature seasonal amplitude changes have very small impact during MAM and SON. However in DJF, the reduced seasonal amplitude produces a surface air temperature decrease in Southern Hemisphere and an increase in Northern Hemisphere over land and ocean; while the enhanced seasonal amplitude produces a surface air temperature increase in Southern Hemisphere and a decrease in Northern Hemisphere over
Figure 9. Seasonal cycle of surface air temperature over land in °K data from Schutz and Gates (1974) (solid line), reduce seasonal amplitude by 0.5 times case (dotted line) and enhance seasonal amplitude by 1.5 times case (dashed line) at 85°N, 45°N, 5°N and 85°S.
Figure 10. Seasonal cycle of surface air temperature over ocean in °K data from Schutz and Gates (1974) (solid line), reduce seasonal amplitude by 0.5 times case (dotted line) and enhance seasonal amplitude by 1.5 times case (dashed line) at 85°N, 45°N, 5°N and 85°S.
Figure 11. Seasonally averaged surface air temperature over land in °K data from Schutz and Gates (1974) (solid line), reduce seasonal amplitude by 0.5 times case (dotted line) and enhance seasonal amplitude by 1.5 times case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
Figure 12. Seasonally averaged surface air temperature over ocean in °K data from Schutz and Gates (1974) (solid line), reduce seasonal amplitude by 0.5 times case (dotted line) and enhance seasonal amplitude by 1.5 times case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
land and ocean. The largest changes occur in the polar regions, with a change of about $5^\circ$K. The effects decrease toward the equator. In JJA, the situation is reversed, with an increase in Southern Hemisphere and a decrease in Northern Hemisphere surface air temperature for the reduced seasonal amplitude case, and a decrease in Southern Hemisphere and an increase in Northern Hemisphere for the enhanced seasonal amplitude case.

Figures 13 - 16 give the three resulting precipitation rates for each season, zonally averaged over land and ocean. Each figure also gives the three components of the precipitation rate. These three components are: (1) precipitation rate due to latent heat convergence in a zone (Figures 13b-16b), (2) precipitation rate due to evaporation-precipitation recycling within a zone (Figures 13c-16c), (3) precipitation rate due to baroclinic eddies (Figures 13d-16d). From these figures, it is clear that evaporation-precipitation recycling play very important role in the tropics, baroclinic eddies are important in mid-latitudes, and latent heat convergence defines the precipitation distribution mostly in the tropics but also globally.

Generally precipitation rates change in almost the same patterns as the surface air temperatures change in corresponding regions and seasons, with similar patterns of increase and decrease during DJF and JJA, and very small changes during MAM and SON. An analysis of each season follows.

In DJF (Figure 13a), precipitation rates increase and decrease in the same patterns as surface air temperatures in the two hemisphere. In the Southern Hemisphere smaller precipitation rates result from a reduced seasonal amplitude of temperature when air temperatures are lower than the control case, and larger precipitation rates result from an
Figure 13. Precipitation rate seasonally and zonally averaged over land and ocean for DJF in m/s$^{-1}$ computed using surface air temperature data from Schutz and Gates (1974) (Solid line), reduce surface air temperature seasonal amplitude by 0.5 times case (dotted line) and enhance surface air temperature seasonal amplitude by 1.5 times case (dashed line): (a) total precipitation rate, (b) precipitation rate due to latent heat convergence in a zone, (c) precipitation rate due to evaporation-precipitation rate recycling within a zone, (d) precipitation rate due to baroclinic eddies.
Figure 14. Precipitation rate seasonally and zonally averaged over land and ocean for MAM in ms$^{-1}$ computed using surface air temperature data from Schutz and Gates (1974) (Solid line), reduce surface air temperature seasonal amplitude by 0.5 times case (dotted line) and enhance surface air temperature seasonal amplitude by 1.5 times case (dashed line): (a) total precipitation rate, (b) precipitation rate due to latent heat convergence in a zone, (c) precipitation rate due to evaporation-precipitation rate recycling within a zone, (d) precipitation rate due to baroclinic eddies.
Figure 15. Precipitation rate seasonally and zonally averaged over land and ocean for JJA in m s\(^{-1}\) computed using surface air temperature data from Schutz and Gates (1974) (Solid line), reduce surface air temperature seasonal amplitude by 0.5 times case (dotted line) and enhance surface air temperature seasonal amplitude by 1.5 times case (dashed line): (a) total precipitation rate, (b) precipitation rate due to latent heat convergence in a zone, (c) precipitation rate due to evaporation–precipitation rate recycling within a zone, (d) precipitation rate due to baroclinic eddies.
Figure 16. Precipitation rate seasonally and zonally averaged over land and ocean for SON in m$^{-1}$ computed using surface air temperature data from Schutz and Gates (1974) (Solid line), reduce surface air temperature seasonal amplitude by 0.5 times case (dotted line) and enhance surface air temperature seasonal amplitude by 1.5 times case (dashed line): (a) total precipitation rate, (b) precipitation rate due to latent heat convergence in a zone, (c) precipitation rate due to evaporation-precipitation rate recycling within a zone, (d) precipitation rate due to baroclinic eddies.
enhanced seasonal amplitude of temperature when air temperature are higher than the control case. The reversed pattern is seen in the Northern Hemisphere. The pattern in the tropics is determined by precipitation rate due to latent heat convergence (Figure 13b). Figure 13a also shows that the precipitation rates are larger in Southern Hemisphere than in Northern Hemisphere. This is associated with higher temperatures in Southern Hemisphere during DJF. In the poleward most regions there is a relatively larger temperature change, however the precipitation rate changes are relatively small. This is because the saturation vapor pressure $e_s$ is an exponential function of $(-1/T)$, so the smaller temperature $T$ is the smaller the exponential and thus the smaller $e_s$ is. Therefore when temperatures are low, small differences will not make much difference in the saturation vapor pressure. Since all three precipitation components are directly or indirectly the function of saturation vapor pressure, there is little impact of polar temperature changes on precipitation rate. This can be seen clearly in Figures 13b through 13d.

In MAM (Figure 14a), the largest changes occur in the tropics. This is because in the tropics, temperatures are higher, so even small change in temperature will have relatively large effect on saturation vapor pressure. This will result in large differences in precipitation rate due to latent heat convergence (Figure 14b) (which is also determined by temperature gradient) and precipitation rate due to evaporation-precipitation recycling within a zone (Figure 14c), and therefore the largest changes in the precipitation rate in MAM.

In JJA (Figure 15a), precipitation rates are larger in the Northern Hemisphere than in the Southern Hemisphere, the opposite of the situation in DJF. Major differences occur in the Northern Hemisphere, with the same patterns as the temperature changes. During JJA the Northern Hemisphere has the highest temperature during the year. Temperature differences in this season therefore produce large changes in saturation vapor pressure $e_s$,
thus causing larger differences in all three precipitation rate components (Figures 15b, 15c and 15d). Therefore, the changes in the amplitude of the surface air temperature seasonal cycle result in larger precipitation differences in Northern Hemisphere in this season. In the Southern Hemisphere the precipitation rates show no large differences because there are no large differences in the surface air temperatures in JJA.

In SON (Figure 16a), similar to MAM, with almost no changes.

b. Sensitivity to surface air temperature mean value

In this experiment group surface air temperature mean values are changed, one with reducing mean value of control case surface air temperature by 5°C, and second with increasing mean value of control case surface air temperature by 5°C.

Figures 17 and 18 show the surface air temperatures for these cases and the control case over land and over ocean respectively versus season for 85°S, 5°N, 45°N, and 85°N. Figures 19 and 20 show these temperatures over land and over ocean respectively versus latitude for each season. The surface air temperature mean value changes are always same for all latitudes and four seasons. The impact of these changes on the precipitation rate can be seen in Figure 21 as discussed below.

Figure 21 gives the precipitation rates computed from the temperature regimes described above seasonally and zonally averaged over land and ocean for each season. In each season, just like the temperature changes, precipitation rates also increase or decrease at all latitudes, but instead of changing the same magnitude at all latitudes, precipitation rate changes are larger when the precipitation rate is larger. This can be explained by looking at one season, for example Figure 22 gives the DJF precipitation rate (Figure 22a) and its three components. The temperature gradients remain the same for each of these three cases,
Figure 17. Seasonal cycle of surface air temperature over land in °K data from Schutz and Gates (1974) (solid line), reduce mean value by 5°K case (dotted line) and enhance mean value by 5°K case (dashed line) at 85°N, 45°N, 5°N and 85°S.
Figure 18. Seasonal cycle of surface air temperature over ocean in °K data from Schutz and Gates (1974) (solid line), reduce mean value by 5°K case (dotted line) and enhance mean value by 5°K case (dashed line) at 85°N, 45°N, 5°N and 85°S.
Figure 19. Seasonally averaged surface air temperature over land in °K data from Schutz and Gates (1974) (solid line), reduce mean value by 5°C case (dotted line) and enhance mean value by 5°C case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
Figure 20. Seasonally averaged surface air temperature over ocean in °K data from Schutz and Gates (1974) (solid line), reduce mean value by 5°C case (dotted line) and enhance mean value by 5°C case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
Figure 21. Precipitation rate seasonally and zonally averaged over land and ocean in ms\(^{-1}\) using surface air temperature data from Schutz and Gates (1974) (solid line), reduce surface air temperature mean value by 5°K case (dotted line) and enhance surface air temperature mean value by 5°K case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
Figure 22. Precipitation rate seasonally and zonally averaged over land and ocean for DJF in ms⁻¹ computed using surface air temperature data from Schutz and Gates (1974) (Solid line), reduce surface air temperature mean value by 5ºK case (dotted line) and enhance surface air temperature mean value by 5ºK case (dashed line): (a) total precipitation rate, (b) precipitation rate due to latent heat convergence in a zone, (c) precipitation rate due to evaporation-precipitation rate recycling within a zone, (d) precipitation rate due to baroclinic eddies.
so only the magnitude of the temperatures play an important role. Each of the three precipitation rate components is a function of saturation vapor pressure, which is an exponential function of temperature, so when temperature is higher a change in temperature will make larger difference in saturation vapor pressure. This results in larger differences in precipitation rate. The precipitation rate due to latent heat convergence is a function of the difference in the moisture transport across the boundaries of a zone. The larger that difference, i.e. the larger the magnitude of the convergence, the larger the change in precipitation rate (Figure 22b), but the effect is smaller than for evaporation-precipitation recycling. The term most sensitive to the changes in temperature is the evaporation-precipitation recycling within a zone, because it is directly a function of saturation vapor pressure (Figure 22c). The precipitation rate due to baroclinic eddies is a function of both saturation vapor pressure and temperature gradient. These combine to produce the largest changes in precipitation rate in mid latitudes (Figure 22d). Overall, the computed precipitation rates change the most in response to temperature change in regions where the temperature is the greatest, with smaller effects due to changes in latent heat convergence.

In summary, the changes in the surface air temperature seasonal amplitude can cause seasonal changes in precipitation rate, while changes in the mean surface air temperature can cause changes in the mean precipitation rate with the largest effect occurring when temperatures are relatively high.
Sensitivity to Mean Meridional Moisture Flux

In this experiment set, two experiments are performed using surface air temperature data from Schutz and Gates (1974). One with the mean meridional moisture flux computed as described in chapter II (control case), and a second in which mean meridional moisture flux is doubled.

Figure 23 shows these two sets of mean meridional moisture fluxes for the four seasons. The major differences between these two results occur in lower and mid latitude zones. The largest differences occur in tropics and northern subtropics in DJF, tropics and southern subtropics in JJA, tropics and the sub-tropics in both hemispheres in MAM and SON. These differences in mean meridional moisture fluxes produce large changes in latent heat convergence, and thus changes in precipitation rates due to latent heat convergence in the same regions and seasons (Figure 24). This results in large changes in the total precipitation rates in the corresponding regions and seasons (Figures 25).

In DJF, a large change in the mean meridional moisture flux gradient, when the flux is doubled, occurs between 10°S-10°N (Figure 23a). This causes about a 100% increase in moisture flux convergence around equator (Figure 24a), which results in an increase in precipitation rate of about 2 x 10^{-8} m s^{-1} (0.63 m yr^{-1}) around equator, about a 30% increase in the precipitation rate from the control case (Figure 25a). Another large change in mean meridional moisture flux gradient, when the flux is doubled, occurs between 20°N-40°N (Figure 23a), which cause about 100% decrease in moisture flux convergence around 30°N (Figure 24a). This results in a decrease in precipitation rate of about 2 x 10^{-8} m s^{-1} (0.63 m yr^{-1}) around 30°N, about a 95% decrease in the precipitation rate from the control case (Figure 25a). Similarly, for the other seasons, the large changes in mean meridional
moisture flux gradient cause corresponding large latent heat convergence changes, resulting large precipitation rate changes in same regions.
Figure 23. Seasonally and zonally averaged mean meridional moisture flux in W computed using surface air temperature data from Schutz and Gates (1974) for the control case (solid line), double mean meridional moisture flux case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
Figure 24. Precipitation rate due to latent heat convergence in a zone seasonally and zonally averaged over land and ocean in m s$^{-1}$ computed using surface air temperature data from Schutz and Gates (1974) for the control case (solid line), double mean meridional moisture flux case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
Figure 25. Precipitation rate seasonally and zonally averaged over land and ocean in ms$^{-1}$ computed using surface air temperature data from Schutz and Gates (1974) for the control case (solid line), double mean meridional moisture flux case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
Sensitivity to Zonal Wind Speed

In this experiment set, we change the zonal wind to 75% and 125% of the value computed in the control case. These experiments are initialized with surface air temperature data from Schutz and Gates (1974). These variations in the zonal winds are shown in Figure 26, and the resulting changes in the meridional winds, mean meridional moisture fluxes, and precipitation rates for four seasons are shown in Figures 27 - 29 respectively.

Figure 26 shows the zonal winds for these cases. The largest differences occur at about 20°, 50° and polar regions in both hemispheres, and are on the order of 0.5 m s$^{-1}$ for the four seasons. Figure 27 shows that the corresponding major changes of meridional winds occur in north and south subtropics, depending on season, with changes on the order of 1 m s$^{-1}$. Meridional wind differences between these three cases are larger near the south pole than near the north pole. This is because zonal wind changes between these cases are larger in south polar region, and meridional wind magnitude is parameterized here as the square of zonal wind. The changes in meridional winds result in large changes in the mean meridional moisture flux on the order of $1 \times 10^{15}$ W in the tropics and subtropics (Figure 28). The changes in the mean meridional moisture fluxes in polar region are very small. This is because in the polar regions the surface air temperatures are the lowest globally, which causes the saturation vapor pressure to be very small. The meridional moisture flux is parameterized as a function of vapor pressure, which is a measure of the amount of moisture in the air, and thus when this is small the mean meridional moisture fluxes are small.

Figure 29 gives the resulting precipitation rates for changes in zonal wind. The major differences in precipitation rates occur in tropics and northern subtropics for DJF,
tropics and southern subtropics for JJA, tropics and subtropics in both hemispheres for MAM and SON. These changes are on the order of $1 \times 10^{-8} \text{ m s}^{-1}$ from the control case. These differences are caused by corresponding changes in the mean meridional moisture fluxes (Figure 28), which produce changes in latent heat convergence in same regions and seasons, therefore affect the precipitation rates (similar to earlier sensitivity study of mean meridional moisture flux).
Figure 26. Seasonally and zonally averaged zonal wind in ms$^{-1}$ computed using surface air temperature data from Schutz and Gates (1974) for the control case (solid line), 75% zonal wind case (dotted line), and 125% zonal wind case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
Figure 27. Seasonally and zonally averaged meridional wind in ms$^{-1}$ computed using surface air temperature data from Schütz and Gates (1974) for the control case (solid line), 75% zonal wind case (dotted line), and 125% zonal wind case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
Figure 28. Seasonally and zonally averaged mean meridional moisture flux in W computed using surface air temperature data from Schutz and Gates (1974) for the control case (solid line), 75% zonal wind case (dotted line), and 125% zonal wind case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
Figure 29. Precipitation rate seasonally and zonally averaged over land and ocean in m s\(^{-1}\) computed using surface air temperature data from Schutz and Gates (1974) for the control case(solid line), 75% zonal wind case (dotted line), and 125% zonal wind case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
Sensitivity to Relative Humidity

In all the cases described above, the relative humidity is specified as 0.74 (CCSI model, Ledley, 1988 a&b). In this experiment set, using surface air temperature data from Schutz and Gates (1974), two changes in the relative humidity are presented: one is reduced to 0.7, second is increased to 0.8.

Figures 30 through 32 give the mean and eddy meridional moisture fluxes, and precipitation rates averaged over land and ocean resulting from these changes in relative humidity. We can see these two sets results are fairly close, this implies that sensitivity of parameterization to relative humidity is very small.
Figure 30. Seasonally and zonally averaged mean meridional moisture flux in W computed using surface air temperature data from Schutz and Gates (1974) for the relative humidities equal to 0.7 case (solid line), equal to 0.8 case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
Figure 31. Seasonally and zonally averaged eddy meridional moisture flux in W computed using surface air temperature data from Schutz and Gates (1974) for the relative humidities equal to 0.7 case (solid line), equal to 0.8 case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
Figure 32. Precipitation rate seasonally and zonally averaged over land and ocean in m s\(^{-1}\) computed using surface air temperature data from Schutz and Gates (1974) for the relative humidities equal to 0.7 case (solid line), equal to 0.8 case (dashed line): (a) DJF, (b) MAM, (c) JJA, and (d) SON.
B. Comparison with Observations and Other Modeling Studies

In this section, the results produced by our parameterization will be presented and compared to the available observational data and other modeling results. This will give an idea of the accuracy of the parameterization and indicate components which may need improvement. All our results here are calculated using surface air temperature data from Schutz and Gates (1974). The parameterization results and data are presented in Figures 33 - 44.

The data

The data with which the parameterization results are compared come from a number of sources and are of variable quality and limited extent.

The sources for wind speed data used in Figures 33 - 34 are Newell (1973), Saltzman (1983), and Oort and Rasmusson (1971). Saltzman's data is only available for DJF and JJA, Oort and Rasmusson's data is available from 10°S to 75°N.

The moisture flux data used in Figures 35 - 39 come from Newell (1973), Oort and Rasmusson (1971), Sellers (1965), Gabites (1950), Benton and Estoque (1954), and Starr (1958). Newell's and Oort and Rasmusson's data are the sources with monthly values and only four months: January, April, July, and October. Newell's data covers 15°S - 35°N. Oort and Rasmusson's data covers only the Northern Hemisphere as does Starr's data.

The precipitation rate data for Figures 40 - 44 come from Schutz and Gates (1971, 1972, 1973, 1974, Schutz and Gates obtained their data from Möller 1951, which was from various unspecified years), and Sellers (1965). We also present here the GCM results
from Washington and Meehl (1984) and Manabe (1985). Washington and Meehl's results are only available for DJF and JJA.

Comparison

1. Wind

The seasonally and zonally averaged zonal and meridional winds calculated by the parameterization are shown with data in Figures 33 and 34 respectively for the four seasons. We can see both zonal and meridional winds appear to be well simulated in sign, and seasonal changes. The latitudinal distribution of the computed winds are very similar to the observed with centers of peaks in approximately the correct latitudes, except the computed zonal winds are lower in subtropics. The computed zonal winds are also too weak in the Southern Hemisphere. The reason for this is probably due to the stronger circulation over the dominant ocean because it is uninterrupted by land, while the parameterization was developed for more average climatic conditions.
Figure 33. Seasonally and zonally averaged zonal wind in m s\(^{-1}\) computed using surface air temperature data from Schutz and Gates (1974) (solid line), observations from Newell (1973) (Δ), Saltzman (1983) (o), and Oort and Rasmusson (1971) (x): (a) DJF, (b) MAM, (c) JJA, (d) SON.
Figure 34. Seasonally and zonally averaged meridional wind in m/s computed using surface air temperature data from Schutz and Gates (1974) (solid line), observations from Newell (1973) (Δ), Saltzman (1983) (o), and Oort and Rasmusson (1971) (x): (a) DJF, (b) MAM, (c) JJA, (d) SON.
2. Meridional moisture flux

The Figures 35 and 36 give the zonally averaged mean meridional moisture fluxes and eddy meridional moisture fluxes respectively for four months, Figure 37 gives the annual average of the total meridional moisture fluxes. By comparing with observed data, we find the computed values are generally smaller with the greatest differences occurring at the peaks. These differences between computed and observed data tend to be associated with differences in observed and computed wind speeds. For example, in January (Figure 35a) around 10°N and in July (Figure 35c) around 10°S the computed mean meridional moisture fluxes are lower than observed, this is also where the computed zonal wind speeds are lower than observed (Figure 33a and 33c respectively). Figure 36 shows computed eddy meridional moisture fluxes are smaller around 30° than observed. The lower computed mean and eddy meridional moisture fluxes at lower mid latitudes combine together to give the differences in the annual total meridional moisture fluxes (Figure 37). Figure 37 also shows that the latitudinal amplitude of computed results is lower in lower and mid latitudes than observed.

Generally, the meridional moisture fluxes appear to be well-simulated in latitudinal distribution and seasonal cycle, only the magnitudes have some differences with observed data.

3. Zonal moisture flux convergence

Figures 38 - 39 show the computed and observed zonal moisture flux convergences over land and over ocean respectively for four different months. The agreement of computed flux with observations in magnitude, sign, latitudinal distribution, and seasonal cycle are fairly close. In Figure 38 we find that computed zonal moisture flux convergence
over land has a dip around 55°S. This occurs because in the model the zonal moisture flux convergence is computed as the net moisture flux into the air over an area divided by that area. Around 55°S the land fraction is very small, so by dividing a very small area, here we get the large divergence around 55°S
Figure 35. Zonally averaged mean meridional moisture flux in W computed using surface air temperature data from Schutz and Gates (1974) (solid line), observations from Newell (1973) (Δ), and Oort and Rasmusson (1971) (x): (a) January, (b) April, (c) July, (d) October.
Figure 36. Zonally averaged eddy meridional moisture flux in W computed using surface air temperature data from Schutz and Gates (1974) (solid line), observations from Newell (1973) (Δ), and Oort and Rasmusson (1971) (x): (a) January, (b) April, (c) July, (d) October.
Figure 37. Annually and zonally averaged total meridional moisture flux in W computed using surface air temperature data from Schutz and Gates (1974) (solid line), observations from Sellers (1965) (△), Gabites (1950) (x), Benton and Estoque (1954) (o), and Starr (1958) (+).
Figure 38. Zonal moisture flux convergence over land in W m\(^{-2}\) computed using surface air temperature data from Schutz and Gates (1974) (solid line), observations from Newell (1973) (+): (a) January, (b) April, (c) July, (d) October.
Figure 39. Zonal moisture flux convergence over ocean in W m\(^{-2}\) computed using surface air temperature data from Schutz and Gates (1974) (solid line), observations from Newell (1973) (+): (a) January, (b) April, (c) July, (d) October.
4. Precipitation rate

The comparison between the computed and observed precipitation rates are given in Figures 40 - 44.

Figures 40 and 41 give precipitation rates over land and over ocean respectively for four seasons, and Figure 42 gives annually averaged precipitation rates over land and over ocean. In Figures 40 and 42a we find that around 65°S, precipitation rates computed over land are much larger than observed, while at 55°S in DJF and JJA (Figures 40a and 40c) observed values are higher than computed. There may be a number of reasons for these differences. First, in this region, the land fraction is small. The precipitation due to latent heat convergence is computed as the net moisture flux into the air over an area divided by that area. So by dividing a very small area, we can get a large convergence or divergence such as the large convergence around 65°S and large divergence around 55°S. The high observed precipitation rates may be caused by the presence of mountainous land which may produce a local uplifting of the moist surface air that comes from over ocean that would produce precipitation. This effect is not included in by the parameterization.

Generally, computed precipitation rates simulate seasonal patterns, and the latitudinal distribution of computed precipitation rates are found to be similar to the observed, with peaks and valleys approximately in the correct latitude zones. The differences that do occur seem to be within the natural variability of the precipitation rate.

Figures 43 and 44 give precipitation rates zonally averaged over land and ocean for four seasons and for annually averaged respectively. Compared with observed data and GCM results, the computed precipitation rates are well simulated in magnitude, and latitude distribution, however, there are several differences.
The computed precipitation rates show a seasonal cycle that is strongly a function of temperature, while the seasonal cycle of observed precipitation rates is not very clear.

In our parameterization, the precipitation rate is partially determined from the saturation vapor pressure, which is a function of surface air temperature. Therefore, the computed precipitation rates should decrease when temperatures are lower, so computed precipitation rates have lower values during the winter as compared to summer in both hemispheres. We find for the four seasons (Figure 43) and the annually averaged case (Figure 44), the observed precipitation rates are higher in southern than in the northern mid latitudes. Since sea fractions are larger in southern than in the northern mid latitudes, and considering that evaporation is larger over ocean than over land, above data analysis implies that besides considering precipitation rates dependent on surface air temperature, the precipitation parameterization should also consider the differing impacts of land and ocean surfaces on the precipitation rate.

Considering limited observed data sources and the natural variability of precipitation rate, the computed precipitation rates simulate the observed data and GCM results fairly well.
Figure 40. Seasonally averaged precipitation rate over land in m s\(^{-1}\) computed using surface air temperature data from Schutz and Gates (1974) (solid line), observations from Schutz and Gates (1971, 1972, 1973, 1974) (+): (a) DJF, (b) MAM, (c) JJA, (d) SON.
Figure 41. Seasonally averaged precipitation rate over ocean in ms$^{-1}$ computed using surface air temperature data from Schutz and Gates (1974) (solid line), observations from Schutz and Gates (1971, 1972, 1973, 1974) (+): (a) DJF, (b) MAM, (c) JJA, (d) SON.
Figure 42. Annually averaged precipitation rate in ms\(^{-1}\) computed using surface air temperature data from Schutz and Gates (1974) (solid line), observations from Schutz and Gates (1971,1972,1973,1974) (+): (a) over land, (b) over ocean.
Figure 43. Precipitation rate seasonally and zonally averaged over land and ocean in m$^{-1}$s$^{-1}$ computed using surface air temperature data from Schutiz and Gates (1974) (solid line), observations from Schutz and Gates (1971,1972,1973,1974) (+), and GCM results from Washington and Meehl (1984) (o): (a) DJF, (b) MAM, (c) JJA, (d) SON.
Figure 44. Precipitation rate annually and zonally averaged over land and ocean in ms$^{-1}$ computed using surface air temperature data from Schutz and Gates (1974) (solid line); observations from Schutz and Gates (1971, 1972, 1973, 1974) (+), Sellers (1965) (∆); and GCM results from Manabe (1985) (o).
Summary

Allowing for a natural variability and observational errors, the computed and observed values of precipitation rates, winds, and moisture fluxes agree fairly well. The parameterization succeeded in simulating the seasonal cycle and latitudinal distribution. The amplitude of the seasonal cycle of observed precipitation rates is not very large; however, the parameterization simulates a seasonal cycle that is strongly a function of temperature. The major differences between the computed and observed precipitation rates occur in mid latitudes. These differences may be improved by considering the differing impacts of land and ocean on precipitation in the computation.

Overall, the quantitative computed results have some inconsistencies with observed data, but the qualitative results are a good simulation of the features of the atmospheric circulation and precipitation.
IV. DISCUSSION AND SUMMARY

In summary, a parameterization for the computation of moisture transports and precipitation rates, based upon the energy balance model of Sellers (1973) and precipitation parameterization of Schneider and Thompson (1978), has been developed. This parameterization has been tested by comparing computed energy transports and precipitation rates with available observations and some GCM results, and by evaluating its sensitivity to variations in the values of specified parameters.

The results of our experiments on the sensitivity to surface air temperature showed that winds and meridional moisture fluxes are sensitive to the temperature variations in the tropics; precipitation rates are sensitive to temperature variations at lower and mid latitudes. The precipitation rates are also sensitive to variations in winds, and meridional moisture fluxes. Changes in surface air temperature can produce similar patterns of change in precipitation rate, for example, a change in the seasonal amplitude of surface air temperature may cause a seasonal change in precipitation rate; and a change in the mean temperature may cause a change in the mean precipitation rate.

By analyzing parameterization sensitivity with respect to changes in the zonal wind speed, we find that the mean meridional moisture flux is more sensitive to the zonal and meridional winds changes in lower latitudes. The changes of meridional moisture flux then cause changes in latent heat convergence (a major term in precipitation parameterization), therefore affecting the precipitation rate.

The experiments also indicated that the sensitivity of the parameterization to relative humidity may be very small.
The results produced by parameterization are compared with a number of sources of observational data and some GCM modeling results. The comparison shows that both the winds and meridional moisture fluxes appear to be well-simulated in sign, seasonal change and latitudinal distribution, except the computed magnitudes are slightly lower in subtropics than observed. The computed zonal winds are also weak in the Southern Hemisphere, and computed eddy meridional moisture fluxes are lower in mid latitudes than observed. Zonal moisture flux convergences are simulated fairly well in magnitude, sign, latitudinal distribution and seasonal cycle.

Comparing the computed and observed (also GCM results) precipitation rates, the major differences occur in the mid latitudes, which suggest that the precipitation parameterization should reflect the differing moisture condition at a land surface and an ocean surface.

In conclusion, despite above uncertainties, the computed and observed values agree qualitatively fairly well. The parameterization does a good job of simulating the seasonal cycle and latitudinal distribution of many of the computed variables. The parameterization experiments conducted up to now suggest a number of modifications. Some of these include 1) incorporating the differing moisture conditions of land and ocean surfaces and their impact on latent heat flux and precipitation, 2) considering the impact of stronger circulation over dominant ocean in Southern Hemisphere, and 3) try to determine why the computed meridional moisture fluxes in the lower and mid latitudes are smaller than observed. This parameterization will be incorporated into the CCSI model. The expanded CCSI model will be used to examine the direct effect of environmental changes on the energy and moisture exchanges between the different parts of the climate system and their subsequent effect on the climate.
Appendix A. Derivation of Exponent $a_1$ for Vertical Profile of Specific Humidity

(equation 15)

Vertical variation of specific humidity $q$:

$$\bar{q} = \bar{q}_0 \left(\frac{p}{p_0}\right)^{a_1}$$

Precipitable water, $w_a$

$$\bar{w}_a = \frac{1}{g} \int_0^{p_0} \bar{q} dp$$

$$\bar{w}_a = \frac{1}{g} \int_0^{p_0} \bar{q}_0 \left(\frac{p}{p_0}\right)^{a_1} dp$$

$$\bar{w}_a = \frac{\bar{q}_0}{gp_0^{a_1}} \int_0^{p_0} p^{a_1} dp$$

$$\bar{w}_a = \frac{\bar{q}_0 p_0}{g} \left(\frac{1}{a_1 + 1}\right)$$

$$\Rightarrow$$

$$a_1 + 1 = \frac{\bar{q}_0 p_0}{g \bar{w}_a}$$

Substitute for $\bar{q}_0$ using

$$\bar{q}_0 = \frac{0.622 \bar{e}_0}{p_0}$$

then

$$1 + a_1 = \frac{0.622 \bar{e}_0}{g \bar{w}_a}$$
Appendix B. Derivations of coefficients, equations (16) - (19)

For $a_G$, start with equation (3):

$$ G_S = \frac{1}{g} \int_0^{\infty} \frac{\partial \bar{s}}{\partial t} \, dp = \frac{1}{g} a_G S_0 \frac{\partial S_0}{\partial t} p_0 $$(3)

Let $s = Lq$, and introduce (13):

$$ \bar{q} = q_0 (p/p_0)^{a_1} $$

we get

$$ \frac{1}{g} \int_0^{\infty} \frac{\partial \bar{s}}{\partial t} \, dp = \frac{1}{g} \int_0^{\infty} \frac{\partial \bar{Lq}}{\partial t} \, dp $$

$$ = \frac{1}{g} \frac{\partial \bar{Lq}_0}{\partial t} \int_0^{\infty} \left( \frac{p}{p_0} \right)^{a_1} dp $$

$$ = \frac{1}{g} \frac{1}{1 + a_1} \frac{\partial \bar{Lq}_0}{\partial t} p_0 $$

compare with the right-hand side of (3)

$$ a_{GV} = \frac{1}{1 + a_1} $$

(16)

For $a_x$, start with equation (4):

$$ \Delta_x S = \frac{1}{g} \int_0^{\infty} \frac{\partial \bar{u}}{\partial x} \bar{s} \, dp = \frac{1}{g} \int_0^{\infty} \bar{u} \frac{\partial \bar{s}}{\partial x} \, dp = \frac{1}{g} a_{xS} \bar{u}_0 \frac{\partial S_0}{\partial x} p_0 A_L $$

(4)

Let $s = Lq$, and introduce (11) and (13):

$$ \bar{u} = \bar{u}_0 - (p_0 - p) \frac{\partial \bar{u}}{\partial p} $$

(11)
\[ \overline{q} = \overline{q}_0 \left( \frac{p}{p_0} \right)^{a_1} \]  

(13)

we get

\[ \frac{1}{g} \int_0^{p_0} \overline{u} \frac{\partial \overline{s}}{\partial x} \, dp = \frac{1}{g} \int_0^{p_0} \overline{u} \frac{\partial \overline{Lq}}{\partial x} \, dp \]

\[ = \frac{1}{g} \int_0^{p_0} \left( \overline{u}_0 + \left( p - p_0 \right) \frac{\partial \overline{u}}{\partial p} \right) \frac{\partial \left( L\overline{q}_0 \left( \frac{p}{p_0} \right)^{a_1} \right)}{\partial x} \, dp \]

\[ = \frac{1}{g} \left( \overline{u}_0 - p_0 \frac{\partial \overline{u}}{\partial p} \right) \frac{\partial \left( L\overline{q}_0 \right)}{\partial x} \left( p_0^{a_1} \right) + \frac{1}{g} \frac{\partial \overline{u}}{\partial p} \frac{\partial \left( L\overline{q}_0 \right)}{\partial x} \left( p_0^{a_1} \right) \int_0^{p_0} p a_1 + 1 \, dp \]

\[ = \frac{1}{g} \left( \overline{u}_0 - p_0 \frac{\partial \overline{u}}{\partial p} \right) \frac{\partial \left( L\overline{q}_0 p_0 \right)}{\partial x} \left( 1 + a_1 \right) + \frac{1}{g} \frac{\partial \overline{u}}{\partial p} \frac{\partial \left( L\overline{q}_0 p_0^2 \right)}{\partial x} \left( 2 + a_1 \right) \]

\[ = \frac{1}{g} \frac{1}{1 + a_1} \left( 1 - \frac{1}{2 + a_1} \frac{p_0}{\overline{u}_0} \frac{\partial \overline{u}}{\partial p} \right) \overline{u}_0 \frac{\partial \overline{Lq}_0}{\partial x} p_0 \]

compare with right-hand side of (4)

\[ a_Mv = \frac{1}{1 + a_1} \left( 1 - \frac{1}{2 + a_1} \frac{p_0}{\overline{u}_0} \frac{\partial \overline{u}}{\partial p} \right) \]

(17)

For \( a_M \), start with equation (6):

\[ M_S = \frac{1}{g} \int_0^{p_0} \overline{v} \frac{\overline{s}}{\partial p} \, dp = \frac{1}{g} a_M \overline{v}_0 \overline{s}_0 p_0 \]

(6)

Let \( s = Lq \), and introduce (12) and (13):

\[ \overline{v} = \overline{v}_0 \left( \frac{2p - p_0}{p_0} \right) \]

(12)

\[ \overline{q} = \overline{q}_0 \left( \frac{p}{p_0} \right)^{a_1} \]

(13)
we get
\[
\frac{1}{g} \int_0^{p_0} \vec{\tau} \cdot \vec{s} \, dp = \frac{1}{g} \int_0^{p_0} \vec{\tau} \cdot \vec{L} \cdot \vec{q} \, dp
\]
\[
= \frac{1}{g} \int_0^{p_0} \vec{\tau}_0 \left( \frac{2p - p_0}{p_0} \right) \vec{L} \vec{q}_0 \left( \frac{p}{p_0} \right)^{a_1} \, dp
\]
\[
= \frac{1}{g} \vec{\tau}_0 \vec{L} \vec{q}_0 \int_0^{p_0} \left( \frac{2p - p_0}{p_0} \right) \frac{p^{a_1}}{p_0^{a_1+1}} \, dp
\]
\[
= \frac{1}{g} \frac{a_1}{(1 + a_1)(2 + a_1)} \vec{\tau}_0 \vec{L} \vec{q}_0 \, p_0
\]

compare with the right-hand side of (6)
\[
a_{MV} = \frac{a_1}{(1 + a_1)(2 + a_1)}
\]  

(18)

For \( a_E \), start with equation (7):
\[
E_S = - \frac{1}{g} \int_0^{p_0} K_S \frac{\partial \vec{s}}{\partial y} \, dp = - \frac{1}{g} a_{E}K_S \frac{\partial \vec{s}_0}{\partial y} \, p_0
\]

(7)

Let \( s = \vec{L} \vec{q} \), and introduce (13):
\[
\vec{q} = \vec{q}_0 (p/p_0)^{a_1}
\]

(13)

we get
\[
- \frac{1}{g} \int_0^{p_0} K_S \frac{\partial \vec{s}}{\partial y} \, dp = - \frac{1}{g} \int_0^{p_0} K_S \frac{\partial \vec{L} \vec{q}}{\partial y} \, dp
\]
\[
= - \frac{1}{g} \int_0^{p_0} K_S \frac{\partial \vec{L}}{\partial y} \left( \vec{q}_0 \left( \frac{p}{p_0} \right)^{a_1} \right) \, dp
\]
\[
= - \frac{1}{g} \frac{1}{1 + a_1} K_S \frac{\partial \vec{L} \vec{q}_0}{\partial y} \, p_0
\]
compare with the right-hand side of (7)

\[ a_{EV} = \frac{1}{1 + a_1} \]  

(19)
Appendix C. Derivation of precipitation rate due to latent heat convergence, equation (41)

1 gram of water = 1 cm\(^3\) of water, so 1 gm cm\(^{-2}\) = 1 cm precipitable water.

Latent heat of vaporization = 2470 J gm\(^{-1}\).

The unit of \(L_H\) is J s\(^{-1}\) m\(^2\).

If 1 J s\(^{-1}\) m\(^2\) of latent heat is released for a day then through a units conversion, see below, the amount of water condensed or precipitated would be 0.0350 mm day\(^{-1}\)

\[
\frac{J}{s m^2} \times \frac{1 \text{ gm}}{2470 \text{ J}} \times \frac{86400 \text{ s}}{\text{day}} \times \frac{10^3 \text{ mm}^3}{\text{gm}} \times \frac{m^2}{10^6 \text{ mm}^2} = 0.0350 \text{ mm day}^{-1}
\]

\[
\Rightarrow
\]

\[p_L = 0.0350 \ L_H \ (\text{mm d}^{-1}) \quad (41)\]
Appendix D. Definition of some physical terms

**Barotropic and baroclinic atmospheres**  A portion of the atmosphere in which the surfaces of pressure, density, temperature are all parallel, $\rho = \rho(p)$, is called barotropic. The atmosphere is approximately barotropic in large regions in the tropics. A baroclinic atmosphere is characterized by the presence of a solenoidal field formed by the intersection of temperature and pressure surfaces, i.e., $\rho = \rho(p,T)$. The Earth's atmosphere is mainly baroclinic.

**Baroclinic eddies**  The observed large scale meridional temperature gradient reaches its largest magnitude in mid latitudes. This temperature gradient produces large-scale wave disturbances that interrupt the zonal symmetry of the mid latitude flow. Along with the development of these waves a weak thermally indirect meridional cell develops in middle latitudes. Gradually the kinetic energy of the atmospheric motions levels off and the atmosphere reaches some sort of equilibrium in this "wave regime". This state of equilibrium does not represent a steady, undisturbed flow; rather it is a continually changing mass and motion field in which the wave disturbances are forming and decaying at more or less the same rate, in a statistical sense. These changes are called baroclinic eddies and are commonly experienced as the passage of low pressure systems in mid latitudes. The above wave disturbances are called baroclinic waves; the process by which they amplify is called baroclinic instability (Wallace and Hobbs, 1977).

**Precipitable water**  The total atmospheric water vapor contained in a vertical column of unit cross-sectional area extending between any two specified levels, commonly expressed in terms of the height to which that water substance would stand if completely condensed and collected in a vessel of the same unit cross-section.
**Absolute humidity**  The density of water vapor, $p_v$.

**Specific humidity**  The ratio of the mass of vapor in a certain volume to the total mass of air and vapor in the same volume.

**Relative humidity**  The ratio of the actual mass of vapor in a volume to the saturation mass of vapor in the same volume at same temperature and pressure.
## Appendix E. Physical variables and constants, units

<table>
<thead>
<tr>
<th>Quantity</th>
<th>Symbol</th>
<th>Units and numerical values</th>
</tr>
</thead>
<tbody>
<tr>
<td>Acceleration of gravity</td>
<td>$g$</td>
<td>$9.8 \text{ m s}^{-2}$</td>
</tr>
<tr>
<td>surface air pressure</td>
<td>$p_0$</td>
<td>$10^5 \text{ Pa}$</td>
</tr>
<tr>
<td>Latent heat of condensation</td>
<td>$L$</td>
<td>$2.5 \times 10^6 \text{ J kg}^{-1}$</td>
</tr>
<tr>
<td>Specific humidity</td>
<td>$q$</td>
<td>unitless</td>
</tr>
<tr>
<td>Specific heat of air at constant pressure</td>
<td>$c_p$</td>
<td>$1004 \text{ J kg}^{-1} \text{ (°K)}^{-1}$</td>
</tr>
<tr>
<td>Air temperature</td>
<td>$T$</td>
<td>°K</td>
</tr>
<tr>
<td>Height above sea level</td>
<td>$h$</td>
<td>m</td>
</tr>
<tr>
<td>Heat capacity of soil or water times density</td>
<td>$C$</td>
<td>$\text{ J m}^{-3} \text{ (°K)}^{-1}$</td>
</tr>
<tr>
<td>Soil or water temperature</td>
<td>$T_E$</td>
<td>°K</td>
</tr>
<tr>
<td>East-west wind velocity component</td>
<td>$u$</td>
<td>m s$^{-1}$</td>
</tr>
<tr>
<td>North-south wind velocity component</td>
<td>$v$</td>
<td>m s$^{-1}$</td>
</tr>
<tr>
<td>Vertical wind velocity component</td>
<td>$\omega$</td>
<td>Pa s$^{-1}$</td>
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<tr>
<td>Eddy diffusivity coefficient for latent heat</td>
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<td>m$^2$ s$^{-1}$</td>
</tr>
<tr>
<td>Coriolis parameter</td>
<td>$f$</td>
<td>s$^{-1}$</td>
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<tr>
<td>Gas constant for dry air</td>
<td>$R_d$</td>
<td>$287.04 \text{ J kg}^{-1} \text{ (°K)}^{-1}$</td>
</tr>
<tr>
<td>Pressure in atmospheric layers</td>
<td>$P_a$</td>
<td>$5 \times 10^4 \text{ Pa}$</td>
</tr>
<tr>
<td>Precipitable water</td>
<td>$W_a$</td>
<td>m</td>
</tr>
<tr>
<td>Surface vapor pressure</td>
<td>$e_0$</td>
<td>Pa</td>
</tr>
<tr>
<td>Angular velocity of rotation</td>
<td>$\Omega$</td>
<td>$7.292 \times 10^{-5} \text{ rad s}^{-1}$</td>
</tr>
<tr>
<td>Density</td>
<td>$\rho$</td>
<td>$\text{ kg m}^{-3}$</td>
</tr>
<tr>
<td>Friction force</td>
<td>$F_r$</td>
<td>N</td>
</tr>
<tr>
<td>Specific volume</td>
<td>$\alpha$</td>
<td>m$^3$ kg$^{-1}$</td>
</tr>
</tbody>
</table>
Latent heat transport by

mean meridional motions \( M_V \) \( \text{W m}^{-1} \)
eddy meridional motions \( E_V \) \( \text{W m}^{-1} \)
mean zonal motions \( \Delta_s S \) \( \text{W m}^2 \)

Meridional latent heat convergence \( L_H \) \( \text{W m}^2 \)

\( L_H \) converted to water equivalent \( p_L \) \( \text{mm d}^{-1} \)
Saturation vapor pressure \( e_s \) \( \text{Pa} \)
Relative humidity \( r \) \( \text{unitless} \)

Temperature gradient using in

precipitation parameterization \( \Delta T \) \( ^\circ\text{K/1000 km} \)
REFERENCES


