In order to understand the atmospheric branch of the earth's hydrologic cycle on the global scale, an atmospheric moisture balance is diagnostically analyzed from the January and July data of the OSU atmospheric general circulation model, which has been integrated for thirty-nine months of simulation with seasonally-varying sea-surface temperature and solar insolation. The model hydrologic processes analyzed for the balance include the surface evaporation, the precipitation by large-scale and cumulus condensation, the vertical transport by large-scale and cumulus mass fluxes, and the horizontal transport of water vapor. The large-scale transports include the contributions from the standing and transient components of motion. Also analyzed are the potential and stream functions of horizontal transport, and the statistics of seasonal and interannual variabilities of the global and hemispheric effects of the hydrologic processes.

As a result of these analyses, the hydrologic cycle is constructed and understood for both January and July of the model. Large-scale vertical transport moistens the upper layer; the standing and transient
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Comparison with other models and observational data indicates that the model reproduced some basic features of the atmospheric branch of the hydrologic cycle and its seasonal variation. The intense evaporation ($\geq 5 \text{ mm day}^{-1}$) over the Pacific and Atlantic oceans and the rain belts in the tropics are well simulated for both January and July. The poleward transport in the northern middle and high latitudes is in good agreement with observations. The maximum toward-thermal-equator transport in the tropics occurs, however, at the geographic equator for both January and July, indicating that these maxima are about 5 degrees of latitude closer to the seasonal thermal equator than the observed maxima. Nevertheless the global statistics of the model atmosphere are not significantly different from that of the real atmosphere.
Among others, we mention the following common features of the January and July moisture balances in the present model. Most precipitation of penetrating convection occurs in regions of strong surface evaporation even though some occurs in the moisture convergence zones where most of heavy mid-level convection is located. In the regions of intense penetrating convection, however, the standing part of surface evaporation is much larger in magnitude than the negative transient part which is essentially due to the positive correlation between the turbulence intensity and surface humidity over wet surfaces. Moreover, the horizontal structure of the standing part conforms to that of the standing vapor pressure difference between the air and the underlying surface. A strategy for further studies is recommended on the basis of our understanding of these features.
Analysis of the Atmospheric Water Vapor Transport and the Hydrologic Cycle Simulated in a Global Circulation Model

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1. INTRODUCTION

The atmospheric water vapor transport often fluctuates to cause floods and droughts, and therefore constitutes an important branch of the hydrologic cycle. The present study is intended for understanding major components of the atmospheric moisture balance by analyzing the simulation data of the Oregon State University two-level atmospheric general circulation model (hereafter called the AGCM for brevity).

In order to make a clear perspective of the problems of the present study, the observational and modeling studies of the atmospheric hydrologic cycle are reviewed in section 1.1, and then in section 1.2, motivation and definition of the problems will be presented.

1.1 A brief review of the previous studies

Atmospheric water vapor is a primary climatic variable since the general circulation depends absolutely on condensational and radiative effects of water vapor and clouds. Considerable investigation on the atmospheric hydrologic cycle has been made on a broad spectrum of scales. They may be classified into observational and modeling studies according to the sources of data used for them.

A. Observational studies

The annual statistics of water vapor and its flux by transient eddies were first computed by White (1951) using the northern hemispheric data north of 25°N. Then an attempt was made by Starr and White (1955) to evaluate the meridional transport using the data of 1950. (Subsequent investigations using the same data are well summarized in Peixoto (1958).) Starr et al. (1969) completed the first global analysis of mean annual humidity conditions, using the 1958 IGY data.
The General Circulation Data Library (hereafter called the GCDL for brevity) of the mean monthly statistics collected at the MIT by Starr et al. (1970) for the extended period of 1 May 1958 - 30 April 1963 opened the way to a new generation of observational studies of humidity conditions over the Northern Hemisphere and also over the southern hemispheric tropics. Even though a pilot study of the GCDL by Rasmusson (1967) indicated a need for additional low-level winds in view of the importance of the low-level fluxes in the tropics, the average annual water vapor balance was presented by Peixoto (1970) using the GCDL.

It is generally recognized that the water vapor content of the atmosphere is linked to the terrestrial hydrologic cycle; this recognition, gradually obtained during the 1950's and 1960's, seems to be prompted by the large variability of atmospheric water vapor transport according to Peixoto (1973). Thus while the total water vapor content of the atmosphere is only a small fraction (about 0.001%, e.g., see Eagleson (1970)) of the hydrospheric mass, the large-variability of the atmospheric transport makes it important in the global water balance.

A recent study by Rosen et al. (1976) of the interannual variability of the mean meridional circulation indicates that the annual statistics for the year of 1 May 1967 - 30 April 1968 differed significantly from those for one of the original first five years of the GCDL (which now includes data up to 30 April 1973). Even though Oort (1977) contends that many of the zonal mean statistics from the third five-year period of the GCDL resemble those of the first five years, and the statistics of the first five years may be representative of the total mean conditions, similarity in the zonal mean statistics did not imply that significant regional differences did not occur. Rosen et al. (1979a, 1979b) show the evidence of the interannual variability in the atmospheric transport and also in the stream function of its rotational component. In analyzing the year-to-year changes of the individual annual mean fluxes, however, they used once-daily measurements so that sampling bias might have exaggerated the documented variability.

In all these observational studies several difficulties are noteworthy. First, efforts to define the humidity conditions suffer from
great gaps in the rawinsonde and surface networks (e.g., see Fig. 5.1 of Newell et al. (1972)). The lack of data or insufficient data coverage precludes a definitive investigation of the regional water balance over much of the globe and therefore gives rise to uncertainty in the zonally averaged quantities.

Second, the bulk of the horizontal flux of water vapor takes place below 800 mb in the tropics and a central analysis problem is the accurate evaluation of low-level winds, and to a lesser extent, of low-level humidity fields. The examination of rawinsonde data with high vertical resolution from several tropical stations clearly indicates, according to Rasmusson (1972), the need for 50-mb resolution in order to properly evaluate the fluxes in the low levels, and the flux between the surface and 1000 mb may not be ignored particularly in the tropics. The difficulty with the low-level flux evaluation is also due to the highly transient winds in the low levels. The low-level winds of the continental and tropical island stations show marked diurnal oscillations resulting in a net water vapor flux which often overshadows other effects in the lowest one or two kilometers according to Rasmusson (1972). The pronounced influence of local topography and land-sea contrast on the low-level diurnal winds is well known, while diurnal winds associated with large-scale topographic features such as major mountains and continents may be ignored (see for example Rasmusson (1966)). Nevertheless, topographically-induced wind fields have an adverse effect on the analysis because smaller-scale features in areas of sparse data will be interpreted as large-scale features of the flow.

Third, there are instrumental problems which are probably of lesser importance in the large-scale water balance studies but should nevertheless be kept in mind for better interpretation. Missing observations of humidity due to the inability of a sensor to respond to small water vapor concentration introduces some errors primarily in the middle and upper troposphere. More importantly, the transport of water in liquid and solid forms is not accounted for by conventional sensors.

Finally, local observations of precipitation are often inadequate for estimating the large-scale pattern and therefore the zonal means
are not quite reliable. Moreover, it is very difficult to observationally identify the generic mechanisms of an observed total precipitation.

B. Model studies

Climate modeling with an atmospheric general circulation model was first attempted by Phillips (1956). He employed a two-level quasi-geostrophic model of the mid-latitude large-scale circulations but assumed a zonally uniform heating which was independent of the state. He succeeded in simulating the subtropical jet and three-celled mean meridional circulation in spite of other simplifying assumptions on the model processes including friction. There were no hydrologic processes in this model whose domain was confined to the middle latitudes of the Northern Hemisphere.

Smagorinsky (1963) then used a GCM based upon a primitive-equation system valid in the tropics as well as higher latitudes and opened the way for incorporation of the hydrologic cycle into an AGCM. Manabe et al. (1965) first included the atmospheric moisture as an important state variable and analyzed in detail the hydrologic processes (i.e., evaporation, transports and precipitation) in the model where the lower boundary was a uniformly wet and flat surface. They simulated, however, the zonal mean features of precipitation such as the tropical rain belt, subtropical bands of meager precipitation and mid-latitude rain belts. The ground hydrology was considered by Manabe (1969) later when he investigated the climatic effects of the land-sea contrast in wetness by using an AGCM on a limited domain with an idealized geography. A further detailed analysis by Manabe and Holloway (1975) of their model showed that the seasonal variation of the hydrologic cycle was well simulated including the correct locations of the watersheds of the major rivers of the world.

Even though moisture fields were explicitly calculated in some models cited in the previous paragraph, they were not used in modeling the effects of clouds or the radiative transfer. In fact, climatological data were used for the atmospheric moisture and clouds needed for calculating the radiative transfer in the models. Employing the stable computational method of Arakawa (1966), Mintz (1965) made a long-term
integration of the primitive equation AGCM on the global domain with realistic geography and topography. (Arakawa (1972) describes the radiation and cloud schemes, and further details may be found in Katayama (1972).) Mintz et al. (1972) then extended the time integration beyond one annual cycle with interactive moisture fields, and succeeded in simulating the seasonal and interannual variations of tropospheric circulation on the global scale. The hydrologic simulation with fully interactive moisture fields was not reported and the hydrologic processes were then relatively new components of their models.

1.2 Motivation and definition of the problem

From the review in the last section, it is hard to avoid feeling that observational data are inadequate for better understanding the processes of the hydrologic cycle and that numerical simulation of the atmospheric general circulation by AGCMs have become more realistic with the inclusion of hydrologic processes. (See Table 1.2.1 for a subjective comparison of the characteristics of the simulated and observed data.) Thus it seems of significant scientific merit to analyze the moisture balance and hydrologic cycle of an AGCM and to understand the model hydrologic processes, i.e., the evaporation, transport and precipitation, which contribute to the requirements and fulfillments of the balance. Recently Schlesinger and Gates (1981) made a three-year integration of the AGCM to simulate the observed annual cycle including a complete hydrologic cycle, but analyzed the precipitation only in regard to the hydrologic processes.

The purposes of the present study are to understand how atmospheric water vapor is transported (i.e., by large-scale standing and transient motions, cumulus convection, boundary-layer turbulence and internal evaporation of free-falling water drops) and how the global hydrologic cycle occurs in the AGCM. In order to solve these problems, the hydrologic processes of the moisture balance will be analyzed, the global hydrologic cycle constructed, and their variability calculated. The results are then compared to reliable observations, and a strategy for model improvement is recommended on the basis of this comparison.
Table 1.2.1 Subjective comparison of OSU model and observational data

<table>
<thead>
<tr>
<th>data resolution</th>
<th>model</th>
<th>observation</th>
</tr>
</thead>
<tbody>
<tr>
<td>horizontal resolution</td>
<td>4° lat x</td>
<td>variable</td>
</tr>
<tr>
<td>5° long (*)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>station density</td>
<td>uniform</td>
<td>nonuniform</td>
</tr>
<tr>
<td>globally</td>
<td>and data-sparse</td>
<td></td>
</tr>
<tr>
<td>vertical resolution</td>
<td>worse (**)</td>
<td>better</td>
</tr>
<tr>
<td>time interval</td>
<td>uniform at every 6 hours</td>
<td>nonuniform</td>
</tr>
<tr>
<td>data interpretation</td>
<td>lengths of records</td>
<td>39 months</td>
</tr>
<tr>
<td>interpretative advantage</td>
<td>high</td>
<td>low</td>
</tr>
<tr>
<td>level of understanding</td>
<td>high</td>
<td>low</td>
</tr>
<tr>
<td>data reliability</td>
<td>aliasing error</td>
<td>smaller</td>
</tr>
<tr>
<td>other processing errors</td>
<td>truncation</td>
<td>instrumental</td>
</tr>
</tbody>
</table>

*see Fig. 1.2.1

**OSU AGCM has only two levels in the vertical direction
Fig. 1.2.1. Geographic distribution of the model surface elevation (meters). Dashed lines are for the datum at 200 m; solid lines are for higher elevations at contour interval of 100 m. Peaks of Himalayas, east and west of Antarctica, Greenland, Andes, Rockies and Africa have elevations of 4410 m, 3616 m, 3549 m, 2603 m, 1769 m, 1754 m, and 1062 m, respectively.
In the next two chapters, the model equations and moisture transports are discussed. In Chapter 4, data and methods of analysis are described. Results are then discussed in Chapter 5 followed by the conclusions in the last chapter.
2. MODEL EQUATIONS

2.1 Inviscid diabatic system for large scale circulation

Under the traditional approximation (Eckart, 1960; Phillips, 1966) the hydrostatic primitive equations on the spherical surface with the modified Phillips (1957) vertical coordinate \( \sigma \) for the inviscid diabatic atmospheric system are

\[
\frac{\partial u}{\partial t} = -V \cdot \nabla u - \frac{a \partial u}{\partial \sigma} - \frac{1}{a \cos \phi} \left( \frac{\partial \Phi}{\partial \lambda} + \frac{\partial p}{\partial \lambda} \right) + \left( f + \frac{u \tan \phi}{a} \right)v, \tag{2.1.1}
\]

\[
\frac{\partial v}{\partial t} = -V \cdot \nabla v - \frac{a \partial v}{\partial \sigma} - \frac{1}{a} \left( \frac{\partial \Phi}{\partial \phi} + \frac{\partial p}{\partial \phi} \right) - \left( f + \frac{u \tan \phi}{a} \right)u, \tag{2.1.2}
\]

\[
\frac{\partial T}{\partial t} = -V \cdot \nabla T - \frac{a}{c_p} \frac{\partial T}{\partial \sigma} + \frac{1}{c_p} \omega a + \frac{g}{\pi c_p} \frac{\partial R}{\partial \sigma} + \frac{1}{c_p} \frac{\partial \mathcal{L}}{\partial \sigma}, \tag{2.1.3}
\]

\[
\frac{\partial q}{\partial t} = -V \cdot \nabla q - \frac{a}{c_p} \frac{\partial q}{\partial \sigma} - C, \tag{2.1.4}
\]

\[
\frac{\partial \pi}{\partial t} = -V \cdot (\pi V) - \frac{a}{\partial \sigma} (\pi \phi), \tag{2.1.5}
\]

\[
\frac{\partial \Phi}{\partial \sigma} = -\pi a \tag{2.1.6}
\]

where \( u \) and \( v \) are the zonal and meridional components of horizontal wind vector \( V \), \( V \) is horizontal gradient operator, \( \Phi \) is the specific geopotential energy, \( p \) is the pressure, \( f \) is Coriolis parameter, \( a \) is the radius of the earth, \( \phi \) is the vertical velocity in the \( \sigma \)-coordinate, \( T \) is the temperature, \( \omega \) is the individual time derivative of pressure, \( g \) is the gravity, \( R \) is the net upward radiation flux, \( a \) is the specific volume, \( c_p \) is the heat capacity at constant pressure, \( q \) is the water vapor mixing ratio, \( C \) is the net condensation rate per unit mass of dry air, \( L \) is the latent heat of evaporation, \( \pi \) is the pressure thickness of the model atmosphere, i.e., \( \pi = p_s - p_T \), \( p_s \) and \( p_T \) are the pressures at the
surface and model top, respectively, \( t \) is time, \( \lambda \) and \( \phi \) are the longitude and latitude, respectively, and the vertical coordinate \( \sigma \) is defined as

\[
\sigma = \frac{p - p_T}{p_S - p_T}.
\] (2.1.7)

The hydrostatic approximation is justified because the horizontal scale of large scale atmospheric disturbances is much larger than the scale height of the atmosphere.

Since the present investigation is related to the hydrological simulation of the AGCM, we will only be concerned with the basic equation of water vapor in the rest of this chapter.

2.2 Continuity equation of water vapor

Combining (2.1.4) and (2.1.5), we can obtain the flux form of the conservation equation for atmospheric water vapor as

\[
\frac{\partial}{\partial t} (\pi q) = -\nabla \cdot (\pi \nabla q) - \frac{\partial}{\partial \sigma} (\pi \partial q) - \pi c
\] (2.2.1)

Since we will not solve this equation analytically, we have to develop a budget equation for the grid box (defined in next section) in the model atmosphere. First, we introduce the convention of the ensemble average operation.

For variable \( A \), which has no mass in its dimension, we define

\[
\bar{A} (\lambda, \phi, \sigma, t) = \frac{\int_{\varepsilon} \pi (\lambda, \phi, t; \varepsilon) A(\lambda, \phi, \sigma, t; \varepsilon) d\varepsilon}{\int_{\varepsilon} \pi (\lambda, \phi, t; \varepsilon) d\varepsilon},
\]

\[
A (\lambda, \phi, \sigma, t; \varepsilon) = \bar{A} (\lambda, \phi, \sigma, t) + A' (\lambda, \phi, \sigma, t; \varepsilon)
\] (2.2.2)

where \( \varepsilon \) represents the ensemble of realizations.

For variable \( B \), which has mass in its dimension, we define

\[
\hat{B} (\lambda, \phi, \sigma, t) = \frac{\int_{\varepsilon} B(\lambda, \phi, \sigma, t; \varepsilon) d\varepsilon}{\int_{\varepsilon} d\varepsilon}
\]

\[
B (\lambda, \phi, \sigma, t; \varepsilon) = \hat{B} (\lambda, \phi, \sigma, t) + B'' (\lambda, \phi, \sigma, t; \varepsilon).
\] (2.2.3)
Of course, 
\[ \dot{A} = 0 \]  
and  
\[ \dot{B} = 0 \]  
Thus, the averaged balance equation over the grid volume is  
\[ \frac{\partial}{\partial t} (\tilde{\pi q}) = -\nabla \cdot (\tilde{\pi} \tilde{V} q) - \frac{\partial}{\partial \sigma} (\tilde{\pi} \tilde{\sigma} q) - \tilde{\pi} C \]  
(2.2.5)  

where \( \tilde{C} \) is the net large-scale condensation, and the last two terms are the unresolved subgrid-scale effects due to the parameterized model convective and turbulent processes.

2.3 The model equation of water vapor balance

The model characteristics are briefly described here and more details can be seen in Schlesinger and Gates (1979). The atmosphere is vertically divided into two layers with equal mass from the surface to 200 mb, as shown in Fig. 2.3.1, and horizontally resolved by the model's 4° lat x 5° long global grid by which continents with smoothed topography are outlined, as seen in Fig. 1.2.1. The surface elevation is obtained from the tabulation of Gates and Nelson (1975) on a 1°x1° grid. For convenience we consider the water mass budget for each model layer. For brevity of notation, we will drop the bar and circumflex without confusion hereafter.

For the upper layer the water mass budget is  
\[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_1 \right) = -\nabla \cdot \left( \frac{\pi \Delta \sigma}{g} \nabla q_1 \right) - \frac{\pi}{g} \left( \delta q_1 \right) - \frac{\pi \Delta \sigma}{g} C_1 \]  
\[ + \left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_1 \right) \right]_{MLC} + \left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_1 \right) \right]_{PC} , \]  
(2.3.1)
Fig. 2.3.1. The vertical structure and principal variables of the two-level atmospheric general circulation model. Here $\sigma = (p - p_T)/p_S - p_T$ where $p$ is pressure at the earth's surface, $u$ and $v$ are the eastward and northward velocity components, $T$ the temperature, $\phi$ the geopotential, $q$ the water vapor mixing ratio, $P$ the precipitation rate, $E_S$ the surface evaporation rate and $GW$ the ground wetness. $\dot{\sigma}$ is the vertical velocity in $\sigma$-coordinate and $\dot{\sigma} = 0$ at the model top and bottom. Here 0, 1, 2, 3 and 4 on the right margin denote the top, upper, middle, lower and surface levels, respectively.
where the last two terms are the subgrid-scale effects due to mid-level and penetrating convection. We will discuss their parameterization in the next section. The subscripted indices 1, 2 and 3 refer to the upper, middle and lower levels, respectively, as seen in Fig. 2.3.1.

For the lower layer we have

\[
\frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_3 \right) = - \nabla \cdot \left( \frac{\pi \Delta \sigma}{g} V q_3 \right) + \frac{\pi}{g} (\Delta q)_2 - \frac{\pi \Delta \sigma}{g} C_3 \\
+ \left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_3 \right) \right]_{MLC} + \left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_3 \right) \right]_{PC} + E_s \tag{2.3.2}
\]

where \( E_s \) is the surface evaporation. Note that the dimension of each term is mass per unit time and per unit area and that it is equivalent to water mass flux.

### 2.4 Parameterization of subgrid-scale processes

The present model considers middle-level convection, penetrating convection and surface evaporation as the parameterized subgrid-scale processes.

#### A. Middle-level convection

Assuming that the mid-level convective cloud mass is in a statistically steady state. Then from the schematic diagram in Fig. 2.4.1 we can formulate its effect on the environmental air of the upper layer as

\[
\left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_1 \right) \right]_{MLC} = \eta M_c (q_{c,1} - q_2) \tag{2.4.1}
\]

where \( q_{c,1} \) is the specific humidity of the detrained air from the cloud into its environment (\( q_{c,1} \) will be derived in the next chapter), \( q_2 \) is the mixing ratio of the environment at level 2, and \( \eta M_c \) is the cloud-induced mass flux. We may note that \( \eta M_c \) is constant everywhere since there is no accumulation or consumption of cloud mass under steady state assumption. In the middle-level convection, there is no influence
Fig. 2.4.1. Schematic diagram of middle-level convective cloud ensemble; $\eta M$ is the mass flux induced by the cloud ensemble. Here 0, 1, 2, 3 and 4 denote the top, upper, middle, lower and surface levels, respectively.
from the boundary layer, so only the ambient air in the lower layer can be entrained. Similarly, we can write the effect on the lower layer as

$$\frac{\partial}{\partial t} \left( \frac{\pi A \sigma}{g} q_3 \right)_{MLC} = \eta M_c (q_2 - q_3)$$  \hspace{1cm} (2.4.2)

To determine the cloud mass flux, we have to know the sufficient condition which favors the occurrence of this kind of convection. We define \( h = c_p T + g z + L q \) as moist static energy and assume that if the moist static energy of the lower layer is larger than the saturated moist static energy of the upper layer, i.e.,

$$h_3 > h_1 \quad (\equiv c_p T_1 + q z_1 + L q_s (T_1, p_1))$$  \hspace{1cm} (2.4.3)

then the middle-level convection will occur. Where the superscript \( (\cdot)^* \) denotes a saturation value, and \( q_s(T,p) \) is the saturation mixing ratio at temperature \( T \) and pressure \( p \). This subgrid-scale process removes a conditional moist-convective instability of the environment, within a "turnover" time period during which the instability will be reduced to \( e^{-1} \) of the initial intensity. In order to save space in the main text, we will put the detailed derivation in Appendix A, provided by Schlesinger (1979, personal communication), and write down the formula for the cloud mass flux \( \eta M_c \) and the turnover time period \( \tau \);

$$\eta M_c = \frac{1}{\tau} \frac{\pi A \sigma}{g} \left[ \frac{h_3 - h_1^*}{(h_3-h_1^*) + L(q_3-q_2) + \gamma_1 (s_1-s_2) + s_1 + s_3 - 2s_2} \right]$$  \hspace{1cm} (2.4.4)

where \( \tau \) is assumed to be one hour, and \( s \) is the dry static energy

$$s = c_p T + g z = h - L q .$$  \hspace{1cm} (2.4.5)

B. Penetrating convection

The penetrating convection diagramatically shown in Fig. 2.4.2 occurs under the following conditions:
Fig. 2.4.2. Schematic diagram of penetrating convective cloud ensemble; $M$ is the mass flux induced by the cloud ensemble. Here 0, 1, 2, 3 and 4 denote the top, upper, middle, lower and surface levels, respectively.
The net effect on the moisture change is

\[
\frac{\partial}{\partial t} \left( \frac{\rho \Delta \sigma}{g} q_1 \right)_{PC} = M_c (q_{c,1} - q_2)
\]

\[
\frac{\partial}{\partial t} \left( \frac{\rho \Delta \sigma}{g} q_3 \right)_{PC} = M_c (q_2 - q_4)
\]

Following Schlesinger and Gates (1979), the cloud mass flux is

\[
M = \frac{1}{\tau} \frac{\rho \Delta \sigma}{g} \frac{h_4 - h_3^*}{[1 + \gamma_3 - \frac{\partial T_4}{\partial T_3}]} (s_2 - s_4) + \frac{L}{\rho \Delta \sigma} (q_4 - q_2)
\]

where \(T_4\) and \(q_4\) are the surface air temperature and mixing ratio which are diagnostically related to \(T_3\) and \(q_3\) among others, \(\partial T_4/\partial T_3\) and \(\partial q_4/\partial q_3\) are the partial derivatives and \(\gamma \equiv \frac{L}{\rho \Delta \sigma} \frac{\partial q_3}{\partial T_3}\). The detailed derivation can be seen in Appendix A.

C. Surface evaporation

There is no explicit treatment of the planetary boundary layer (PBL) in the present version of the model but we need to parameterize the turbulent flux of water vapor at the earth's surface. Using the bulk aerodynamic method the surface evaporation \(E_s\) is

\[
E_s \equiv - \frac{\rho_s C_D V_s \beta}{g} \left( q_g - q_4 \right)
\]

where \(q_g \equiv q_s (T_g, p_s), \rho_s\) is surface air density, \(C_D\) is the drag coefficient given by

\[
C_D = \begin{cases} 
0.002 \left( 1 + \frac{3z_s}{5000} \right), & \text{over land and ice} \\
\text{Min}\left[0.001 \left( 1 + 0.07 \left| V_4 \right| \right), 0.0025\right], & \text{over water}
\end{cases}
\]
along with the surface elevation $z_s$ in meters. Here $V_s$ is surface wind speed given by

$$V_s = \text{Max} \left(0.7 \ |V_4|, \ 2 \text{ m s}^{-1}\right) \quad (2.4.12)$$

using the extrapolated wind

$$V_4 = \frac{3}{2} V_3 - \frac{1}{2} V_1 \quad (2.4.13)$$

and the gustiness of $2 \text{ m s}^{-1}$. The coefficient $\beta$ represents the ratio of the actual evapotranspiration to the potential evapotranspiration and, following Arakawa (1972), depends on the ground wetness $w$ as

$$\beta = \text{Min} \ (2w, \ 1) \quad (2.4.14)$$

except for $\beta$ taken equal to unity if either $q_4 > q_g$ (dew deposit) or the surface is snow-covered. The ground wetness is defined as the ratio of the amount of water stored in the ground to the field capacity (taken as 15 g cm$^{-2}$).

In order to determine surface air mixing ratio $q_4$, surface evaporation is assumed to be equal to the mean $K$-theory flux below the lower prognostic level following Arakawa (1972);

$$E_s = -\rho_s K \left(\frac{q_3 - q_4}{z_3 - z_s}\right) \quad (2.4.15)$$

where $K \ (\text{m}^2 \text{s}^{-1})$ is

$$K = \begin{cases} \text{Min} \ \{10 \ \exp \left[0.32 \ (T_g - T_4)\right], 100\} & \text{over land and ice} \\ \text{Min} \ \{10 \ \exp \left[0.32 \ (T_g - T_4)\right], 15\} & \text{over water} \end{cases} \quad (2.4.16)$$

From (2.4.10) and (2.4.15) we can obtain

$$q_4 = \frac{C_D V_s \beta q_g + K}{C_D V_s \beta + \frac{K}{z_3 - z_s}} q_3 \quad (2.4.17)$$
3. FORMULATIONS OF MOISTURE TRANSPORTS

In the model, the effective moisture transports by some of the processes are not explicitly calculated. However, the emphasis of the present study is placed on the diagnostic analysis of the simulation data and it is desirable to formulate and estimate those moisture transports consistently with the model physics. Large-scale and sub-grid-scale processes will be investigated separately in this chapter.

3.1 Large-scale processes

In the balance equation for the atmospheric moisture, there are three terms related to large-scale processes, i.e., horizontal convergence, vertical convergence and large-scale condensation. We will investigate them one by one.

A. Kinematic analysis of horizontal transport

Let the horizontal moisture flux vector \( \mathbf{F} \) be

\[
\mathbf{F}(\lambda, \phi, \sigma) = k \times \nabla \psi + \nabla \chi \tag{3.1.1}
\]

where the first and second terms on the right hand side are the rotational (nondivergent) and divergent (irrotational) parts, respectively. We call \( \psi \) stream function and \( \chi \) scalar potential. The curl of (3.1.1) yields

\[
\nabla^2 \psi = \nabla \cdot \nabla \times \mathbf{F} \tag{3.1.2}
\]

and the divergence of (3.1.1) yields

\[
\nabla^2 \chi = \nabla \cdot \mathbf{F} \tag{3.1.3}
\]

Usually, if there are proper boundary conditions, the stream function and scalar potential can be determined. Since the elliptic Poisson
equations are linear, the solutions influenced by different sources can be linearly superposed. For example, if \( \nabla^2 x_1 = S_1, \nabla^2 x_2 = S_2 \) and \( \nabla^2 x = S_1 + S_2 \), then

\[ x = x_1 + x_2. \]  
\[(3.1.4)\]

Thus, if \( S_1 \) and \( S_2 \) are the flux divergences due to quasi-stationary and transient motions, respectively, then the total scalar potential is the sum of the potentials due to the respective motions. Detailed solution method is in Appendix B.

According to (3.1.1), the moisture sources are located at the relative minima of \( x \); the water vapor fluxes diverge from the regions of low to high potential where the water vapor may experience a phase change to precipitate. However, the rotational part of the flux is parallel to the contours of stream function; the water vapor carried by this flux moves along the streamline without any precipitation. The moving direction is clockwise around the "high" (relative maximum); the sense of the moving direction around the "low" (relative minimum) is just the opposite.

B. The large-scale vertical transports

In the present OSU AGCM there are two grid-scale vertical transport processes. One is by the large-scale vertical motion itself and the other is due to the large-scale condensation and subsequent internal evaporation at a lower altitude. They will be discussed respectively in the following.

(i) Large-scale vertical motion

Since the vertical velocity \( \sigma \) in the model is carried at each of the even levels, and \( \sigma_0 = \sigma_4 = 0 \) in particular, the vertical moisture transport we need to know is that at the middle level only;

\[
\frac{\pi}{g} \sigma_2 \frac{\partial q_2}{\partial \sigma} = \frac{\partial}{\partial \sigma} \left( \frac{\pi \Delta \sigma}{g} \sigma q \right)_1 = -\frac{\partial}{\partial \sigma} \left( \frac{\pi \Delta \sigma}{g} \sigma q \right)_3. \]
\[(3.1.5)\]
When $\dot{\sigma}_2$ is positive, the flux is downward so that the lower layer gains water vapor while the upper layer loses the same amount of water vapor. The opposite situation occurs when $\dot{\sigma}_2$ is negative.

(ii) Large-scale condensation

Due to the advection effects of large-scale motions the specific humidity at any level $\kappa$ ($\kappa = 1$ or $3$) can be larger than the saturation specific humidity $q_s(T, p)$. Under this condition the large-scale condensation occurs. The temperature increases from $T^{(0)}_{\kappa}$ to $T^{(1)}_{\kappa}$ and the specific humidity decreases from $q^{(0)}_{\kappa}$ to $q^{(1)}_{\kappa} = q_s(T^{(1)}_{\kappa}, p)_{\kappa}$;

$$c_p(T^{(1)}_{\kappa} - T^{(0)}_{\kappa}) = L[q^{(0)}_{\kappa} - q_s(T^{(1)}_{\kappa}, p)_{\kappa}]$$

(3.1.6)

An alternative interpretation of the above equation is that the moist static energy is conserved during the course of condensation. We can iteratively solve (3.1.6) for $T^{(1)}_{\kappa}$ by the Newton-Raphson method (at least two iterations); the first-order approximation is

$$T^{(1)}_{\kappa} = T^{(0)}_{\kappa} - \frac{F(T^{(0)}_{\kappa})}{F'(T^{(0)}_{\kappa})}$$

(3.1.7)

where

$$F(T^{(0)}_{\kappa}) = -\frac{L}{c_p} [q^{(0)}_{\kappa} - q_s(T^{(0)}_{\kappa}, p)_{\kappa}]$$

(3.1.8)

and

$$F'(T^{(0)}_{\kappa}) = 1 + \gamma^{(0)}_{\kappa}$$

(3.1.9)

along with $\gamma^{(0)}_{\kappa} \equiv \gamma(T^{(0)}_{\kappa}, p)_{\kappa}$. Substituting (3.1.8) and (3.1.9) into (3.1.7) yields the temperature change over a time interval $\Delta t$ as

$$\Delta T^{(1)}_{\kappa} = T^{(1)}_{\kappa} - T^{(0)}_{\kappa} = \frac{L}{c_p} \frac{[q^{(0)}_{\kappa} - q_s(T^{(0)}_{\kappa}, p)_{\kappa}]}{1 + \gamma^{(0)}_{\kappa}}$$

(3.1.10)
The moisture change per unit time is accordingly

$$C'_{k} = \frac{c}{L} \frac{\Delta T}{\Delta t} \quad (k = 1 \text{ or } 3) \quad (3.1.11)$$

The water condensing in the upper layer, $C'_1 \Delta t$, falls into the lower layer where it completely evaporates. The lower layer is thereby cooled and moistened as a result of this transport. Condensed water in this layer, however, falls to the ground as large-scale precipitation;

$$p_{LS} = C'_3. \quad (3.1.12)$$

In the water balance equation the net large-scale condensation of each layer is

$$C_1 = C'_1, \quad (3.1.13)$$

and

$$C_3 = C'_3 - C'_1. \quad (3.1.14)$$

### 3.2 Subgrid-scale processes

In order to distinguish the transport and condensation due to subgrid-scale convection we have to make a detailed derivation. Suppose that the air is saturated in the clouds. Then

$$q_i = q_s(T_i, p_i) \quad (3.2.1)$$

for the $i$-th cloud. If the effect on $q_s$ of a pressure difference between the cloud and the environment is neglected, then

$$q_i = q_s(T_i, \bar{p}) \approx \bar{q}_s + \frac{\partial q_s}{\partial T} (T_i - \bar{T}) \quad (3.2.2)$$
where $\overline{q_s} = q_s(\overline{T}, \overline{p})$ and the over-bar stands for the environment. Also, from the definition of the moist static energy,

$$
\overline{h_i} - \overline{h^*} = c_p \left[ 1 + \frac{L}{c_p} \overline{\frac{\partial q_s}{\partial T}} \right] (\overline{T_i} - \overline{T})
$$

$$
= c_p (1 + \gamma) (\overline{T_i} - \overline{T}) \quad (3.2.3)
$$

From (3.2.2) and (3.2.3)

$$
q_i - q_s = \frac{1}{L} \frac{\gamma}{1 + \gamma} \left( h_i - \overline{h^*} \right) \quad (3.2.4)
$$

Assume that all of the clouds in the grid volume are identical (horizontally uniform on the isobaric surface) i.e., $q_i = q_c$ and $h_i = h_c$. Then

$$
q_c = q_s + \frac{1}{L} \frac{\gamma}{1 + \gamma} \left( h_c - \overline{h^*} \right) \quad (3.2.5)
$$

where $c$ represents the cloud ensemble. For simplicity we will drop the overbar. Now we can consider the application of (3.2.5) to the OSU two-level AGCM.

A. Penetrating convection

In the present version of the model we ignore the effect of lateral entrainment on the cloud ensemble. The moist static energy in this cloud is then

$$
h_c = h_4 \quad (3.2.6)
$$

where $h_4$ is the moist static energy of the surface boundary layer. Since we are interested in computing the precipitation from the lower and upper portions of the cloud ensemble respectively, we may use (3.2.5) and (3.2.6) to determine the mixing ratio in the cloud ensemble at the middle level, viz.,
\[ q_{c,2} = q_2^* + \frac{1}{L} \frac{\gamma_2}{1 + \gamma_2} (h - h_2^*) \]  

(3.2.7)

where \( q_2^* = q_s(T_2, p_2) \) and \( h_2^* = s_2 + Lq_2^* \). Therefore, the condensations from the upper and lower portions of the cloud ensemble are given respectively by

\[ C_{PC1} = M_c(q_{c,2} - q_{c,1}) \]  

(3.2.8)

and

\[ C_{PC3} = M_c(q_4 - q_{c,2}) \]  

(3.2.9)

The net upward transport by the convection is given by

\[ T_{PC2} = M_c(q_{c,2} - q_2) \]  

(3.2.10)

It is required that the sum of the condensations as given by (3.2.8) and (3.2.9) be equal to the precipitation which is the sum of the net effects on the two model layers, viz.,

\[ P_{PC} = C_{PC1} + C_{PC3} \]

\[ = -\left[ \frac{\partial}{\partial t} \left[ \frac{\pi dG}{g} (q_1 + q_3) \right] \right]_{PC} \]

\[ = M_c(q_4 - q_{c,1}) . \]  

(3.2.11)

B. Mid-level convection

For this subgrid-scale process the influence of the surface boundary layer is assumed negligible. The moist static energy entrained from the lateral sides is that of the lower level, \( h_3 \); the cloud ensemble has then
From (3.2.5) and (3.2.12) the cloud mixing ratio at level 2 is obtained as

\[ q_{c,2} = q_{2}^* + \frac{1}{L} \frac{\gamma_{2}}{1 + \gamma_{2}} (h_{3} - h_{2}^*) \]  

(3.2.13)

Therefore, the condensations from the upper and lower portions of the cloud ensemble are given respectively by

\[ C_{MLC1} = n_{M} (q_{c,2} - q_{c,1}) \]  

(3.2.14)

and

\[ C_{MLC3} = n_{M} (q_{3} - q_{c,2}) \]  

(3.2.15)

The net upward transport by the convection is given by

\[ T_{MLC2} = n_{M} (q_{c,2} - q_{2}) \]  

(3.2.16)

All of these quantities are positive definite as well as those in (3.2.8) - (3.2.10). The precipitation due to the mid-level convection is

\[ P_{MLC} = C_{MLC1} + C_{MLC3} \]

\[ = - \left[ \frac{\partial}{\partial t} \left( \frac{n_{M} \Delta \sigma}{g} (q_{1} + q_{3}) \right) \right]_{MLC} \]

\[ = n_{M} (q_{3} - q_{c,1}) . \]  

(3.2.17)

Therefore water vapor balance equations (2.3.1) and (2.3.2) can be rewritten as
\[
\frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_1 \right) = -\nabla \cdot \left( \frac{\pi \Delta \sigma}{g} \nabla q_1 \right) - \frac{\pi}{g} \hat{\alpha}_2 q_2 - \frac{\pi \Delta \sigma}{g} (C + C_{PC} + C_{MLC})_1 \\
+ [M_{PC}(q_{PC,2} - q_2) + \eta_{MLC}(q_{MLC,2} - q_2)]_1 \tag{3.2.18}
\]

for the upper layer,

\[
\frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_3 \right) = -\nabla \cdot \left( \frac{\pi \Delta \sigma}{g} \nabla q_3 \right) + \frac{\pi}{g} \hat{\alpha}_2 q_2 - \frac{\pi \Delta \sigma}{g} (C + C_{PC} + C_{MLC})_3 \\
- [M_{PC}(q_{PC,2} - q_2) + \eta_{MLC}(q_{MLC,2} - q_2)]_3 + E_s \tag{3.2.19}
\]

for the lower layer.
4. DATA AND METHODS OF ANALYSIS

In this chapter we will describe the data and methods of analysis.

4.1 Description of data

A thirty-nine month simulation with the OSU AGCM has been made in 1979. The extent of the simulation is from 1 December of year 0 through 28 February of year 4. The model state was sampled every 6 hours (at simulated GMT 00Z, 06Z, 12Z and 18Z, respectively of each day) to form the history. The state variables recorded in the history, however, belong to time points just prior to the calculation of vertical moisture transport and forcing terms, which was performed only once every hour in the model. Therefore we calculated the effects of forcing terms as well as vertical moisture transport using the history, and obtained new data set for the present analysis. The thirty-nine monthly data is very extensive in size, and moreover, most models have been diagnosed for their results for January and July, which are the mid-season months of winter and summer, respectively. In the present analysis, therefore, we diagnose the data only for January and July.

4.2 Methods of data analysis

A. Time and space means

For any variable \( A \), the time mean and departure therefrom can be defined as

\[
\bar{A} = \left(\frac{1}{T} \int_{t}^{t+T} \pi_A dt\right)/\pi
\]

(4.2.1)

where \( \pi = \frac{1}{T} \int_{t}^{t+T} \pi_A dt \) with \( T \) being the averaging period, and

\[
A' = A - \bar{A}
\]

(4.2.2)

From these definitions we obtain in particular
\[ A' = 0 \] 

\[ A'B' = AB - \overline{A} \overline{B} \]

where \( A'B' \) and \( \overline{A} \overline{B} \) are called the transient and quasi-stationary components of the second moment statistics \( \overline{AB} \) centered at zero. In calculating a monthly mean according to (4.2.1), we used the trapezoidal integration with \( T = 1 \) month.

For the spatial average, the zonal mean is again a mass-weighted average; the weighting mass itself would be time-averaged for the zonal mean of a time mean. The hemispheric and global means are similarly weighted averages taking into account the latitudinally-varying grid area.

B. Composite mean and variability

Since we have four simulations of January and three of July, we can use the monthly means to calculate the composite mean and variability. For example, the composite January mean is just the arithmetic average of four January monthly means, whereas the composite January variability is their standard deviation. The sample size \( (N = 3 \text{ or } 4) \) is too small for the variability to be a statistically meaningful measure of interannual variability, but the estimates are nevertheless computed for our interest in finding the physical causes of interannual variability.
5. RESULTS

In this chapter, results obtained from the present analysis will be discussed in six sections. The first section shows how most of the atmospheric water vapor is originated over oceans. Various transports of water substance are then discussed in section two and three. The next section exhibits each component in the water vapor balance equation. Model verification and simulated hydrologic cycle appear in the last two sections, respectively.

5.1 Origin of atmospheric water vapor

In the water vapor balance for the entire atmosphere the surface evaporation is the only source. The parameterized formula for turbulent moisture flux at the top of the constant flux layer indicates that evaporation is mainly proportional to the drag coefficient, the ground wetness, the wind speed at anemometer level and the vapor pressure difference between ground or sea surface and the surface air. The detailed formulation is in section 2.4.

Fig. 5.1.1 shows the global distributions of the monthly mean ground wetness in January and July of the third model year. Most inland regions of the continents are very dry (below 50%) in January except the northwestern part and central band of North America, the Amazon river basin, part of Europe, the northeastern part of Asia and small islands. It should be noted that sea and ice-covered surfaces have the wetness of one. In July only the southern part of South America and small islands are relatively wet except that Greenland is very wet in both January and July. It is therefore expected that most water vapor is evaporated from the oceans.

The surface evaporation also in January and July of the third year is shown in Fig. 5.1.2. Strong evaporation (larger than $6 \times 10^{-5} \text{ kg s}^{-1} \text{m}^{-2}$) takes place mostly to the east of continents in middle latitudes and over the subtropical oceans. Seasonal variation can be visualized from the expansion and movement of these regions. For example, strong evaporation occurs to
Fig. 5.1.1. Ground wetness (dimensionless) of January and July (continued) of the third year. Dashed lines are for 0.5; solid lines are for 0.75; dotted lines are for lower values at 0.25, 0.15 and 0.05, consecutively.
Fig. 5.1.1. Continued
Fig. 5.1.2. Surface evaporation rate (10^{-5} \text{kg s}^{-1}\text{m}^{-2}) of January and July (continued) of the third year. Dashed lines are for 3; solid lines are for higher values at contour interval of 1.5; dotted lines are for lower values at contour interval of 1.
the east of Asia and North America with the contour of \(6 \times 10^{-5} \text{ kg s}^{-1}\text{m}^{-2}\) enclosing a wider area in January than in July, whereas the one located to the west of the southern part of North America in January is shifted southwestward farther away from the coast in July.

Causes of these centers of strong evaporation are not obvious from the above description. A diagnostic scheme was designed to reveal them. Since the surface evaporation is a subgrid-scale process which transports moisture from the earth's surface to the ambient atmosphere through turbulence, we can separate the parameterized formula into two parts. One represents the intensity of turbulence, \(\rho_s C_D V_s\); the other expresses the intensity of "air mass transformation", or the available moisture content to be evaporated, \(\beta (q^* - q_4)\).

The turbulence intensity in the surface layer in Fig. 5.1.3 shows a remarkable contrast between the continents and oceans, and also between winter and summer. Since in the present model the drag coefficient over land is linearly proportional to the elevation and the surface wind speed is linearly extrapolated from that at two prognostic levels in the model free atmosphere, the drag tends to be large over high terrain and the turbulence is more active over continents than over oceans. The seasonal change is larger over the continents in middle and high latitudes than over the oceans and low-latitude continents, and this might be a good indication of seasonal variation of baroclinicity in the model.

The intensity of "air mass transformation" during January and also July shown in Fig. 5.1.4 indicates that all of the strong activity (\(\geq 4.5 \text{ g kg}^{-1}\)) takes place over oceans approximately between 40°S and 40°N; the intensity is then determined by \(q^*_g - q_4\) over the oceans. The mixing ratio of surface air shown in Fig. 5.1.6 essentially decreases poleward over the oceans except near the coasts, but the sea surface temperature shown in Fig. 5.1.6 reveals relatively cold temperature in subtropical oceans to the west of continents where one therefore expects relatively weak air mass transformation as is evident in Fig. 5.1.4.
Fig. 5.1.3. Surface turbulence intensity \((10^{-2}\text{kg s}^{-1}\text{m}^{-2})\) of January and July (continued) of the third year. Dashed lines are for 1.25; solid lines are for higher values at contour interval of 0.5; dotted lines are for lower values at contour interval of 0.25.
Fig. 5.1.3. Continued
Fig. 5.1.4. Intensity of air mass transformation (g kg$^{-1}$) of January and July (continued) of the third year. Dashed lines are for 3; solid lines are for higher values at contour interval of 1.5; dotted lines are for lower values at contour interval of 1.
Fig. 5.1.5. Mixing ratio of surface air (g kg\(^{-1}\)) of January and July (continued) of the third year. Dashed lines are for 9; solid lines are for higher values at contour interval of 2; dotted lines are for lower values at contour interval of 2.
Fig. 5.1.6. Temperature at the earth's surface (°C) of January and July (continued) of the third year. Dashed lines are 14 and 18 for January and July, respectively. Solid lines are for high values at contour intervals of 6 and 5, respectively. Dotted lines are for lower values at contour intervals of 10 and 5, respectively.
Fig. 5.1.6. Continued
Shown in Fig. 5.1.7 is the standing component of surface evaporation, i.e., the product of monthly mean intensities of surface-layer turbulence and air mass transformation. Comparison between this figure and Fig. 5.1.2 indicates that strong surface evaporation ($\geq 6 \times 10^{-5}$ kg s$^{-1}$ m$^{-2}$) occurs essentially in the standing mode. It should be noted that the areas occupied by this strong standing mode are slightly different from those occupied by the total evaporation. For example, areas are connected in the North Pacific and Atlantic oceans in January, and in July there are three additional areas of strong standing mode which occur to the southeast of the Great Lakes in North America, Lake Baikal in Asia, and along the Niger River in west Africa. They are associated with strong turbulence and moderate air mass transformation. Areas of strong standing evaporation over the Indian and Southwest Pacific oceans are connected in July.

The difference between the total and the standing evaporation is the transient mode as shown in Fig. 5.1.8. This mode represents the temporal correlation between the intensities of turbulence and air mass transformation which are negatively correlated over most of the globe, particularly in the summer hemisphere. Areas of weakly positive correlation are mainly located in northern hemispheric continents in January, but over southern hemispheric continents, particularly over Antarctica, in July. This transient component finally removes those three areas of strong standing evaporation over continents in July and separates areas of strong standing mode over oceans.

Whenever the turbulence is relatively strong (weak) the surface air over most of the ocean between 40°S and 40°N is relatively moist (dry). The cause of this negative contribution to the total evaporation might be due to the present formulation to determine $q_4$.

In summary, the air mass transformation is more significant than the turbulence due to surface drag to maintain or form the strong evaporation centered over the oceans. Also cold sea surface temperature does not favor air mass transformation between 40°S and 40°N, particularly along the east coasts of Asia and Africa in January and to the west of continents in both January and July.
Fig. 5.1.7. Standing component of surface evaporation rate ($10^{-5}$ kg s$^{-1}$ m$^{-2}$) of January and July (continued) of the third year. Dashed lines are for 3; solid lines are for higher values at contour interval of 1.5; dotted lines are for lower values at contour interval of 1.
Fig. 5.1.8. Transient component of surface evaporation rate ($10^{-5}$kg s$^{-1}$m$^{-2}$) of January and July (continued) of the third year. Dashed lines are for -0.25; solid lines are for 0.25; dotted lines are for lower values at contour interval of 0.25.
Fig. 5.1.8. Continued
5.2 Large-scale horizontal transport

Total (monthly mean) horizontal moisture flux can be divided into standing (quasi-stationary) and transient parts through time averaging as in Section 4.2. Also any flux vector can be partitioned into divergent and rotational parts as in (3.1.1) after Poisson equations for potential and stream function are solved. However, the divergent part is more meaningful than the rotational part as far as the water vapor balance is concerned. Therefore the detailed presentation in the following is for the divergent part rather than the rotational part.

The potential function of total moisture flux is shown in Figs. 5.2.1 and 5.2.2. Moisture sources in January in the lower layer are located to the west of continents, i.e., South America (95°W, 2°S), Africa (35°W, 22°N) and Australia (110°E, 26°S) while there are three sink regions around (70°E, 42°N), (75°E, 6°S) and (70°E, 58°S) in the longitude band between 50°E and 100°E. In the upper layer sink regions are located to or in the west of continents, i.e., North America (115°W, 34°N), South America (90°W, 6°S) and Africa (10°W, 30°N); sources are in the equatorial Indian Ocean, (85°E, 6°S) and (130°E, 6°S).

In July sources in the lower layer are located to the west of South America (105°W, 2°S) and Africa (10°W, 10°S) and to the east of Madagascar (60°E, 26°S), and a sink occurs in the Pacific Ocean to the southeast of Asia (145°E, 26°N). Source at (145°E, 10°N) and sinks at (75°W, 26°S) and (5°W, 14°S) in the upper layer corresponding to sink and sources in the lower layer, respectively, can be easily visualized except to the east of Madagascar.

In Tables 5.2.1a, 5.2.1b, 5.2.2a and 5.2.2b is the water vapor budget for all physical processes at the center of source (*) or sink (**) region. It should be noted that horizontal divergence (convergence) occurs at source (sink).

In the upper layer, the source is maintained by large-scale upward transport from below whose magnitude is larger than that of horizontal divergence. These two transports at the source center are stronger in July than in January. Middle-level convection is much stronger than penetrating convection in January, but not in July.
Fig. 5.2.1. Potential function of total moisture flux of composite January (10^8 kg s^{-1}). In the upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.1 are for higher and lower values, respectively. In the lower layer (continued) dashed lines are for -0.8; solid and dotted lines at contour interval of 0.2 are for higher and lower values respectively.
Fig. 5.2.1. Continued
Fig. 5.2.2. Potential function of total moisture flux of composite July ($10^8$ kg s$^{-1}$). In upper layer dashed lines are for -0.35; solid lines at contour interval of 0.15 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.3 are for higher and lower values, respectively.
Fig. 5.2.2. Continued
Table 5.2.1a. Budget for important stations of the upper layer in January.

<table>
<thead>
<tr>
<th>Location</th>
<th>Layer (Column)</th>
<th>Horizontal convergence</th>
<th>Vertical convergence</th>
<th>Large scale condensation</th>
<th>Large scale evaporation</th>
<th>Middle level convection</th>
<th>Penetrating convection</th>
<th>Surface evaporation</th>
<th>Storage</th>
</tr>
</thead>
<tbody>
<tr>
<td>(115°W, 34°N)</td>
<td>Upper**</td>
<td>2.767</td>
<td>-0.806</td>
<td>-1.814</td>
<td>-0.080</td>
<td>0.0</td>
<td>1.314</td>
<td>0.434</td>
<td>0.068</td>
</tr>
<tr>
<td></td>
<td>Lower</td>
<td>4.170</td>
<td>0.806</td>
<td>1.814</td>
<td>-0.151</td>
<td>0.0</td>
<td>1.314</td>
<td>0.503</td>
<td>1.114</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>6.937</td>
<td>-9.331</td>
<td>-1.814</td>
<td>-0.231</td>
<td>0.0</td>
<td>0.0</td>
<td>1.937</td>
<td>1.571</td>
</tr>
<tr>
<td>(90°W, 6°S)</td>
<td>Upper**</td>
<td>1.885</td>
<td>-1.845</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.040</td>
<td>0.226</td>
</tr>
<tr>
<td></td>
<td>Lower</td>
<td>-4.775</td>
<td>1.845</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>3.171</td>
<td>0.241</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>-2.890</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>3.171</td>
<td>0.281</td>
</tr>
<tr>
<td>(10°W, 30°N)</td>
<td>Upper**</td>
<td>2.169</td>
<td>-1.459</td>
<td>-0.640</td>
<td>-0.019</td>
<td>-0.014</td>
<td>0.0</td>
<td>0.040</td>
<td>0.336</td>
</tr>
<tr>
<td></td>
<td>Lower</td>
<td>-1.489</td>
<td>1.459</td>
<td>-1.609</td>
<td>0.640</td>
<td>-0.029</td>
<td>-0.060</td>
<td>1.463</td>
<td>0.377</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>0.68</td>
<td>-2.248</td>
<td>0.640</td>
<td>-0.048</td>
<td>-0.074</td>
<td>1.463</td>
<td>0.413</td>
<td>0.377</td>
</tr>
<tr>
<td>(85°E, 6°S)</td>
<td>Upper*</td>
<td>-1.827</td>
<td>5.033</td>
<td>-0.125</td>
<td>-2.774</td>
<td>-0.293</td>
<td>0.0</td>
<td>0.040</td>
<td>0.144</td>
</tr>
<tr>
<td></td>
<td>Lower</td>
<td>6.948</td>
<td>-5.033</td>
<td>-0.594</td>
<td>0.125</td>
<td>-3.613</td>
<td>-1.58</td>
<td>3.375</td>
<td>0.372</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>5.121</td>
<td>-0.719</td>
<td>0.125</td>
<td>-6.387</td>
<td>-1.873</td>
<td>3.375</td>
<td>-0.358</td>
<td>0.377</td>
</tr>
<tr>
<td>(130°E, 6°S)</td>
<td>Upper*</td>
<td>-1.699</td>
<td>4.332</td>
<td>-0.126</td>
<td>-1.700</td>
<td>-0.658</td>
<td>0.0</td>
<td>0.151</td>
<td>0.228</td>
</tr>
<tr>
<td></td>
<td>Lower</td>
<td>6.133</td>
<td>-4.332</td>
<td>-0.208</td>
<td>0.126</td>
<td>-2.527</td>
<td>-3.40</td>
<td>4.436</td>
<td>0.228</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>4.434</td>
<td>-0.334</td>
<td>0.126</td>
<td>-4.227</td>
<td>-4.058</td>
<td>4.436</td>
<td>0.377</td>
<td></td>
</tr>
</tbody>
</table>

* and ** indicate source and sink, respectively; units are $10^{-5} \text{kg s}^{-1} \text{m}^{-2}$. 
Table 5.2.1b. Budget for important stations of the lower layer in January

<table>
<thead>
<tr>
<th>Location</th>
<th>Layer (Column)</th>
<th>Horizontal convergence</th>
<th>Vertical convergence</th>
<th>Large scale condensation</th>
<th>Large scale evaporation</th>
<th>Middle level convection</th>
<th>Penetrating convection</th>
<th>Surface evaporation</th>
<th>Storage</th>
</tr>
</thead>
<tbody>
<tr>
<td>(95°W, 2°S)</td>
<td>Upper</td>
<td>1.261</td>
<td>-1.311</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>3.680</td>
<td>0.208</td>
</tr>
<tr>
<td></td>
<td>Lower*</td>
<td>-4.781</td>
<td>1.311</td>
<td>0.0</td>
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<td>0.0</td>
<td>0.0</td>
<td>3.680</td>
<td>0.160</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>-3.52</td>
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<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>(35°W, 22°N)</td>
<td>Upper</td>
<td>1.138</td>
<td>-1.054</td>
<td>-0.034</td>
<td>-0.015</td>
<td>-0.122</td>
<td></td>
<td>-0.086</td>
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<tr>
<td></td>
<td>Lower*</td>
<td>-4.516</td>
<td>1.054</td>
<td>-0.150</td>
<td>0.034</td>
<td>-0.022</td>
<td>-0.90</td>
<td>4.368</td>
<td>-0.128</td>
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<tr>
<td></td>
<td>Total</td>
<td>-3.378</td>
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<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>(70°E, 42°N)</td>
<td>Upper</td>
<td>1.221</td>
<td>-0.100</td>
<td>-0.980</td>
<td>-0.082</td>
<td>0.0</td>
<td>0.0</td>
<td>0.059</td>
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</tr>
<tr>
<td></td>
<td>Lower**</td>
<td>3.471</td>
<td>0.100</td>
<td>-4.796</td>
<td>0.980</td>
<td>-0.182</td>
<td>0.0</td>
<td>0.594</td>
<td>0.168</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>4.692</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(75°E, 6°S)</td>
<td>Upper</td>
<td>-2.495</td>
<td>4.858</td>
<td>-0.247</td>
<td>-0.998</td>
<td>-1.068</td>
<td></td>
<td>0.050</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Lower**</td>
<td>6.146</td>
<td>-4.858</td>
<td>-0.202</td>
<td>0.247</td>
<td>-1.319</td>
<td>-5.41</td>
<td>5.062</td>
<td>-0.335</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>3.651</td>
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<td></td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(70°E, 58°S)</td>
<td>Upper</td>
<td>0.099</td>
<td>0.323</td>
<td>-0.399</td>
<td>-0.002</td>
<td>0.0</td>
<td>0.0</td>
<td>0.021</td>
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</tr>
<tr>
<td></td>
<td>Lower**</td>
<td>1.092</td>
<td>-0.323</td>
<td>-1.596</td>
<td>0.399</td>
<td>-0.004</td>
<td>0.0</td>
<td>0.597</td>
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<td></td>
<td>Total</td>
<td>1.191</td>
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<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>(110°E, 26°S)</td>
<td>Upper</td>
<td>1.185</td>
<td>-1.247</td>
<td>-0.005</td>
<td>0.0</td>
<td>-0.011</td>
<td>-0.97</td>
<td>6.800</td>
<td>-0.078</td>
</tr>
<tr>
<td></td>
<td>Lower*</td>
<td>-6.398</td>
<td>1.247</td>
<td>-0.084</td>
<td>0.005</td>
<td>0.0</td>
<td>-0.97</td>
<td>6.800</td>
<td>0.596</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>-5.213</td>
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<td></td>
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<td></td>
<td>0.522</td>
</tr>
</tbody>
</table>
Table 5.2.2a. Budget for important stations of the upper layer in July

<table>
<thead>
<tr>
<th>Location</th>
<th>Layer (Column)</th>
<th>Horizontal convergence</th>
<th>Vertical convergence</th>
<th>Large scale condensation</th>
<th>Large scale evaporation</th>
<th>Middle level convection</th>
<th>Penetrating convection</th>
<th>Surface evaporation</th>
<th>Storage</th>
</tr>
</thead>
<tbody>
<tr>
<td>(75°W, 26°S)</td>
<td>Upper**</td>
<td>3.579</td>
<td>-2.728</td>
<td>-0.573</td>
<td>0.0</td>
<td>0.0</td>
<td>0.278</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Lower</td>
<td>-3.81</td>
<td>2.728</td>
<td>-0.781</td>
<td>0.573</td>
<td>0.0</td>
<td>1.762</td>
<td>0.48</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>-0.231</td>
<td>-1.354</td>
<td>0.573</td>
<td>1.762</td>
<td>0.750</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(5°W, 14°S)</td>
<td>Upper**</td>
<td>1.908</td>
<td>-1.966</td>
<td>-0.001</td>
<td>0.0</td>
<td>0.0</td>
<td>-0.058</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Lower</td>
<td>-6.08</td>
<td>1.966</td>
<td>-0.012</td>
<td>0.001</td>
<td>0.0</td>
<td>4.261</td>
<td>0.14</td>
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</tr>
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<td>Total</td>
<td>-4.172</td>
<td>-0.013</td>
<td>0.001</td>
<td>4.261</td>
<td>0.077</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>(145°E, 10°N)</td>
<td>Upper*</td>
<td>-3.226</td>
<td>5.297</td>
<td>-0.272</td>
<td>-0.983</td>
<td>-0.962</td>
<td>-0.145</td>
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</tr>
<tr>
<td></td>
<td>Lower</td>
<td>6.32</td>
<td>-5.297</td>
<td>-0.176</td>
<td>0.272</td>
<td>-1.535</td>
<td>5.700</td>
<td>0.003</td>
<td></td>
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<tr>
<td></td>
<td>Total</td>
<td>3.094</td>
<td>-0.448</td>
<td>0.272</td>
<td>-2.518</td>
<td>5.700</td>
<td>-0.142</td>
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</tr>
</tbody>
</table>
Table 5.2.2b. Budget for important stations of the lower layer in July.

<table>
<thead>
<tr>
<th>Location</th>
<th>Layer (Column)</th>
<th>Horizontal convergence</th>
<th>Vertical convergence</th>
<th>Large scale condensation</th>
<th>Large scale evaporation</th>
<th>Middle level convection</th>
<th>Penetrating convection</th>
<th>Surface evaporation</th>
<th>Storage</th>
</tr>
</thead>
<tbody>
<tr>
<td>(105°W, 2°S)</td>
<td>Upper</td>
<td>1.037</td>
<td>-1.033</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>4.440</td>
<td>0.004</td>
</tr>
<tr>
<td></td>
<td>Lower*</td>
<td>-5.33</td>
<td>1.033</td>
<td>-0.017</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>4.440</td>
<td>0.13</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>-4.293</td>
<td>-0.017</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>4.440</td>
<td>0.130</td>
</tr>
<tr>
<td>(10°W, 10°S)</td>
<td>Upper</td>
<td>1.585</td>
<td>-1.610</td>
<td>0.0</td>
<td>0.0</td>
<td>-0.070</td>
<td>-0.99</td>
<td>4.944</td>
<td>-0.095</td>
</tr>
<tr>
<td></td>
<td>Lower*</td>
<td>-5.95</td>
<td>1.610</td>
<td>0.0</td>
<td>0.0</td>
<td>-0.99</td>
<td>4.944</td>
<td>-0.38</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>-4.365</td>
<td>-1.060</td>
<td>0.0</td>
<td>0.0</td>
<td>-0.99</td>
<td>4.944</td>
<td>-0.481</td>
<td></td>
</tr>
<tr>
<td>(60°E, 26°S)</td>
<td>Upper</td>
<td>0.383</td>
<td>-0.408</td>
<td>0.0</td>
<td>0.0</td>
<td>-0.032</td>
<td>-1.31</td>
<td>5.408</td>
<td>-0.058</td>
</tr>
<tr>
<td></td>
<td>Lower*</td>
<td>-4.86</td>
<td>0.408</td>
<td>-0.051</td>
<td>0.0</td>
<td>-1.31</td>
<td>5.408</td>
<td>-0.41</td>
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</tr>
<tr>
<td></td>
<td>Total</td>
<td>-4.477</td>
<td>-0.051</td>
<td>-0.051</td>
<td>0.0</td>
<td>-1.342</td>
<td>5.408</td>
<td>-0.462</td>
<td></td>
</tr>
<tr>
<td>(145°E, 26°N)</td>
<td>Upper</td>
<td>-0.871</td>
<td>3.136</td>
<td>0.0</td>
<td>-2.094</td>
<td>-0.052</td>
<td>3.644</td>
<td>0.120</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Lower**</td>
<td>4.22</td>
<td>-3.136</td>
<td>-0.046</td>
<td>0.0</td>
<td>-3.495</td>
<td>-0.29</td>
<td>3.644</td>
<td>0.90</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>-3.349</td>
<td>-0.046</td>
<td>-5.589</td>
<td>-0.342</td>
<td>3.644</td>
<td>1.016</td>
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<td></td>
</tr>
</tbody>
</table>

*Note: Lower* and Lower** indicate specific measurements or conditions at these locations.
The sink in the upper layer can be classified into two categories. First, if a convergence center is located over the continents, it can be essentially maintained by large-scale condensation and downward transport. If there is significant topography nearby, e.g., (115°W, 34°N), large-scale condensation is dominant; otherwise, large-scale downward transport is dominant, e.g., (10°W, 30°N). The second category has the center over the ocean. If the center is not close to a mountainous coast, it is essentially maintained by large-scale downward transport; otherwise, e.g., (75°W, 26°S), large-scale condensation can play a minor role.

The sink in the lower layer is strongly influenced by the seasonal change. In January in the tropics, (75°E, 6°S), the deficit of evaporation over precipitation needs a relatively strong horizontal convergence to maintain the strong upward transport. The penetrating convection is more significant than middle-level convection. The remaining areas at (70°E, 42°N) and (70°E, 58°S) have the large-scale precipitation maintained by horizontal convergence. In July the lower layer sink is sustained by large-scale upward transport while the deficit of evaporation over precipitation is not very significant. Middle-level convection is the major precipitation process for this case.

For the source of the lower layer significant excess of evaporation over precipitation is required. Moderate large-scale downward transport can also be found for most of the sources except the one at (60°E, 26°S) to the east of Madagascar in July which has relatively weak downward transport.

The zonal components of the total divergent moisture flux are shown in Figs. 5.2.3 and 5.2.4 for January and July, respectively. In January strong easterly (westward moisture flux) in the upper layer occurs off the west coasts of continents, i.e., North and South America, Africa and Australia. Easterly transport also takes place in the Indian Ocean to the east of Africa and on the east side of the Tibetan plateau. Westerly (eastward moisture flux) occurs mostly over oceans and is relatively intense in the Southern Hemisphere. The direction of moisture flux
Fig. 5.2.3. Zonal component of total divergent moisture flux of composite January (\(10^2\text{kg s}^{-1}\text{m}^{-1}\)).
In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.03 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.1 are for higher and lower values, respectively.
Fig. 5.2.3. Continued
Fig. 5.2.4. Zonal component of total divergent moisture flux of composite July \((10^2 \text{kg s}^{-1} \text{m}^{-1})\). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.05 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.1 are for higher and lower values, respectively.
Fig. 5.2.4. Continued
in the lower layer is opposite to that in the upper layer, especially off the east and west coasts of continents. This indicates moisture is transported from oceans to continents in the lower layer, particularly off the west coasts of South America and Africa.

In July relatively weak easterly flow in the upper layer occurs off the west coasts of North and South America, Europe and Africa but relatively intense easterly flows exist to the east of Africa. Intense westerly flow occurs in the middle of oceans such as the Pacific and Atlantic Oceans. In the lower layer moisture flows out from North America but into South America relatively intensely. Along the west and southeast coasts of Africa and the west coast of Australia there are more intense moisture flows from oceans than along the rest coasts of respective continents. Only one exception occurs along the east coast of Africa to the south of the Arabian Peninsula where there is strong eastward moisture flow into the Indian Ocean (West Arabian Sea).

Another way to look at the zonal divergent component is to draw the longitude-height cross section at certain latitudes as shown in Figs. 5.2.5a and 5.2.5b. To take seasonal variation into account, 15°S and the equator are chosen for January and July, respectively. The node represents zero of this flux. The dot denotes the location of the ascending branch and the cross denotes that of the descending branch. From the longitude distribution of nodes, the zonal cells can be deduced. This is the so-called "Walker" circulation. It should be noted that the "vertical" flux is deduced from ignoring the meridional divergent component which can mask parts of zonal cells. The main causes that produce this phenomenon might be the sea-land heating contrast and orographic effects. For example, relatively cold sea surface temperature occurs to the west of South America and Africa where downward motion takes place. The upward motion can be seen in the east part of South America and to the east of Africa. The ascending branch at Australia in January or to the east of New Guinea in July is related to convective motion.

Figs. 5.2.6 and 5.2.7 show meridional components of the total divergent moisture flux for January and July, respectively. In January strong southerly (northward flux) in the upper layer occurs to the north of the intertropical convergence zone (ITCZ) in the Indian Ocean. Intense
Fig. 5.2.5a. Longitudinal distribution of nodes at which the zonal component of total divergent moisture flux vanishes along 15°S in January. The arrow indicates the implied Walker circulation. • and x denote the upward and downward branch, respectively.
Fig. 5.2.5b. Longitudinal distribution of nodes at which the zonal component of total divergent moisture flux vanishes along equator in July. The arrow indicates the implied Walker circulation. * and x denote the upward and downward branch, respectively.
Fig. 5.2.6. Meridional component of total divergent moisture flux of composite January ($10^2$ kg s$^{-1}$m$^{-1}$). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.04 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.1 are for higher and lower values, respectively.
Fig. 5.2.7. Meridional component of total divergent moisture flux of composite July ($10^2$ kg s$^{-1}$m$^{-1}$). In upper layer dashed lines are for 0; solid and dotted lines at contour of 0.05 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.15 are for higher and lower values.
moisture flow, therefore, moves into the southern part of Asia. In the southern coasts of the east part of Asia and North America there exists moisture flow into continents from oceans. The remaining portions of continents mainly have moisture outflow. In the lower layer, the ITCZ in the Indian Ocean makes the moisture move out from continents, i.e., Indian subcontinent and Australia, to the ocean. The flow direction is opposite for the remaining coasts of the continents.

In July moisture of the upper layer flows out from North America and the southern part of Africa but into South America, the Indian subcontinent, East Asia, West Europe and the north part of Africa. In the lower layer moisture moves into North America, the west part of North Africa and Europe, part of the south coast of Australia and the south-east coast of Asia, but out from continents elsewhere. From meridional components in both layers the location of ITCZ in the Indian Ocean is at about 6°N.

To determine the contribution from the standing and transient moisture fluxes to the total, their respective potential and stream functions must be calculated first. From their definition, the corresponding fluxes of divergent and rotational components can then be obtained. The standing part is presented followed by the transient part.

In Figs. 5.2.8 and 5.2.9 the standing patterns of potential functions of January and July for both the upper and lower layers are very similar to the total, as shown in Figs. 5.2.1 and 5.2.2, except for some minor differences. This indicates that most of the divergent part of the total moisture flux is carried by its standing part.

However, transient potential functions in January for both layers in Fig. 5.2.10 show that all sources are located in tropical areas except one in the upper layer over Australia. We can expect that transient motion transports moisture poleward. In July Fig. 5.2.11 presents the similar pattern with some centers of sources shifted to the north of those in January.

The standing zonal component of divergent moisture flux is shown in Figs. 5.2.12 and 5.2.13 for January and July, respectively. It can be visualized that this component is almost the same as its total shown in Figs. 5.2.3 and 5.2.4. Therefore, the transient mode is much weaker than the standing mode.
Fig. 5.2.8. Potential function of standing moisture flux of composite January \(10^8\text{ kg s}^{-1}\).
In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.1 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.3 are for higher and lower values, respectively.
Fig. 5.2.9. Potential function of standing moisture flux of composite July (10^8 kg s^{-1}). In upper layer dashed lines are for -0.2; solid and dotted lines at contour interval of 0.15 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0.6; solid and dotted lines at contour interval of 0.3 are for higher and lower values, respectively.
Fig. 5.2.9. Continued
Fig. 5.2.10. Potential function of transient moisture flux of composite January \((10^8 \text{ kg s}^{-1})\). In upper layer dashed lines are for -0.045; solid and dotted lines at contour interval of 0.015 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for -0.4; solid and dotted lines at contour interval of 0.1 are for higher and lower values, respectively.
Fig. 5.2.10. Continued
Fig. 5.2.11. Potential function of transient moisture flux of composite July \( (10^8 \text{ kg s}^{-1}) \).
In upper layer dashed lines are for -0.12; solid and dotted lines at contour interval of 0.02 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for -0.75; solid and dotted lines at contour interval of 0.1 are for higher and lower values, respectively.
Fig. 5.2.11. Continued
Fig. 5.2.12. Zonal component of standing divergent moisture flux of composite January ($10^2$ kg s$^{-1}$ m$^{-1}$). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.04 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.1 are for higher and lower values, respectively.
Fig. 5.2.12. Continued
Fig. 5.2.13. Zonal component of standing divergent moisture flux of composite July ($10^2$ kg s$^{-1}$ m$^{-1}$). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.05 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.1 are for higher and lower values, respectively.
The transient zonal moisture flux in January, as shown in Fig. 5.2.14, is from oceans to continents, particularly North America and the west coasts of Europe and Africa. It should be noted that the mixing ratio of water vapor across the west coasts decreases in the eastward direction. Therefore, positive transient flux implies that there is downgradient moisture transport into continental air; otherwise, negative flux along the west coast means upgradient transport into marine air. According to this definition a few weak upgradient transports occur in both the upper and lower layers, e.g., along the west coasts of South America, and Africa. The most distinctive transport in the lower layer from ocean to continent occurred over North America.

In July the zonal transient flux shown in Fig. 5.2.15 exhibits upgradient transport in a few places, e.g., southern part of South America. Relatively intense transport onto continents from the west coasts of South America and Africa can be seen in both upper and lower layers.

The standing meridional component of divergent moisture flux is shown in Figs. 5.2.16 and 5.2.17 for both January and July. Attention is paid only to coasts with east-west orientation. In January, moisture in the upper layer is essentially blown out from North and South America, but onto Africa, Eurasia and Australia except part of south coast of Africa. In the lower layer moisture is transported into North and South America, but out from the south coast of Asia, South America, and northeast and southwest coasts of Africa.

In July moisture in the upper layer is transported out from North America, the northeast and central west coasts of Africa, part of the North coasts of Eurasia and the south coast of Australia but onto continents elsewhere. In the lower layer intense inflow to North America and the south coast of West and East Africa, but weak outflow elsewhere from continents can be observed in Fig. 5.2.17.

Since the mean moisture essentially decreases poleward, positive transient meridional flux in the Northern Hemisphere indicates downgradient transport. Figs. 5.2.18 and 5.2.19 show that the transient mode of meridional flux is essentially downgradient in the upper and
Fig. 5.2.14. Zonal component of transient divergent moisture flux of composite January (10^2 kg s^{-1}m^{-1}). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.01 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.03 are for higher and lower values, respectively.
Fig. 5.2.14. Continued
Fig. 5.2.15. Zonal component of transient divergent moisture flux of composite July \((10^2 \text{kg s}^{-1}\text{m}^{-1})\). In upper layer dashed lines are for 0; solid and do-ted lines at contour interval of 0.01 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.02 are for higher and lower values, respectively.
Fig. 5.2.15. Continued
Fig. 5.2.16. Meridional component of standing divergent moisture flux of composite January (10^2 kg s^{-1} m^{-1}). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.04 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.1 are for higher and lower values, respectively.
Fig. 5.2.17. Meridional component of standing divergent moisture flux of composite July ($10^2$ kg s$^{-1}$m$^{-1}$). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.07 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.2 are for higher and lower values, respectively.
Fig. 5.2.18. Meridional component of transient divergent moisture flux of composite January \((10^2 \text{kg s}^{-1} \text{m}^{-1})\). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.01 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.05 are for higher and lower values, respectively.
Fig. 5.2.19. Meridional component of transient moisture flux of composite July $(10^2 \text{kg s}^{-1} \text{m}^{-1})$. In upper layer dashed lines are for $0$; solid and dotted lines at contour interval of $0.02$ are for higher and lower values, respectively. In lower layer (continued) dashed lines are for $0$; solid and dotted lines at contour interval of $0.06$ are for higher and lower values, respectively.
lower layers for both January and July over the whole globe. Another feature is that the most intense transports occur in middle latitudes and are associated with cyclonic motions in January. The whole pattern is however shifted northward in July. There are also more areas of negative transient meridional flux in northern middle latitudes in July than in January. The transport in July is therefore not so apparently downgradient as in January.

The stream function is related to the rotational component of the moisture flux. However, the large-scale atmospheric motion is quasirotational. Major patterns of general circulation can still be visualized under modification of moisture distribution. Therefore, only stream functions of total moisture flux are shown in Figs. 5.2.20 and 5.2.21.

In January most of the anticyclones in the upper layer are located in the northern tropics and subtropics, i.e., over the Pacific Ocean to the southeast of Asia, over Central Africa, and to the north of South America. One anticyclone can be seen in southern high latitudes. Cyclones are located over central South America, to the east of Africa, over north Australia, and over North America. Westerlies occur in middle latitudes but those in the Northern Hemisphere are relatively strong. In the lower layer there are two subtropical anticyclones in the Northern Hemisphere and three polar anticyclones at high latitudes. Most cyclones are over the oceans except two, one over northwest Europe and the other over northeast Australia. Westerlies can be seen at middle latitudes but easterlies occur in tropical areas.

In July westerlies in the upper layer of the Southern Hemisphere are more intense than that in the Northern Hemisphere. There are four centers for anticyclones located in northern subtropics, and only one cyclone occurs over the tropical South Pacific Ocean. At southern middle latitude there is one anticyclone. In the lower layer there exists a cyclonic band in the southern tropics. Equatorial easterlies, northern subtropical anticyclones, middle latitude westerlies, cyclones and anticyclones at higher latitudes can be seen. The seasonal variation of the wind system in the Indian Ocean to the east of Africa can be easily distinguished in the lower layer.

From the stream function we can compute rotational components of the total moisture flux. The zonal component is shown in Figs. 5.2.22 and 5.2.23 for both January and July.
Fig. 5.2.20. Stream function of total moisture flux of composite January ($10^8$ kg s$^{-1}$). In upper layer dashed lines are for 2.5; solid and dotted lines at contour interval of 0.3 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 2.2; solid and dotted lines at contour interval of 0.3 are for higher and lower values, respectively.
Fig. 5.2.20. Continued
Fig. 5.2.21. Stream function of total moisture flux of composite July ($10^8$ kg s$^{-1}$). In upper layer dashed lines are for 1.6; solid and dotted lines at contour interval of 0.3 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 1.25; solid and dotted lines at contour interval of 0.5 are for higher and lower values, respectively.
Fig. 5.2.22. Zonal component of total rotational moisture flux of composite January ($10^2$ kg s$^{-1}$ m$^{-1}$). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.3 are for higher and lower values, respectively. In lower layer (continued) dashed lined are for 0; solid and dotted lines at contour interval of 0.5 are for higher and lower values, respectively.
Fig. 5.2.22. Continued
Fig. 5.2.23. Zonal component of total rotational moisture flux of composite July ($10^2$ kg s$^{-1}$ m$^{-1}$). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.35 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.5 are for higher and lower values, respectively.
Fig. 5.2.23. Continued
In January relatively intense westerlies in the upper layer occur around 30°N. Easterlies in the tropics are stronger than those in polar areas due to poleward decrease of moisture content. Tropical easterlies, strong middle latitude westerlies and weak polar easterlies can be seen in the lower layer.

In July similar structures can be seen. However, westerlies in the upper layer are more intense in the Southern Hemisphere than in the Northern Hemisphere. In the lower layer strong westerlies can be found to the south of Asia. These two facts are obviously related to the seasonal variation.

Meridional components of January and July are shown in Figs. 5.2.24 and 5.2.25, respectively. In January relatively intense northward fluxes in the upper layer occur in the east or southeast coasts of continents, e.g., North and South America, Southeast Asia, and Australia. Intense northerlies occur in oceanic regions and at the west coasts of continents, e.g., South America, South Africa and Australia. In the lower layer relatively intense southerlies occur at the west and south coasts of continents. Most of the intense northerlies can be found in oceanic regions.

In July isotachs of the upper layer are elongated in the east-west direction more over oceanic regions than over continents. This might be due to an orographic effect. In the lower layer northward flux to the east of Africa occurs in response to the seasonal variation of the monsoon.

5.3 Vertical transport of water substance in the atmosphere

In this section all vertical transports by large-scale and subgrid-scale processes are presented except surface evaporation and precipitation. In the present version of the model, the large-scale condensation in the upper layer completely evaporates in the lower, so that it is also a transport process related to phase change.

In Fig. 5.3.1 is the total vertical transport by large-scale motion, \((\pi/g)\vec{\delta}_2\). When it is negative (positive) the lower layer loses (gains) water vapor and the upper layer gains (loses) the same amount. In January there is a downward band within the northern tropics
Fig. 5.2.24. Meridional component of total rotational moisture flux of composite July ($10^2$ kg s$^{-1}$ m$^{-1}$). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.1 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.2 are for higher and lower values, respectively.
Fig. 5.2.24. Continued
Fig. 5.2.25. Meridional component of total rotational moisture flux of composite July ($10^2$ kg s$^{-1}$ m$^{-1}$). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.05 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 0.25 are for higher and lower values, respectively.
Fig. 5.2.25. Continued
Fig. 5.3.1. Total large-scale vertical transport \((10^{-5}\text{kg s}^{-1}\text{m}^{-2})\) in January and July (continued). Dashed lines are for 0; solid and dotted lines at contour interval of 1 are for higher and lower values, respectively.
and subtropics. Three areas of downward transport are located to the west of continents where oceanic deserts of low precipitation can be expected. Since descending motion is heated in a stably stratified atmosphere, the chance of supersaturation is much reduced. Precipitation is rare in those areas. There are also scattered areas of downward transport over high latitudes. Areas of intense upward transport are mostly located over the Indian Ocean and southern continents and oceans between the equator and 30°S. It can be seen that the band of the Pacific Ocean is separated into two branches by the southern subtropic anticyclone. There is a similar pattern over the south Atlantic Ocean. In the Northern Hemisphere, upward transports occur to the east of mountain ranges, i.e., Rocky Mountains and Tibetan Plateau, and in the track of cyclonic motion.

In July the downward transports occur to the west of continents, in the southern subtropics and in the west or north side of Tibetan, Greenland and Antarctica mountains (highlands). The remaining areas are scattered over high latitudes. Areas of upward transport are shifted northward in general.

To further understand the vertical transports by large-scale motion the standing and transient modes are shown in Figs. 5.3.2 and 5.3.3, respectively. In January areas of standing upward transport in the low latitudes are interrupted by those of downward transport along the west coasts of continents. Subtropic bands of descending motion are also broken by ascending motion induced by cyclones over oceans and the lee side of mountains. In the northern middle latitudes, the upward motion is followed by downward motion. However, this pattern does not happen so obviously in the southern middle latitudes. It might be due to different size of continents over both hemispheres. Downward transport is dominant over polar regions. In July generally similar patterns can be observed except they are shifted northward, and the Northern subtropic band of downward transport is broken into several pieces.

From the standing part of the vertical transport by the large-scale motion, Hadley and Ferrel cells are interrupted by Walker circulation which is a planetary scale flow. This planetary scale flow is influenced by orography and continent-ocean heating contrasts so that
Fig. 5.3.2. Standing large-scale vertical transport \((10^{-5}\text{kg s}^{-1}\text{m}^{-2})\) in January and July (continued). Dashed lines are for 0; solid and dotted lines at contour interval of 1 are for higher and lower values, respectively.
Fig. 5.3.2. Continued
Fig. 5.3.3. Transient large-scale vertical transport ($10^{-5}$ kg s$^{-1}$ m$^{-2}$) in January and July (continued). Dashed lines are for 0; solid lines at contour interval of 0.2 are for higher values; dotted lines at contour interval of 0.3 are for lower values.
Fig. 5.3.3. Continued
it is highly longitude dependent. Thus meridional cells are also longitude dependent. Seasonal variation of intensity and location of intense upward transport is very significant in low latitudes.

Since moisture content is decreasing with height, the negative (positive) transient transport is downgradient (upgradient) or upward (downward). In January most of the intense upward transports ($\leq -0.9 \times 10^{-5} \text{kJ s}^{-1} \text{m}^{-2}$) are distributed over middle latitudes. This is due to baroclinic processes of cyclone motions. There are upgradient transports scattered over low latitudes except for a wide area over central Africa. In July areas of intense upward transport ($\leq -0.9 \times 10^{-5} \text{kg s}^{-1} \text{m}^{-2}$) are changed in location and deformed in shape, and even disappear, i.e., the one which occurs in North America in January. However, areas of upgradient transport are expanded, particularly those in Africa, Indian Ocean and Asia which are connected.

Large-scale condensation in the upper layer is shown in Fig. 5.3.4. In January most of the intense ($\geq 0.9 \times 10^{-5} \text{kg s}^{-1} \text{m}^{-2}$) activity occurs in the northern middle latitudes, e.g., in the northwest Pacific Ocean with center at the west coast of North America (115°W, 34°N) and in Europe over the Black Sea and part of the Mediterranean Sea. There is one intense area over East Africa. Since large-scale condensation is associated with cyclone motion, it occurs more frequently in middle latitudes than elsewhere. In July areas of intense activity ($\geq 0.9 \times 10^{-5} \text{kg s}^{-1} \text{m}^{-2}$) are located in the southern middle latitudes and in the Indian Ocean and Central Africa. According to the seasonal variation, large-scale condensation in the upper layer is more intense in the winter hemisphere than in the summer one.

There are two subgrid-scale vertical transports in the model atmosphere. One is by middle-level convection and the other by penetrating convection. Both are shown in Figs. 5.3.5 and 5.3.6.

In January strong middle-level convective transport ($\geq 0.4 \times 10^{-5} \text{kg s}^{-1} \text{m}^{-2}$) occurs between 30°S and the equator. In these latitude belts the interruption by large-scale downward motion as shown in Fig. 5.3.2 can be visualized. In July all of the areas of strong activity are between the equator and 50°N. This indicates that seasonal variation of middle-level convection is very significant in the model atmosphere.
Fig. 5.3.4. Large-scale condensation rate ($10^{-5}$ kg s$^{-1}$ m$^{-2}$) in upper layer of January and July (continued). Dashed lines are for 0.3; solid lines at contour interval of 0.6 are for higher values.
Fig. 5.3.4. Continued
Fig. 5.3.5. Vertical transport \((10^{-5}\, \text{kg s}^{-1}\, \text{m}^{-2})\) by middle-level convection in January and July (continued). Dashed lines are for 0.1; solid lines at contour interval of 0.3 and 0.6 are for higher values of January and July, respectively.
Fig. 5.3.5. Continued
Fig. 5.3.6. Vertical transport \((10^{-5} \text{kg s}^{-1} \text{m}^{-2})\) by penetrating convection in January and July (continued). Dashed lines are for 0.5; solid lines at contour interval of 1 and 2.5 are for higher values of January and July, respectively.
Fig. 5.3.6. Continued
Areas of strong penetrating convective transport ($\geq 1.5 \times 10^{-5} \text{ kg s}^{-1} \text{m}^{-2}$) in January are highly coincident with those of strong surface evaporation ($\geq 3 \times 10^{-5} \text{ kg sec}^{-1} \text{m}^{-2}$) (in Fig. 5.4.8). Also areas of strong activity, e.g., in equatorial Indian Ocean and near South Africa are correlated to low-level convergence (in Fig. 5.4.1). Thus penetrating convection is favored not only by surface evaporation but also by low-level convergence due to cyclone motion, particularly in middle latitudes. All of these activities occur in middle oceanic regions and to the east or south of continents.

In July most areas of intense transport ($\geq 3 \times 10^{-5} \text{ kg s}^{-1} \text{m}^{-2}$) are located in oceanic regions, especially in Indian Ocean. The similar correlation with surface evaporation and low-level convergence is also observed. However, the seasonal variation can be seen in the northern hemisphere to the east or south of continents. The intensity and location of strong activity in equatorial Indian Ocean is changed significantly from January to July.

### 5.4 Atmospheric water vapor balance

To facilitate the comprehension and interpretation of the results, each term in the balance equation of atmospheric water vapor is calculated. Before showing the geographical distribution of each physical process the average over the globe and hemispheres is presented.

Tables 5.4.1a and 5.4.1b are global and hemispheric balances of January for both layers, while Tables 5.4.2a and 5.4.2b are those of July. The first number in each box is the composite mean and the second is the composite standard deviation. In January large-scale upward transport is the primary source for the upper layer. The major sink term is large-scale condensation except in the Southern Hemisphere where it is the net effect of middle-level convection. The interannual variability of vertical transport is the largest, although that of horizontal transport is the second. It should also be noted that middle-level convection has larger interannual variability than large-scale condensation except in the Northern Hemisphere. In the lower layer the major source is surface evaporation. The primary sink is the net effect of penetrating convection, and the secondary
Table 5.4.1a. Global and hemispheric water vapor balance of the upper atmospheric layer in January

<table>
<thead>
<tr>
<th>Units: $10^{-5} \frac{kg}{s \cdot m^2}$</th>
<th>Northern Hemisphere</th>
<th>Southern Hemisphere</th>
<th>Global atmosphere</th>
</tr>
</thead>
<tbody>
<tr>
<td>$-\mathbf{\nabla} \cdot \left( \frac{\Delta \sigma}{g} \mathbf{V} q \right)_1$</td>
<td>$0.1216 \pm 0.2818$</td>
<td>$-0.1216 \pm 0.3208$</td>
<td>$0.0000 \pm 0.3013$</td>
</tr>
<tr>
<td>$-\frac{\partial}{\partial \sigma} \left( \frac{\Delta \sigma}{g} \mathbf{\nabla} q \right)_1$</td>
<td>$0.3779 \pm 0.3772$</td>
<td>$0.9758 \pm 0.5313$</td>
<td>$0.6769 \pm 0.4542$</td>
</tr>
<tr>
<td>$-\frac{\pi \Delta \sigma}{g} \mathbf{C}_1$</td>
<td>$-0.3487 \pm 0.1319$</td>
<td>$-0.2685 \pm 0.1066$</td>
<td>$-0.3086 \pm 0.1193$</td>
</tr>
<tr>
<td>$\left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_1 \right) \right]_{MLC}$</td>
<td>$-0.0905 \pm 0.0660$</td>
<td>$-0.4144 \pm 0.2263$</td>
<td>$-0.2525 \pm 0.1461$</td>
</tr>
<tr>
<td>$\left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_1 \right) \right]_{PC}$</td>
<td>$-0.0652 \pm 0.0757$</td>
<td>$-0.1517 \pm 0.0940$</td>
<td>$-0.1085 \pm 0.0848$</td>
</tr>
<tr>
<td>$\frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_1 \right)$</td>
<td>$-0.0048 \pm 0.0831$</td>
<td>$0.0196 \pm 0.0895$</td>
<td>$0.0074 \pm 0.0863$</td>
</tr>
</tbody>
</table>

\[
\frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_1 \right) = -\mathbf{\nabla} \cdot \left( \frac{\pi \Delta \sigma}{g} \mathbf{V} q \right)_1 - \frac{\partial}{\partial \sigma} \left( \frac{\pi \Delta \sigma}{g} \mathbf{\nabla} q \right)_1 - \frac{\pi \Delta \sigma}{g} \mathbf{C}_1
\]

\[ + \left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_1 \right) \right]_{MLC} + \left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_1 \right) \right]_{PC} \]
Table 5.4.lb. Global and hemispheric water vapor balance of the lower atmospheric layer in January

<table>
<thead>
<tr>
<th>Units: $10^{-5}$ kg s m$^{-2}$</th>
<th>Northern Hemisphere</th>
<th>Southern Hemisphere</th>
<th>Global atmosphere</th>
</tr>
</thead>
<tbody>
<tr>
<td>$-\nabla \cdot \left( \frac{\pi \Delta \sigma}{g} V q \right)_3$</td>
<td>-0.5196 ± 0.9918</td>
<td>0.5196 ± 1.2819</td>
<td>-0.0000 ± 1.1369</td>
</tr>
<tr>
<td>$\frac{\partial}{\partial \sigma} \left( \frac{\pi \Delta \sigma}{g} \delta q \right)_3$</td>
<td>-0.3779 ± 0.3772</td>
<td>-0.9758 ± 0.5313</td>
<td>-0.6759 ± 0.4542</td>
</tr>
<tr>
<td>$- \frac{\pi \Delta \sigma}{g} C'_3$</td>
<td>-1.0038 ± 0.3276</td>
<td>-0.7293 ± 0.2333</td>
<td>-0.8665 ± 0.2804</td>
</tr>
<tr>
<td>$\left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_3 \right) \right]_{MLC}$</td>
<td>-0.1435 ± 0.1095</td>
<td>-0.6758 ± 0.3583</td>
<td>-0.4097 ± 0.2339</td>
</tr>
<tr>
<td>$\left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_3 \right) \right]_{PC}$</td>
<td>-1.5781 ± 0.6831</td>
<td>-1.2835 ± 0.6589</td>
<td>-1.4308 ± 0.6710</td>
</tr>
<tr>
<td>$E_s$</td>
<td>3.2445 ± 0.4409</td>
<td>2.9709 ± 0.3953</td>
<td>3.1077 ± 0.4181</td>
</tr>
<tr>
<td>$\frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_3 \right)$</td>
<td>-0.0298 ± 0.4935</td>
<td>0.0946 ± 0.4651</td>
<td>0.0324 ± 0.4793</td>
</tr>
</tbody>
</table>

$$\frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_3 \right) = -\nabla \cdot \left( \frac{\pi \Delta \sigma}{g} V q \right)_3 - \frac{\partial}{\partial \sigma} \left( \frac{\pi \Delta \sigma}{g} \delta q \right)_3 - \frac{\pi \Delta \sigma}{g} (C'_3 - C_1)$$

$$+ \left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_3 \right) \right]_{MLC} + \left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_3 \right) \right]_{PC} + E_s$$
Table 5.4.2a. Global and hemispheric water vapor balance of the upper atmospheric layer in July

<table>
<thead>
<tr>
<th>Units: $10^{-5}$ kg m⁻² s⁻¹</th>
<th>Northern Hemisphere</th>
<th>Southern Hemisphere</th>
<th>Global atmosphere</th>
</tr>
</thead>
<tbody>
<tr>
<td>$-\nabla \cdot \left( \frac{\Delta \sigma}{g} Vq \right)_1$</td>
<td>$-0.2873 \pm 0.3252$</td>
<td>$0.2873 \pm 0.2337$</td>
<td>$-0.0000 \pm 0.2794$</td>
</tr>
<tr>
<td>$-\frac{\partial}{\partial \sigma} \left( \frac{\Delta \sigma}{g} \delta q \right)_1$</td>
<td>$1.0903 \pm 0.4676$</td>
<td>$0.1679 \pm 0.2929$</td>
<td>$0.6291 \pm 0.3803$</td>
</tr>
<tr>
<td>$-\frac{\Delta \sigma}{g} C_1$</td>
<td>$-0.1685 \pm 0.0866$</td>
<td>$-0.3406 \pm 0.1069$</td>
<td>$-0.2545 \pm 0.0968$</td>
</tr>
<tr>
<td>$[\frac{\partial}{\partial t} \left( \frac{\Delta \sigma}{g} q_1 \right)]_{MLC}$</td>
<td>$-0.4805 \pm 0.2078$</td>
<td>$-0.0645 \pm 0.0454$</td>
<td>$-0.2725 \pm 0.1266$</td>
</tr>
<tr>
<td>$[\frac{\partial}{\partial t} \left( \frac{\Delta \sigma}{g} q_1 \right)]_{PC}$</td>
<td>$-0.1472 \pm 0.1050$</td>
<td>$-0.0721 \pm 0.0588$</td>
<td>$-0.1096 \pm 0.0819$</td>
</tr>
<tr>
<td>$\frac{\partial}{\partial t} \left( \frac{\Delta \sigma}{g} q_1 \right)$</td>
<td>$0.0068 \pm 0.1269$</td>
<td>$-0.0220 \pm 0.0929$</td>
<td>$-0.0076 \pm 0.1099$</td>
</tr>
</tbody>
</table>
Table 5.4.2b. Global and hemispheric water vapor balance of the lower atmospheric layer in July

<table>
<thead>
<tr>
<th>Units: $10^{-5} \ \frac{kg}{s \ m^2}$</th>
<th>Northern Hemisphere</th>
<th>Southern Hemisphere</th>
<th>Global atmosphere</th>
</tr>
</thead>
<tbody>
<tr>
<td>$-\nabla \cdot \left( \frac{\Delta \sigma}{g} \ \nabla q \right)_3$</td>
<td>0.9181 ± 1.0583</td>
<td>-0.9181 ± 0.7974</td>
<td>-0.0000 ± 0.9278</td>
</tr>
<tr>
<td>$-\frac{\partial}{\partial \sigma} \left( \frac{\pi \Delta \sigma}{g} \ \delta q \right)_3$</td>
<td>-1.0903 ± 0.4676</td>
<td>-0.1679 ± 0.2929</td>
<td>-0.6291 ± 0.3803</td>
</tr>
<tr>
<td>$-\frac{\pi \Delta \sigma}{g} C_3^1$</td>
<td>-0.3823 ± 0.1494</td>
<td>-0.9474 ± 0.2519</td>
<td>-0.6648 ± 0.2007</td>
</tr>
<tr>
<td>$\left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_3 \right) \right]_{MLC}$</td>
<td>-0.9386 ± 0.3871</td>
<td>-0.1206 ± 0.0857</td>
<td>-0.5296 ± 0.2364</td>
</tr>
<tr>
<td>$\left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_3 \right) \right]_{PC}$</td>
<td>-1.7855 ± 0.9165</td>
<td>-1.4387 ± 0.6357</td>
<td>-1.6121 ± 0.7761</td>
</tr>
<tr>
<td>$E_S$</td>
<td>3.0873 ± 0.4540</td>
<td>3.1727 ± 0.3356</td>
<td>3.1300 ± 0.3948</td>
</tr>
<tr>
<td>$\frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} q_3 \right)$</td>
<td>-0.0228 ± 0.8521</td>
<td>-0.0795 ± 0.5037</td>
<td>-0.0512 ± 0.6779</td>
</tr>
</tbody>
</table>
sink is the large-scale condensation except in the Southern Hemisphere where it is the large-scale vertical transport. The interannual variability of horizontal transport is however the largest, that of penetrating convection the second, and that of vertical transport the third except for surface evaporation in the Northern Hemisphere.

It should be pointed out that the storage term has the smallest magnitude in the mean compared with all other physical processes. Therefore its interannual variability was not included in the ranking of the standard deviation. Also the global mean of horizontal transport is required to be zero by integral constraint, and the hemispheric means should be equal in magnitude but opposite in sign.

In July the primary source of the upper layer (Table 5.4.2a) is large-scale vertical transport in the Northern Hemisphere, but is large-scale horizontal transport in the Southern Hemisphere. This is because the subsidence in the southern subtropics in July is relatively stronger than that in the northern subtropics in January. As seen in Fig. 5.3.1, the downward branch of the Hadley cell over the Indian and South Pacific oceans in July is stronger than that over the Asian continent and North Pacific ocean in January. Therefore the cross-equator transport from the Northern Hemisphere becomes the major source of southern upper layer. The primary sink is middle-level convection in the Northern Hemisphere, however, in the Southern Hemisphere is large-scale condensation.

In the lower layer the major source is surface evaporation. The largest sink is penetrating convection. The secondary sink is large-scale vertical transport in the Northern Hemisphere but in the Southern Hemisphere is large-scale condensation.

In the upper layer the interannual variability of large-scale vertical transport is the largest, that of horizontal transport is the second and that of middle-level convection is the third except that of large-scale condensation is the third in the Southern Hemisphere. In the lower layer horizontal convergence has the largest interannual variability, penetrating convection has the second and surface evaporation has the third except large-scale vertical transport in the Northern Hemisphere.

Since most of the moisture is in the lower layer, the budget in the lower layer is more important than that in the upper layer. It can
be expected that variability of low-level wind can cause the largest variability in horizontal convergence. The most intense precipitation process due to penetrating convection can then have the second largest interannual variability. Therefore, the anomalous behavior in low-level circulation can cause flood or drought over continents.

In summary, the global balance of atmospheric water vapor gives the following features:

1. Only large-scale vertical transport moistens the upper troposphere, the lower layer can be moistened by surface evaporation and large-scale condensation from the upper layer.
2. The largest contribution to total precipitation is from penetrating convection while the smallest contribution to total precipitation is from middle-level convection in January and from large-scale precipitation in July.
3. The atmosphere is slightly moistened in January but slightly dried in July.

The hemispheric balances bring out the following features:

1. Horizontal moisture transport is from winter to summer hemisphere.
2. Large-scale vertical moisture transport in the summer hemisphere is larger than that in the winter hemisphere.
3. Surface evaporation is weaker in the summer hemisphere than in the winter one.
4. Large-scale condensation favors to occur in the winter hemisphere.
5. Middle level convection is much more intense in the summer hemisphere than in the winter one.
6. Penetrating convection in the lower layer of the Northern Hemisphere is always stronger than that of the Southern Hemisphere while this kind of convection in the upper layer is relatively intense in the summer hemisphere.

To know the relative magnitude of seasonal variation for each process the vertically integrated moisture balance over both hemispheres is shown in Tables 5.4.3a and 5.4.3b. It is obvious that horizontal moisture convergence and middle-level convection are primarily responsible for the seasonal variation.
Table 5.4.3a. Seasonal variation of vertically integrated moisture balance over Northern Hemisphere

<table>
<thead>
<tr>
<th>Units: $10^{-5} \frac{\text{kg}}{\text{s m}^2}$</th>
<th>January</th>
<th>July</th>
<th>Difference*</th>
</tr>
</thead>
<tbody>
<tr>
<td>$-\nabla \cdot \left[ \frac{\pi \Delta \sigma}{g} (V_1 q_1 + V_2 q_3) \right]$</td>
<td>-0.3980</td>
<td>0.6308</td>
<td>1.0288</td>
</tr>
<tr>
<td>$- \frac{\pi \Delta \sigma}{g} C_3$</td>
<td>-1.0038</td>
<td>-0.3823</td>
<td>0.6215</td>
</tr>
<tr>
<td>$\left{ \frac{\partial}{\partial t} \left[ \frac{\pi \Delta \sigma}{g} (q_1 + q_3) \right] \right}_{MLC}$</td>
<td>-0.2340</td>
<td>-1.4191</td>
<td>-1.1851</td>
</tr>
<tr>
<td>$\left{ \frac{\partial}{\partial t} \left[ \frac{\pi \Delta \sigma}{g} (q_1 + q_3) \right] \right}_{PC}$</td>
<td>-1.6433</td>
<td>-1.9327</td>
<td>-0.2894</td>
</tr>
<tr>
<td>$E_S$</td>
<td>3.2445</td>
<td>3.0873</td>
<td>-0.1572</td>
</tr>
<tr>
<td>$\frac{\partial}{\partial t} \left[ \frac{\pi \Delta \sigma}{g} (q_1 + q_3) \right]$</td>
<td>-0.0346</td>
<td>-0.0160</td>
<td>0.0186</td>
</tr>
</tbody>
</table>

* Difference = (July) - (January)
Table 5.4.3b. Seasonal variation of vertically integrated moisture balance over Southern Hemisphere

<table>
<thead>
<tr>
<th>Units: $10^{-5} \frac{\text{kg}}{\text{s m}^2}$</th>
<th>January</th>
<th>July</th>
<th>Difference*</th>
</tr>
</thead>
<tbody>
<tr>
<td>$-\nabla \cdot \left[ \frac{\pi \Delta \sigma}{g} \mathbf{V}_{v} (q_1 + q_3) \right]$</td>
<td>0.3980</td>
<td>-0.6308</td>
<td>-1.0288</td>
</tr>
<tr>
<td>$-\frac{\pi \Delta \sigma}{g} C_3$</td>
<td>-0.7293</td>
<td>-0.9474</td>
<td>-0.2181</td>
</tr>
<tr>
<td>$\left{ \frac{\partial}{\partial t} \frac{\pi \Delta \sigma}{g}(q_1 + q_3) \right}_{MLC}$</td>
<td>-1.0902</td>
<td>-0.1851</td>
<td>0.9051</td>
</tr>
<tr>
<td>$\left{ \frac{\partial}{\partial t} \frac{\pi \Delta \sigma}{g}(q_1 + q_3) \right}_{PC}$</td>
<td>-1.4352</td>
<td>-1.5108</td>
<td>-0.0756</td>
</tr>
<tr>
<td>$E_s$</td>
<td>2.9709</td>
<td>3.1727</td>
<td>0.2018</td>
</tr>
<tr>
<td>$\frac{\partial}{\partial t} \left[ \frac{\pi \Delta \sigma}{g} (q_1 + q_3) \right]$</td>
<td>0.1142</td>
<td>-0.1015</td>
<td>-0.2157</td>
</tr>
</tbody>
</table>

* Difference = (July) - (January)
All of the information mentioned above can serve as a guide to examine the geographic distribution of each process in the balance equations. Since the storage is small on the global average, it can be omitted. Large-scale vertical transport (Fig. 5.3.1) and large-scale condensation in the upper layer (Fig. 5.3.4) have been presented. Only the rest of the processes are shown in this section.

Figs. 5.4.1 and 5.4.2 are geographic distributions of horizontal convergence of moisture flux for both January and July, respectively. In January most areas of intense convergence ($\geq 1 \times 10^{-5}$ kg s$^{-1}$m$^{-2}$) in the upper layer are located to the west of continents except three regions where they are to the southwest of the Tibetan plateau, to the southeast of Asia and to the east of Africa. Intense divergence ($\leq -1 \times 10^{-5}$ kg s$^{-1}$m$^{-2}$) however, occurs to the southeast of the Rocky Mountains and Tibetan plateau, over the continents of the Southern Hemisphere, and over the oceans between 30°S and 30°N. In the lower layer relatively intense divergence ($\leq -3 \times 10^{-5}$ kg s$^{-1}$m$^{-2}$) occurs over the oceans, particularly to the west and the south of continents. Areas of intense convergence ($\geq 1.5 \times 10^{-5}$ kg s$^{-1}$m$^{-2}$) over low latitudes indicate the location of the intertropical convergence zone (ITCZ). Over the south Pacific and Atlantic oceans the NW-SE oriented convergence zones are formed due to the block of subtropic anticyclones. This reveals the interaction between the tropics and extratropics. In the middle and high latitudes cyclonic motion induces moisture convergence especially in windward and lee sides of mountain ranges. It can be visualized that an intense convergence occurs near the southern part of Africa.

In July most of the intense convergence ($\geq 1 \times 10^{-5}$ kg s$^{-1}$m$^{-2}$) in the upper layer takes place in the southern tropical Indian ocean and to the west of continents. The intense divergence ($\leq -1 \times 10^{-5}$ kg s$^{-1}$m$^{-2}$) is mostly located in the tropics. In the lower layer the ITCZ and its seasonal shift can be visualized. Intense divergence ($\leq -4 \times 10^{-5}$ kg s$^{-1}$m$^{-2}$) occurs to the west of continents and in the southern Pacific and Indian oceans. There is also an intense convergence near the southern part of Africa.

Seasonal variation reverses the strong convergence and divergence over the Indian Ocean. This also indicates that a relatively intense
Fig. 5.4.1. Horizontal moisture convergence in January \((10^{-5} \text{kg s}^{-1} \text{m}^{-2})\). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 1 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid and dotted lines at contour interval of 1.5 are for higher and lower values, respectively.
Fig. 5.4.2. Horizontal moisture convergence of July ($10^{-5}$ kg s$^{-1}$ m$^{-2}$). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 1 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for 0; solid lines at contour interval of 4 are for higher values, dotted lines at contour interval of 2 are for lower values.
Fig. 5.4.2. Continued
Hadley cell occurs in the winter hemisphere there. Strong divergence is persistent to the west of continents, i.e., North and South America and Africa.

The net effect of middle-level convection is shown in Figs. 5.4.3 and 5.4.4. In January strong activity is highly connected to the strong low-level horizontal convergence as shown in Fig. 5.4.1. Most of the intense convection \(\leq -1.25 \times 10^{-5} \text{kg s}^{-1}\text{m}^{-2}\) occurs between 30°S and the equator. In July most of the intense middle-level convection \(\leq -1.25 \times 10^{-5} \text{kg s}^{-1}\text{m}^{-2}\) takes place between the equator and 50°N. However, it is not so highly connected to low-level horizontal convergence as that in January. The reason is not clear at this moment. This effect is to essentially dry the atmosphere. It is stronger in the lower layer than in the upper layer because the former is more moist than the latter by an order of magnitude.

The net effect of penetrating convection is shown in Figs. 5.4.5 and 5.4.6 for January and July, respectively. In January the upper layer is dried out mainly between 30°S and 20°N but is weakly moistened mostly in the higher latitudes. In the lower layer the effect of the penetrating convection is to remove the moisture. The relatively intense activity \(\leq -3.5 \times 10^{-5} \text{kg s}^{-1}\text{m}^{-2}\) is highly connected to that of surface evaporation \(\geq 6 \times 10^{-5} \text{kg s}^{-1}\text{m}^{-2}\) as shown in Fig. 5.4.7.

In July the upper layer is dried out mainly in the (northern) tropics, particularly over the Indian ocean, and weak moistening happens in higher latitudes. The net effect on the lower layer is purely drying and the pattern of intense convection \(\leq -6.5 \times 10^{-5} \text{kg s}^{-1}\text{m}^{-2}\) is highly connected to that of strong surface evaporation \(\geq 6 \times 10^{-5} \text{kg s}^{-1}\text{m}^{-2}\) as shown in Fig. 5.4.7.

The composite means of surface evaporation for January and July are shown in Fig. 5.4.7. According to comparison with Fig. 5.1.2, which is for the third year, there is no significant difference in the pattern of strong surface evaporation \(\geq 6 \times 10^{-5} \text{kg s}^{-1}\text{m}^{-2}\).

Large-scale precipitation is the same as the large-scale condensation in the lower layer as shown in Fig. 5.4.8. This kind of precipitation is favored to occur in middle and high latitudes. The intensity is stronger in winter hemisphere than in the summer one. Heavy
Fig. 5.4.3. Net effect of middle-level convection of January ($10^{-5}$ kg s$^{-1}$m$^{-2}$). In upper layer dashed lines are for -0.25; solid lines are for 0; dotted lines at contour interval of 1 are for lower values. In lower layer (continued) dashed lines are for -0.4; dotted lines at contour interval of 1 are for lower values.
Fig. 5.4.3. Continued
Fig. 5.4.4. Net effect of middle-level convection of July ($10^{-5}$ kg s$^{-1}$ m$^{-2}$). In upper layer dashed lines are for -0.25; solid lines are for 0; dotted lines at contour interval of 1 are for lower values. In lower layer (continued) dashed lines are for -0.5; dotted lines at contour interval of 1 are for lower values.
Fig. 5.4.5. Net effect of penetrating convection in January ($10^{-5}$ kg s$^{-1}$ m$^{-2}$). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.3 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for -1.5; dotted lines at contour interval of 2 are for lower values.
Fig. 5.4.5. Continued
Fig. 5.4.6. Net effect of penetrating convection of July \((10^{-5}\text{kg s}^{-1}\text{m}^{-2})\). In upper layer dashed lines are for 0; solid and dotted lines at contour interval of 0.6 are for higher and lower values, respectively. In lower layer (continued) dashed lines are for -1.5; dotted lines at contour interval of 5 are for lower values.
Fig. 5.4.7. Surface evaporation rate \(10^{-5}\text{kg s}^{-1}\text{m}^{-2}\) of composite January and July (continued). Dashed lines are for 3; solid lines at contour interval of 1.5 are for higher values, dotted lines at contour interval of 1 are for lower values. (The conversion factor from \(10^{-5}\text{kg s}^{-1}\text{m}^{-2}\) to mm day\(^{-1}\) is 0.864.)
Fig. 5.4.7. Continued
Fig. 5.4.8. Large-scale condensation (precipitation) \(10^{-5}\text{kg s}^{-1}\text{m}^{-2}\) from lower layer of composite January and July (continued). Dashed lines are for 0.5; solid lines at contour interval of 1.5 and 1 are for higher values of January and July, respectively. (The conversion factor from \(10^{-5}\text{kg s}^{-1}\text{m}^{-2}\) to mm day\(^{-1}\) is 0.864.)
Fig. 5.4.8. Continued
precipitation can occur in the windward side of the mountain ranges, e.g., the Rockies and Andes.

5.5 Model verification

In this section the comparison of model results with observed data is to find discrepancies between them. Verification can be made in this manner. Also the above diagnostic information in the preceding part of this chapter can be used to explain discrepancies to a certain degree of understanding.

First, we define the average "residence time" of atmospheric moisture as

$$\tau_r = \frac{\sum_k q_k \Delta \sigma_k}{\sum_i (P_r)_i}$$

(5.5.1)

where the numerator is the total precipitable water and the denominator is the total precipitation due to different processes, i.e., large-scale condensation, middle-level convection and penetrating convection. As shown in Table 5.5.1 the residence time is longer in July than in January. The mean value is 9.36 days which is very close to the observed value (Eagleson, 1970). From another viewpoint this implies that the entire atmospheric moisture must be replaced 39 times each year. This value is the same as the observed one (Drozdov, 1971) that was obtained by Budyko (1963).

The global mean values of annual mean precipitation are given in Table 5.5.2. The present model value is tentatively approximated by the mean value of composite January and July. The simulated precipitation rate is in good agreement with the observed.

From the above verification of global statistics the model atmosphere is not significantly different from the real one. This supports our hypothesis using the model data in the present study.

Water balance at the earth's surface is shown in Fig. 5.5.1. According to the comparison among different data sources the model result is more close to the observed estimate by Budyko (1963) than
Table 5.5.1. The average residence time of atmospheric moisture

<table>
<thead>
<tr>
<th>Month</th>
<th>$q_1$ (g/kg)</th>
<th>$q_3$ (g/kg)</th>
<th>$\Sigma P_i$ ($\frac{kg}{s m^2}$)</th>
<th>$\tau$ (day)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>0.8659</td>
<td>5.2038</td>
<td>3.0679x$10^{-5}$</td>
<td>9.16</td>
</tr>
<tr>
<td>July</td>
<td>0.9325</td>
<td>5.6559</td>
<td>3.1887x$10^{-5}$</td>
<td>9.56</td>
</tr>
<tr>
<td>Mean</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>9.36</td>
</tr>
</tbody>
</table>

Table 5.5.2. Global mean values of annual mean precipitation (cm/year)

<table>
<thead>
<tr>
<th>Investigator(s)</th>
<th>Precipitation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Budyko (1963)</td>
<td>100.0</td>
</tr>
<tr>
<td>Mira Atlas (1965)</td>
<td>102.0</td>
</tr>
<tr>
<td>Mather (1970)</td>
<td>95.5</td>
</tr>
<tr>
<td>Budyko (1970)</td>
<td>102.0</td>
</tr>
<tr>
<td>Baumgartner and Reichel (1973)</td>
<td>97.3</td>
</tr>
<tr>
<td>*Manabe and Holloway (1975)</td>
<td>104.1</td>
</tr>
<tr>
<td>OSU Model</td>
<td>98.8**</td>
</tr>
</tbody>
</table>

*Model data of Geophysics Fluid Dynamics Laboratory (GFDL), Princeton, New Jersey

**This number is the average of January and July. Actually, the annual mean is 98.0 cm/year from Schlesinger and Gates (1981).
Fig. 5.5.1. Water balance of the earth's surface (cm day$^{-1}$)
Manabe's (1969). The general feature is that the gained water can be distributed over continents, and the water lost over the ocean can be compensated by the runoff from the continent. This figure also illustrates that moisture must be transported from marine to continental atmosphere to accomplish the hydrologic cycle. More details are presented in the next section.

In Table 5.5.3 the earth's surface is classified as oceans or continents in Northern and Southern Hemispheres. The approximate annual mean rates of precipitation and evaporation over each region of the model are tabulated, together with the estimates of these quantities by Manabe and Holloway (1975) and Baumgartner and Reichel (1973).

According to this comparison the present model has a better simulation of precipitation over continent of the Northern Hemisphere than GFDL's model does. On the other hand, evaporation and precipitation are underestimated over the Southern Hemisphere. Nevertheless, the difference between them is fairly close to the observed. This implies that the runoff over the model continents is also close to that over actual continents.

Over the oceanic regions of the model precipitation and evaporation are overestimated over the Northern Hemisphere but underestimated over the Southern Hemisphere. The small excess of evaporation results in the export of water vapor from oceanic to continental regions. On the other hand, continents return the water through runoff. Since the area of oceans is much greater than that of continents, the small E-P over oceans is sufficient to compensate for the large P-E over continents, and then the budget of water vapor in the atmosphere over the entire globe in the long term can be maintained.

As a whole the simulation shows that precipitation is slightly larger than evaporation over the Northern Hemisphere. This indicates that the Southern Hemisphere, mostly covered by ocean, exports moisture to the Northern Hemisphere in the model. The magnitude of this export is smaller than that estimated from the real atmosphere.

Table 5.5.3 reveals that the present model underestimates moisture transport from oceans to continents in the atmosphere. The reason might be related to the fact that evaporation and precipitation over
Table 5.5.3. Annual mean precipitation P and evaporation E over various regions in units of centimeters per year

<table>
<thead>
<tr>
<th></th>
<th>Manabe and Holloway (1975)</th>
<th>Baumgartner and Reichel (1973)</th>
<th>OSU Model</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>P</td>
<td>E</td>
<td>P-E</td>
</tr>
<tr>
<td>Continent</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Globe</td>
<td>97.2</td>
<td>51.1</td>
<td>46.1</td>
</tr>
<tr>
<td>N. H.</td>
<td>91.7</td>
<td>49.8</td>
<td>41.9</td>
</tr>
<tr>
<td>S. H.</td>
<td>110.3</td>
<td>54.5</td>
<td>55.9</td>
</tr>
<tr>
<td>Ocean</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Globe</td>
<td>107.7</td>
<td>126.2</td>
<td>-18.5</td>
</tr>
<tr>
<td>N. H.</td>
<td>117.0</td>
<td>138.4</td>
<td>-21.3</td>
</tr>
<tr>
<td>S. H.</td>
<td>99.2</td>
<td>116.7</td>
<td>-17.6</td>
</tr>
<tr>
<td>Earth</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Globe</td>
<td>104.1</td>
<td>104.1</td>
<td>-</td>
</tr>
<tr>
<td>N. H.</td>
<td>107.0</td>
<td>103.2</td>
<td>3.8</td>
</tr>
<tr>
<td>S. H.</td>
<td>101.3</td>
<td>105.1</td>
<td>-3.8</td>
</tr>
</tbody>
</table>
the Southern Hemisphere are significantly underestimated, particularly over the continents.

Next, zonal mean profile of vertically integrated moisture flux is presented along with the observed one from Rosen et al. (1979a). They designate as Year 1 the period 1 May 1958 - 30 April 1959, and so on for the other years. The flux is computed from surface and upper air radiosonde measurements of specific humidity and wind once a day (0000 GMT) from 10°S to the north pole.

Fig. 5.5.2 shows the meridional profiles of the zonal mean zonal moisture transport. The model flux is integrated vertically and averaged over January and July. The eastward and westward transports are well simulated in the middle and low latitudes, respectively. However, the former transport is overestimated probably due to the coarse resolution in the vertical.

Meridional transport is shown in Fig. 5.5.3. The model values of January and July with their average are plotted along with the observed annual averages. The seasonal reversal of the transport can be seen in low latitudes. The average of January and July (which is quite similar to the mean annual average over the three simulated years though not shown here) does not simulate the southward transport between the equator and 20°N even though the northward transport of the model simulation in the northern middle and high latitudes is in good agreement with observation. The lack of the southward transport is evidently due to the dislocation of the extremal values in the low latitudes.

Geographic distribution of the observed surface evaporation from Budyko (1963) as summarized by Schutz and Gates (1971, 1972) is given in Fig. 5.5.4. According to the comparison of this with the model calculation, as shown in Fig. 5.4.7, most of the distinctive features are well simulated. In January evaporation centers off east coasts of North America and Asia are slightly overestimated. The evaporation centers (≥ 5 mm day⁻¹) over the southern oceans are not well simulated in their location, particularly over the Indian ocean. The one which occurs off the southeast coast of Africa should be noted.

In July the location of intense evaporation (≥ 5 mm day⁻¹) is better simulated in Pacific and Atlantic oceans than in the Indian ocean. There is a spurious area off the southeast coast of Africa.
Fig. 5.5.2. Meridional profiles of the zonal mean moisture transport \[\bar{Q}_\lambda\] (10 kg s\(^{-1}\) m\(^{-1}\)).

\[
\bar{Q}_\lambda = \frac{1}{g} \int_{p_T}^{p_S} \bar{u}q \, dp
\]

where "bar" denotes annual mean.
Fig. 5.5.3. Meridional profiles of the total moisture flow $2\pi a \cos \phi [\bar{Q}_\phi]$ ($10^8$ kg s$^{-1}$)
across latitude walls in the atmosphere where $a$ is the radius of the earth,
$\phi$ is latitude and $\bar{Q}_\phi = \frac{1}{g} \int_{P_T}^{P_S} \bar{v}q \, dp$ is the annual mean of $Q_\phi$. 
Fig. 5.5.4. The observed surface evaporation rate (mm day$^{-1}$) for January and July (continued). These are from Budyko (1963) as summarized by Schutz and Gates (1971, 1972b).
Fig. 5.5.4. Continued
The causes of model deficiencies are not clear yet. However, part of the overestimation off the east coast of the winter hemisphere continents as seen in Schlesinger and Gates (1980) has been slightly reduced in composite mean with some revisions described in Schlesinger and Gates (1981).

The geographic distribution of precipitation is shown in Figs. 5.5.5 and 5.5.6 for the observed and simulated respectively. In January the rainbelts in the tropics are well simulated. The overestimated precipitation can be visualized off the east coasts of continents in the Northern Hemisphere. However, the accuracy of the observed data is also in doubt. There are several spurious regions of heavy precipitation in the model, for example, off the south coast of Africa, in the Arabian Sea and off the east coast of Africa.

In July the rainbelts in the tropics seem to be well simulated. The areas of precipitation heavier than five millimeters per day, off the south and east coast of Africa and in the northern middle latitudes, are not observed in the real world.

To understand the reason for these discrepancies, more detailed information should be extracted from the simulated data. Since there are three processes to cause precipitation, their contribution can be easily identified. Figs. 5.5.7 and 5.5.8 show the middle-level and penetrating convective precipitation, while large-scale condensation is shown in Fig. 5.4.8.

In January penetrating convection is responsible for all overestimated precipitation. It should be pointed out that this intense subgrid-scale activity is highly connected to the strong surface evaporation shown in Fig. 5.4.7. This model behavior is unrealistic probably due to the improper parameterization of the PBL processes. On the other hand, middle-level convection is mainly responsible for the successful part of the simulation. This is in contradiction to our observation because middle-level convection like altocumulus rarely causes heavy precipitation.

In July, penetrating convection produces most of the sporadic precipitation to the east and south of Africa and over North America. Middle-level convection contributes to heavy precipitation in the northern middle latitudes. The spurious precipitation near South Africa
Fig. 5.5.5. The observed precipitation rate (mm day⁻¹) from Müller (1951) as summarized by Schutz and Gates (1972a,b) for December, January and February; and for June, July and August (continued).
Simulated total precipitation rate (mm day$^{-1}$) of composite January and July (continued). In January dashed lines are for 6; solid lines at contour interval of 2 are for higher values; dotted lines at contour interval of 1.5 are for lower values. In July dashed lines are for 5; solid lines at contour interval of 5 are for higher values; dotted lines at contour interval of 2 are for lower values.
Fig. 5.5.6. Continued
Fig. 5.5.7. Total penetrating convective precipitation rate (mm day$^{-1}$) of January and July (continued). In January dashed lines are for 6; solid and dotted lines at contour interval of 2 are for higher and lower values, respectively. In July dashed lines are for 5; solid lines at contour interval of 5 are for higher values; dotted lines at contour interval of 2 are for lower values.
Fig. 5.5.8. Total middle-level convective precipitation rate (mm day\(^{-1}\)) of January and July (continued). Dashed lines are 2 and 4 for January and July, respectively; solid and dotted lines at contour interval of 2 and 1 are for higher and lower values, respectively.
is also related to large-scale condensation due to fictitious cyclone activity. The situation is more complex in July than in January.

What kind of action should be taken to improve the model is not a simple problem. A recommendation will be given in the next chapter. Diagnostic analysis does facilitate our understanding of not only the model behavior but also the control mechanisms in the real atmosphere.

5.6 A global view of the hydrologic cycle.

From the model verification the global simulation is more realistic than the geographic distribution. It is adequate to bring out the achievement through a quantitative estimate of the hydrologic cycle on the global scale. In Table 5.6.1 is the model area of continents and oceans over each hemisphere. The total area is slightly smaller than that of a sphere with radius $6.375 \times 10^6$ m. This is due to the approximation of finite difference grid network.

Figs. 5.6.1 and 5.6.2 are the global hydrologic cycle for January and July, respectively. Since all physical quantities are multiplied by the area over each grid, the units become $10^6$ kg s$^{-1}$. The arrow denotes the direction of moisture transport. MF, S, E, T, PC, MLC and LSC represent moisture flux, standing, transient, total, penetrating convection, middle-level convection and large-scale condensation, respectively.

In January horizontal moisture transport in the atmosphere is essentially from ocean to continent, except the standing mode in the upper layer. Vertical transport shows that transient mode is dominant over the ocean but standing mode is dominant over the continent. This indicates that transient disturbance due to short scale baroclinic processes is more effective to induce net upward transfer than semi-permanent circulation due to planetary scale forcing over the ocean. However, the opposite situation occurs over the continent due to the prominent orographic effect. Large-scale condensation (Fig. 5.5.8) is favored to occur in the winter (northern) hemisphere where continent has a relatively large size. However, penetrating convection (Fig. 5.5.7) is weak due to weak surface evaporation (Fig. 5.4.7) over nor-
Table 5.6.1. Model area \((10^{14} \text{ m}^2)\)

<table>
<thead>
<tr>
<th>Continents</th>
<th>Northern Hemisphere</th>
<th>Southern Hemisphere</th>
<th>Globe</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1.019</td>
<td>0.475</td>
<td>1.494</td>
</tr>
<tr>
<td>Oceans</td>
<td>1.532</td>
<td>2.076</td>
<td>3.608</td>
</tr>
<tr>
<td>Total</td>
<td>2.551</td>
<td>2.551</td>
<td>5.102</td>
</tr>
</tbody>
</table>

Table 5.6.2. Runoff \((10^{-5} \text{ kg s}^{-1} \text{ m}^{-2})\)

<table>
<thead>
<tr>
<th>Month</th>
<th>Northern Hemisphere</th>
<th>Southern Hemisphere</th>
<th>Globe</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>0.4720</td>
<td>1.1226</td>
<td>0.6788</td>
</tr>
<tr>
<td>July</td>
<td>0.7372</td>
<td>0.2014</td>
<td>0.5669</td>
</tr>
</tbody>
</table>

Table 5.6.3. Simulated hydroelectric power

\(10^{12} \text{ watts}\) \hspace{2cm} \text{Mean and variability}

<table>
<thead>
<tr>
<th>Month</th>
<th>Mean and variability</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>5.1488 ± 2.5726</td>
</tr>
<tr>
<td>July</td>
<td>5.5925 ± 2.3940</td>
</tr>
</tbody>
</table>
Fig. 5.6.1. Global hydrologic cycle of January $10^6$ kg s$^{-1}$). MFS, MFE and MFT stand for standing, transient and total moisture flux, respectively. PC, MLC and LSC represent penetrating convection, middle-level convection and large-scale condensation, respectively.
Fig. 5.6.2. Global hydrologic cycle of July ($10^6$ kg s$^{-1}$). MFS, MFE and MFT stand for standing, transient and total moisture flux, respectively; PC, MLC and LSC represent penetrating convection, middle-level convection and large-scale condensation, respectively.
thern continents. Therefore, middle-level convection can play the second role to cause precipitation over land. Since strong surface evaporation takes place over oceans as shown in Fig. 5.4.7, penetrating convection can trigger a large amount of precipitation (Fig. 5.5.7). Since part of the intense middle-level convective precipitation occurs over continents, e.g., South America and Africa (Fig. 5.5.8), large-scale condensation (Fig. 5.4.8) gives slightly more contribution.

In July, moisture transport has a similar structure except that the standing mode in the upper layer is larger than the transient mode. Thus the total horizontal transport in the upper layer is from continent to ocean. Over continents evaporation is relatively large over relatively wide northern lands as shown in Fig. 5.4.7, and penetrating convection becomes the first important mechanism to produce precipitation (Fig. 5.5.7). Large-scale condensation is relatively intense over relatively narrow lands in the Southern Hemisphere as shown in Fig. 5.4.8, and thus it gives the least contribution to precipitation over continents. Over oceans penetrating convection is highly connected to surface evaporation according to Figs. 5.5.7 and 5.4.7. This process causes most of the precipitation. Since large-scale condensation is dominant over the Southern Hemisphere, as shown in Fig. 5.4.8 where the ocean is relatively wide, it induces slightly less precipitation than middle-level convection which causes intense precipitation over the northern oceans (see Fig. 5.5.8).

In general evaporation and precipitation over the oceans are much larger than those over the continents. In January the oceanic surplus \((1.708 \times 10^9 \text{ kg s}^{-1})\) of evaporation over precipitation \((E-P)\) is larger than the magnitude of the continental deficit \((1.505 \times 10^9 \text{ kg s}^{-1})\) of \(E-P\). However, in July the oceanic surplus \((4.93 \times 10^8 \text{ kg s}^{-1})\) is smaller than the continental deficit \((7.93 \times 10^8 \text{ kg s}^{-1})\). This information reveals that marine atmosphere is moistened in January but is dried in July. Continental air is always slightly dried in both January and July. As a consequence of seasonal variation of the continental deficit, the runoff from continents to oceans is smaller in July than in January. According to the comparison between continental deficit and runoff, snow mass is built in January but is melted in
July. Therefore, a large amount of oceanic surplus in January not only moistens the marine air but also moves the snowline southward over northern continents. Since there is no budget equation for ice in the present model, snow mass is responsible to take care of the difference between runoff and continental deficit.

As shown in Table 5.6.2, runoff is larger in the summer hemisphere than in the winter hemisphere. It is relatively intense in January over the globe. The geographic distribution of runoff is shown in Fig. 5.6.3. In January relatively intense runoff (\(\geq 3 \text{ mm day}^{-1}\)) occurs in the west side of the Rocky Mountains and over the Amazon River basin. The areas of secondary intense runoff (\(\geq 1 \text{ mm day}^{-1}\)) are distributed in central North America, southern part of South America, Africa and Europe, and Central Eurasia.

In July, the most intense runoff (\(\geq 3 \text{ mm day}^{-1}\)) takes place in the east of Greenland, in the south of South America and in central Africa. The areas with runoff larger than 1 mm day\(^{-1}\) are located in the east and the west of North America, in the north of South America, in the edge of the Asian and African continents, and in the west side of the Tibetan plateau. Further detailed information about the calculation of runoff is given in Appendix C.

Another interesting application of this quantity is to estimate the hydroelectric power which is the product of runoff and surface geopotential. Since the runoff is the discharge rate of the available water in the river, the change of surface elevation can cause potential energy release while flowing downhill. Tables 5.6.3 shows that the total available hydroelectric power is less in January than in July. However, the interannual variability in July is smaller than in January. The former is about 43% of the mean but the latter is about 50%.
Fig. 5.6.3. Geographical distribution of runoff rate (mm day$^{-1}$) of January and July (continued). Dashed lines are for 3; solid and dotted lines at contour interval of 2 are for higher and lower values respectively.
Fig. 5.6.3. Continued
6. CONCLUSIONS AND RECOMMENDATIONS

In this chapter all findings of the present study are summarized. The hydrologic performance of the AGCM is now better understood than ever before even if the present study did not include the moisture balance of the underlying ground and snow mass. Some conclusions on the problems of the present study as laid out in the Introduction will be made, and some suggestions for improving the present hydrologic cycle will also be made in an ordered set of recommendations.

6.1 Conclusions

Strong surface evaporation occurs over the oceans between 40°S and 40°N. The standing part of the surface evaporation is dominant over the negative transient part which is essentially due to the positive correlation between the turbulence intensity and surface air humidity. Moreover, the horizontal structure of the standing part conforms to that of the standing vapor pressure difference between the air and the underlying surface.

In January, moisture sources in the lower layer are located to the west of the southern continents, i.e., South America, Africa and Australia but three sinks are found between 50°E and 100°E. In July, the Australian source has been shifted to the east of Madagascar and sinks occur in the Pacific Ocean to the southeast of Asia. The sources and sinks in the upper layer are located essentially where the lower-layer sinks and sources occur, respectively. The global distribution of sources and its characteristics remain unchanged when the transient component of motion is ignored in calculating the transport; this means that the divergent transport is locally dominated by the standing component of motion in general. Walker circulations found along 15°S in January and also along the equator in July represent the seasonal drift of the total mass circulation converging in the lower layer near the tropical western Pacific Ocean. These findings are based on the fields of divergent component of the horizontal transport.
The divergent transport by transient component of motion is locally small as mentioned earlier, but its effect on the meridional transport is not negligible. In general, it is a remarkably and understandably down-gradient transport in both layers; moisture is thus carried by transient motion from the oceans to continents and from the low to high latitudes. Up-gradient transport occurs, however, in some areas and further analyses and experiments may be necessary for its explanation.

Large-scale vertical transport moistens the upper layer; the standing and transient motions contribute in the tropics and higher latitudes, respectively. The downward transport by large-scale condensation in the upper layer and subsequent evaporation of the precipitating water in the lower layer is associated with the cyclones which bring in moisture to cause supersaturation; therefore it occurs more frequently in the middle latitudes than elsewhere and is more intense in the winter hemisphere than in the summer hemisphere. This transport occurs also in the tropics but it does not lead to precipitation because the tropical lower layer is too warm to be supersaturated.

Most convective precipitation by penetrating cumuli occurs in regions of strong surface evaporation, and some occurs in zones of low-level moisture convergence where convective precipitation by mid-level cumuli is relatively intense.

The characteristics of the global balance of moisture include the following features:

(1) The upper layer is moistened exclusively by large-scale vertical transport against precipitation whereas the lower layer is moistened by both surface evaporation and internal evaporation of free-falling water drops from the large-scale condensation in the upper layer;

(2) The largest contribution to total precipitation is from penetrating convection in both January and in July;

(3) The atmosphere is slightly moistened in January and similarly dried in July.

The characteristics of the hemispheric balance of moisture include the following features:
(1) Surface evaporation is stronger in winter than summer hemisphere;
(2) Large-scale condensation favors the winter hemisphere;
(3) Horizontal transport is from winter to summer hemisphere;
(4) Large-scale vertical transport is stronger in summer than winter hemisphere;
(5) Mid-level convection and the upper-layer effect of penetrating convection are more intense in summer than winter hemisphere;
(6) The lower-layer effect of penetrating convection is stronger in the Northern than Southern Hemisphere.

Interannual variabilities of the lower-layer effects of horizontal moisture convergence and penetrating convection are dominant over those of other processes. The vertically integrated moisture balance over each hemisphere shows that horizontal moisture convergence and middle-level convection are primarily responsible for the seasonal variation.

The simulated residence time of atmospheric moisture is 9.36 days and the simulated global average of annual mean precipitation is 0.988 m/year. (Hereafter, the annual mean refers to the average of January and July when it is with regard to the model.) Both of these estimates are in excellent agreement with those estimated from the observed data. This merely indicates that the global statistics of the AGCM are not significantly different from those of the atmosphere and that the surface water balance is also in close agreement with observation.

Further comparison, however, shows significant discrepancy as well as agreement. Annual mean evaporation and precipitation are overestimated in the northern oceans but underestimated in the Southern Hemisphere; detailed diagnosis reveals that in January penetrating convection is mainly responsible for this overestimation and that all the precipitation processes are involved in producing unrealistic rain in July.

For both January and July, the zonal-mean zonal transport is overestimated in the middle latitudes, and the zonal-mean meridional transport is not well simulated in the low latitudes. The poleward transport in the northern middle and high latitudes is in good agreement with observations. The maximum toward-thermal-equator transport in
the tropics occurs at the geographic equator for both January and July, indicating that these maxima are about 5 degrees of latitude closer to the seasonal thermal equator than the observed maxima. The approximate annual mean does not show the equatorward transport between the equator and 20°N.

The maritime atmosphere is moistened by the excess of evaporation over precipitation in January but not moistened enough by the relatively small excess of evaporation over precipitation in July. The continental atmosphere is dried, however, in both January and July; the continental deficits are of course compensated by the moistening effect of the horizontal transport even though a relatively small transport occurs from the continents to the oceans by the standing motion in the upper layer. As a consequence of the seasonal variation of the maritime moisture budget, the runoff is smaller in July than in January, implying that snow mass builds up in January but melts down in July. Thus the oceanic surplus in January not only moistens the maritime atmosphere but also moves the snowline southward over the northern continents. The penetrating cumuli are a major mechanism of maritime precipitation whereas the large-scale condensation and penetrating cumuli have the dominating effect on the continental precipitation during January and July, respectively; the seasonal precipitation over the northern hemispheric continent concurs with strong surface evaporation in summer and also with strong cyclonic activity in winter.

6.2 Recommendations

The present diagnostic analysis reveals the basic nature of the model behavior. According to the verifications, many observed geographic features of the hydrologic processes are well simulated. Nevertheless, sensitivity studies may be indispensible in order to remove model deficiencies.

The unrealistic rainy zones are primarily due to the penetrating convection which is then connected to the surface evaporation. In the present PBL parameterization the surface wind speed is linearly extrapolated from the prognostic winds at the upper and lower levels and the drag coefficient does not depend on the static stability
of a constant-flux surface layer. These could be partly related to the unrealistic surface evaporation over the Indian Ocean. Another important action may be taken in relation to the spurious precipitation off the southern coast of Africa. According to the suggestion by Lahey (personal communication, 1980) a subgrid-scale horizontal momentum flux may induce a local orographic effect large enough to interrupt the organized cyclonic activity which pumped moisture into this region. Detailed analysis by Manabe and Holloway (1975) seems to support this suggestion.

On the other hand, some of the successful simulations are related to unrealistically heavy precipitation due to mid-level convection which is supported by the low-layer convergence. This might be due to the algorithm of the AGCM which is certainly biased toward mid-level convection; in reality this convection seems to occur with infrequent precipitation. The coarse vertical resolution should also share part of the difficulty.

The overestimated annual mean meridional profile of the zonal transport in the middle latitudes could be related also to the coarse vertical resolution of the AGCM. The unrealistic meridional transport in the low latitudes may need detailed analysis of motion fields for explanation.

Based upon the above arguments a strategy for model improvement is proposed as follows in the designated order:

(1) Detailed analysis of momentum, heat and mass field must be documented for the present AGCM;

(2) Non-slip boundary condition for wind may be imposed on the model bottom to obtain the surface wind speed;

(3) The PBL parameterization designed by Deardorff (1972) may be used for calculating the surface fluxes;

(4) The cumulus parameterization may include an additional criterion proposed by Helfand (1981) which inhibits convection when relative humidity is less than 95% near the top of the PBL;

(5) The non-linear formulation by Smagorinsky (1963) of subgrid-scale horizontal momentum flux may be adopted.
The suggested order of the recommendation is based on the degree of their complexity and importance for the present model. Improvements in the simulated hydrologic cycle and also in the simulated climate should be expected, and practical application of the model would then be more feasible.


Starr, V.P., J.P. Peixoto and N.E. Gaut, 1970: Momentum and zonal kinetic energy balance of the atmosphere from five years of hemispheric data. Tellus, 22, 251-274.

APPENDICES
This appendix is based on Schlesinger (1979, personal communication). Taking the ensemble average for the first law of thermodynamics, we can obtain

$$\frac{\partial}{\partial t} \langle \rho T \rangle = -\nabla \cdot (\bar{\rho} \bar{V} \bar{T}) - \frac{\partial}{\partial \sigma} \langle \bar{\rho} \bar{V}_\sigma \bar{T} \rangle + \frac{\bar{h}}{c_p} \bar{a} + \frac{\bar{g}}{c_p} \frac{\partial R}{\partial \sigma}$$

$$+ \frac{\bar{h}}{c_p} L\bar{C} - \frac{1}{c_p} \nabla \cdot \bar{V} \langle \bar{\rho} \bar{V}^* s^* \rangle - \frac{1}{c_p} \frac{\partial}{\partial \sigma} \langle \bar{\rho} \bar{V}^* s^* \rangle$$

where $s$ is dry static energy. When we derive this equation, we have already used the result of scale analysis for cyclone motion, i.e.,

$$\omega = -\rho gw$$

where $\omega$ and $w$ are vertical velocities in $p$- and $z$-coordinates, respectively. Therefore, the net thermal effect of the cloud ensemble on the environmental air can be written as

$$\left[ \frac{\partial}{\partial t} \langle \pi T \rangle \right]_{\text{clouds}} = \frac{1}{c_p} \nabla \cdot \bar{V} \langle \bar{\rho} \bar{V}^* s^* \rangle - \frac{1}{c_p} \frac{\partial}{\partial \phi} \langle \bar{\rho} \bar{V}^* s^* \rangle$$

In the following the entrained mass flux will be determined for middle-level and penetrating convection, respectively.

(1) Middle-level convection

For the upper layer the net thermodynamic effect can be parameterized as

$$\left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} T \right) \right]_{\text{MLC}} = \frac{\eta M}{c_p} (s_{c,1} - s_2)$$
Since
\[ s_{c,1} - s_2 = c_p (T_{c,1} - T_1) + s_1 - s_2 \]

and
\[ h_{c,1} - h_1^* = c_p (T_{c,1} - T_1) + L(q_{c,1} - q_1^*) \]

\[ \approx c_p \left[ 1 + \frac{L}{c_p} \frac{\partial q_s}{\partial T} (T_{1,p_1}) \right] (T_{c,1} - T_1) , \]

then
\[
\left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} T_1 \right) \right]_{MLC} \approx \frac{nM}{c_p} \left[ \frac{h_{c,1} - h_1}{1 + \gamma_1} + (s_1 - s_2) \right] ,
\]

(A.5)

where the definition for \( s \), \( h \) and \( \gamma \) can be recalled from (2.4.5) and (2.4.9). In this case \( h_{c,1} = h_3 \) which can be easily seen in schematic diagram in section 2.4.

For the lower layer
\[
\left[ \frac{\partial}{\partial t} \left( \frac{\pi \Delta \sigma}{g} T_3 \right) \right]_{MLC} = \frac{nM}{c_p} (s_2 - s_3) .
\]

(A.6)

To determine the mass flux \( nM \), we assume that
\[
\left\{ \frac{\partial}{\partial t} [\pi (h_3 - h_1^*)] \right\}_{MLC} = \frac{\pi (h_3 - h_1^*)}{\tau}
\]

(A.7)

This means that middle-level convective instability can be reduced to \( e^{-1} \) of the initial intensity within time period \( \tau \), which is one hour. Also,
\[
\left\{ \frac{\partial}{\partial t} [\pi (h_3 - h_1^*)] \right\}_{MLC} \approx c_p \left[ \frac{\partial (\pi T_3)}{\partial t} \right]_{MLC} + L \left[ \frac{\partial (\pi q_3)}{\partial t} \right]_{MLC}
\]
\[ - c_p (1 + \gamma_1) \left[ \frac{\partial (\pi T_1)}{\partial t} \right]_{MLC} \]
\[ nM \frac{g}{\Delta \sigma} \left[ \left( s_2 - s_3 \right) + L \left( q_2 - q_3 \right) \right] - (1 + \gamma_1) \left[ \frac{h_3 - h_1^*}{1 + \gamma_1} + \left( s_1 - s_2 \right) \right] \]

\[ nM \frac{g}{\Delta \sigma} \left[ 2s_2 - s_1 - s_3 \right] + L (q_2 - q_3) - \left( h_3 - h_1^* \right) - \gamma_1 (s_1 - s_2) \]  
(A.8)

From (A.7) and (A.8) we can obtain

\[ nM \frac{g}{\Delta \sigma} \left[ \frac{h_3 - h_1^*}{(h_3 - h_1^*) + L(q_3 - q_2) + \gamma_1 (s_1 - s_2) + s_1 + s_3 - 2s_2} \right] \]  
(2.4.4)

(2) Penetrating convection

The net thermodynamic effect for the upper layer can be parameterized as

\[ \frac{\partial}{\partial t} \left[ \left( \pi T_1 \right) \right]_{PC} = M \frac{g}{\Delta \sigma} \frac{1}{c_p} \left( s_{c,1} - s_2 \right) \]  
(A.9)

or similarly as middle-level convection,

\[ \frac{\partial}{\partial t} \left[ \left( \pi T_1 \right) \right]_{PC} = M \frac{g}{\Delta \sigma} \frac{1}{c_p} \left[ \frac{1}{1 + \gamma_1} \left( h_4 - h_1^* \right) + s_1 - s_2 \right] \]  
(A.10)

It should be noted that \( h_c = h_4 \) shown in (2.4.9). The mass flux is purely from planetary boundary layer so that \( \eta = 1 \). Actually, \( \eta \) is defined as the ratio of cloud entrainment mass flux to the mass flux from PBL. For middle-level convection \( \eta \rightarrow \infty \) and \( M \rightarrow 0 \) such that \( nM \) is finite. Since moist static energy is conserved during moist convective processes, \( h_c = h_4 \) for penetrating convection but \( h_c = h_3 \) for middle-level convection. The former does not have mass entrained from lateral boundary of cloud but the latter has negligible mass entrained from the bottom boundary. After this digression the thermal effect on the lower
layer is

$$\left[ \frac{\partial}{\partial t} (T_3) \right]_{PC} = M \frac{g}{\Delta \sigma} \frac{1}{c} (s_2 - s_4) \quad (A.11)$$

Assume that

$$\frac{\partial}{\partial t} [\pi(h_4 - h_3*)]_{PC} = -\frac{\pi(h_4 - h_3*)}{\tau} \quad (A.12)$$

Since

$$\frac{\partial}{\partial t} [\pi(h_4 - h_3*)]_{PC} \equiv L \left( \frac{\partial q_4}{\partial q_3} \right) \left[ \frac{\partial (\pi q_3)}{\partial t} \right]_{PC} + c_p \left( \frac{\partial T_4}{\partial T_3} \right) \left[ \frac{\partial (\pi T_3)}{\partial t} \right]_{PC}$$

$$- c_p \left[ \frac{\partial (\pi T_3)}{\partial T} \right]_{PC} - L \left[ \frac{\partial q_3}{\partial T} \right]_{PC}$$

$$= L \left( \frac{\partial q_4}{\partial q_3} \right) M \frac{g}{\Delta \sigma} (q_2 - q_4) - c_p \left[ 1 + \gamma_3 - \left( \frac{\partial T_4}{\partial T_3} \right) \right] M \frac{g}{\Delta \sigma} \frac{1}{c} (s_2 - s_4)$$

then

$$M = \frac{1}{\tau} \frac{\pi \Delta \sigma}{\rho g} \left\{ \frac{h_4 - h_3*}{\left[ 1 + \gamma_3 \left( \frac{\partial T_4}{\partial T_3} \right) \right] (s_2 - s_4) + L \left( \frac{\partial q_4}{\partial q_3} \right) (q_4 - q_2)} \right\} \quad (2.4.9)$$

To determine \( \frac{\partial T_4}{\partial T_3} \) we have to know the functional relationship between \( T_4 \) and \( T_3 \). Following Schlesinger and Gates (1979) and Arakawa (1972), we have the PBL parameterization for the sensible heat flux \( H_s \) in the form

$$H_s = \rho_s c \frac{C_p V_s}{\Delta s} \left( T - T_4 \right)$$

$$= -\rho_s c K \left[ \left( \frac{T_3}{Z_3} - \frac{T_4}{Z_4} \right) - \left( \frac{T_3}{Z_3} - \frac{T_4}{Z_4} \right) \right] \quad (A.13)$$
where $T_g$ is the ground or sea surface temperature and $Z_3$ is the height at level 3. Following Arakawa (1972),

$$\left(\frac{T_3 - T_4}{Z_3 - Z_s}\right)_{\text{crit}} = RH_4 \left(\frac{T_3 - T_4}{Z_3 - Z_s}\right)_{\text{ma}} + (1 - RH_4) \left(\frac{T_3 - T_4}{Z_3 - Z_s}\right)_{\text{da}} \quad (A.14)$$

where "ma" and "da" express moist and dry adiabatic process, respectively. Since $T_3$ is the temperature at level 3 determined prognostically and $T_4$ is adiabatically extrapolated, (A.14) can be written as

$$\left(T_4\right)_{\text{crit}} = \left(T_4\right)_{\text{da}} + RH_4 \left[\left(T_4\right)_{\text{ma}} - \left(T_4\right)_{\text{da}}\right] \quad (A.15)$$

For dry adiabatic process the potential temperature

$$\Theta = T \left(\frac{p_0}{p}\right)^\kappa \quad (A.16)$$

is conserved where $\kappa = \frac{R}{c_p}$. We can obtain

$$\frac{\partial T}{\partial Z}_{\text{da}} = \frac{g}{c_p}. \quad (A.17)$$

For moist adiabatic process the equivalent potential temperature

$$\Theta_e \equiv \Theta \exp \left(\frac{L q_s}{c_p T}\right) \quad (A.18)$$

is conserved. We can obtain

$$\frac{\partial T}{\partial Z}_{\text{ma}} = \frac{g}{c_p} \left(1 + \gamma - \frac{L q_s}{c_p T}\right)^{-1} \quad (A.19)$$

From (A.17) and (A.19), we can obtain
\[(T_4)_{ma} - (T_4)_{da} = (Z_3 - Z_s) \left[ (-\frac{\partial T}{\partial Z})_{ma} - (-\frac{\partial T}{\partial Z})_{da} \right] \]
\[= -(Z_3 - Z_s) \frac{g}{c_p} (\gamma_4 - \frac{Lq_{s4}}{c_p T_4}) \frac{Lq_{s4}}{1 + \gamma_4 \frac{Lq_{s4}}{c_p T_4}} \]  
\[(A.20)\]

Also,
\[(T_4)_{da} = T_3 \left( \frac{p_4}{p_3} \right)^\kappa \]

and \(Lq_{s4} \ll c_p T_4\). Hence
\[(T_4)_{\text{crit}} = T_3 \left( \frac{p_4}{p_3} \right)^\kappa - \frac{g(Z_3 - Z_s) RH_4 \kappa (\gamma_4 - \frac{Lq_{s4}}{c_p T_4})}{R(1 + \gamma_4)} \]
\[(A.21)\]

From (A.13)
\[T_4 = \left( \frac{K}{Z_3 - Z_s + c_p V} \right)^{-1} \left[ c_p V T g + \frac{K}{Z_3 - Z_s} (T_4)_{\text{crit}} \right] \]
\[(A.22)\]

After combining (A.21) and (A.22) the relationship between \(T_4\) and \(T_3\) can be established and gives
\[\left( \frac{\partial T_4}{\partial T_3} \right) = \left( \frac{K}{Z_3 - Z_s + c_p V} \right)^{-1} \frac{K}{Z_3 - Z_s} \left( \frac{p_4}{p_3} \right)^\kappa \]
\[(A.23)\]

Where \(T_4\)-dependent term in (A.21) are evaluated at the \(T_4\) of the previous time step. From (2.4.24) we can find
\[\frac{\partial q_4}{\partial q_3} = \left( \frac{K}{Z_3 - Z_s + c_p V} \right)^{-1} \frac{K}{Z_3 - Z_s} \]
\[(A.24)\]
APPENDIX B

NUMERICAL SOLUTION OF POISSON EQUATION

B.1 Finite difference analogue

Since the present model has variables distributed in B-type staggered grid system as shown in Fig. B.1.1, we put scalar potential \( \chi \) on \( \pi \)-grid point but stream function \( \psi \) on uv-grid point. Also moisture flux vector \( F \) is oriented as C-type staggered grid system in terms of \( \pi \)-grid point. Therefore the finite difference form of (3.1.1) is

\[
F^\lambda(i,j) = -\frac{\psi(i,j+1) - \psi(i,j)}{\Delta y^\pi(j)} + \frac{\chi(i+1,j) - \chi(i,j)}{\Delta x^\pi(j)}
\]

(B.1.1)

\[
F^\phi(i,j) = \frac{\psi(i,j) - \psi(i-1,j)}{\Delta x^u(j)} + \frac{\chi(i,j) - \chi(i,j-1)}{\Delta y^u(j)}
\]

where superscripts \( \pi \) and \( u \) represent \( \pi \)-grid and uv-grid point respectively. For simplicity the vertical level index has been omitted. All other corresponding expressions can be found from (3.1.1) and Fig. B.1.1. For example, the longitudinal and latitudinal indices are \( i \) and \( j \), respectively.

With the aid of Stokes' theorem, integrating (3.1.2) over the area around the uv-grid point \((i,j)\) gives

\[
\iint \nabla^2 \psi \, ds \bigg|_{i,j} = \oint F \cdot d\ell \bigg|_{i,j}
\]

(B.1.2)

where the left hand side is area integral and the right hand side is line integral. Since the latter can be directly evaluated from Fig. B.1.1, the substitution of the finite difference form of moisture flux yields a discretized formulation for the Poisson equation. That is,
Fig. B.1.1. Distribution of variables over staggered grid system. See text for symbols.
\[ \Delta x^p(i-1) \Delta y^p(j-1) \psi(i,j-1) + \Delta y^u(j) \Delta x^u(j) \psi(i+1,j) + \Delta x^p(j) \Delta y^p(j) \psi(i,j+1) + \Delta y^u(j) \Delta x^u(j) \psi(i,j-1) + \Delta x^p(j-1) \Delta y^p(j-1) \psi(i,j) \]

\[ - \frac{\Delta x^p(j-1)}{\Delta y^p(j-1)} - \frac{\Delta y^u(j)}{\Delta x^u(j)} - \frac{\Delta x^p(j)}{\Delta y^p(j)} - \frac{\Delta y^u(j)}{\Delta x^u(j)} \psi(i,j) \]

\[ = F^\lambda(i,j-1) \Delta x^p(j-1) + F^\phi(i+1,j) \Delta y^u(j) - F^\lambda(i,j) \Delta x^p(j) - F^\phi(i,j) \Delta y^u(j) \]

for 2 < i < 73 and 2 < j < 46.

At the polar points the longitudinal spacing between two adjacent π-grid points vanishes. It should be noted that j = 1 for π-grid is at the south pole.

Similarly, Gauss theorem helps to build the finite difference analogue for the Poisson equation of scalar potential as follows:

\[ \frac{\Delta x^u(j)}{\Delta y^u(j)} \chi(i,j-1) + \frac{\Delta y^p(j)}{\Delta x^p(j)} \chi(i+1,j) + \frac{\Delta x^u(j+1)}{\Delta y^u(j+1)} \chi(i,j+1) + \frac{\Delta y^p(j)}{\Delta x^p(j)} \chi(i-1,j) + \frac{\Delta x^u(j)}{\Delta y^u(j)} \chi(i,j) \]

\[ - \frac{\Delta x^u(j)}{\Delta y^u(j)} - \frac{\Delta y^p(j)}{\Delta x^p(j)} - \frac{\Delta x^u(j+1)}{\Delta y^u(j+1)} - \frac{\Delta y^p(j)}{\Delta x^p(j)} \chi(i,j) \]

\[ = F^\lambda(i,j) \Delta y^p(j) + F^\phi(i,j+1) \Delta x^u(j+1) - F^\lambda(i,j) \Delta x^u(j) - F^\phi(i,j+1) \Delta y^p(j) \]

for 2 < i < 73 and 2 < j < 46.

(B.1.3)
B.2 Boundary conditions

To have a unique and stable solution for the elliptic equation the value or the slope must be specified on a closed boundary (Jackson, 1969). In the present situation we require that the zonal component of moisture flux at the poles should vanish. From (3.1.1) we can obtain

\[ F^x(\lambda, \phi, \sigma) = \frac{\partial \psi}{\partial y} + \frac{\partial \chi}{\partial x} = 0 \quad \text{at} \quad \phi = \pm 90^\circ \]  

This implies that

\[ \frac{\partial \psi}{\partial y} = 0 \]  

(B.2.2)

and

\[ \frac{\partial \chi}{\partial x} = 0 \]  

(B.2.3)

Since there is only one point at each pole, the scalar potential must be constant at the poles. For convenience we can choose zero. Therefore this kind of elliptic equation forms a Dirichlet problem. However, the stream function has Neumann boundary conditions (B.2.2). We shall need to specify the value of \( \psi \) at some single point to obtain a unique answer.

B.3 Solution methods

There are many methods to solve a Poisson equation. For example Smith (1965) presents Jacobi's, Gauss-Seidel, successive overrelaxation, and alternating direction implicit schemes. In the present study Lindzen and Kuo's method (1969) is used. To facilitate the requirement of this scheme we have to expand (B.1.3) and (B.1.4) into tri-diagonal form. Let

\[ \psi(\xi, j) = \text{Re} \left\{ \sum_{m=1}^{72} c(m, j)e^{i\theta} \right\} \]  

(B.3.1)
where
\[ c(m, j) = a(m, j) - i b(m, j) \]

and
\[ \theta = (m - 1)(\ell - 2) \left( \frac{\pi}{36} \right). \]

It should be noted that \( \ell \) is longitudinal index and \( 2 \leq \ell \leq 73 \). Also Fourier transform of (B.3.1) is
\[
\mathcal{F}(\omega, j) = \sum_{\ell=2}^{73} \psi(\ell, j) e^{-i \theta}
\]

Substitution of (B.3.1) into (B.1.3) and operation by Fourier transform eventually yields
\[
\alpha(j)A(m, j-1) + \gamma(j)A(m, j+1)
\]
\[
- (\alpha(j) + (\beta(j) + \delta(j))[1 - \cos \left( \frac{\pi}{36}(m-1) \right)] + \gamma(j)) A(m, j)
\]
\[
= \sum_{i=2}^{36} \mathcal{F}(\omega, j) \cos \left( \frac{\pi}{36}(m-1)(\ell-2) \right)
\]
for \( 1 \leq m \leq 37 \)

\[
\alpha(j)B(m, j-1) + \gamma(j)B(m, j+1)
\]
\[
- (\alpha(j) + (\beta(j) + \delta(j))[1 - \cos \left( \frac{\pi}{36}(m-1) \right)] + \gamma(j)) B(m, j)
\]
\[
= \sum_{i=2}^{36} \mathcal{F}(\omega, j) \sin \left( \frac{\pi}{36}(m-1)(\ell-2) \right)
\]
for \( 2 \leq m \leq 36 \).
where

\[ \alpha(j) = \frac{\Delta x^p(j-1)}{\Delta y^p(j-1)} \]

\[ \beta(j) = \frac{\Delta y^u(j)}{\Delta x^u(j)} \]  \hspace{1cm} (B.3.5)

\[ \gamma(j) = \frac{\Delta x^p(j)}{\Delta y^p(j)} \]

\[ \delta(j) = \frac{\Delta y^u(j)}{\Delta x^u(j)} \]

for stream function \( \psi \),

\[ \alpha(j) = \frac{\Delta x^u(j)}{\Delta y^u(j)} \]

\[ \beta(j) = \frac{\Delta y^p(j)}{\Delta x^p(j)} \]  \hspace{1cm} (B.3.6)

\[ \gamma(j) = \frac{\Delta x^u(j+1)}{\Delta y^u(j+1)} \]

\[ \delta(j) = \frac{\Delta y^p(j)}{\Delta x^p(j)} \]

for scalar potential \( \chi \), and

\[ A(m,j) = \begin{cases} a(m,j) & \text{for } m = 1 \\ 2a(m,j) & \text{for } 2 \leq m \leq 36 \\ a(m,j) & \text{for } m = 37 \end{cases} \]  \hspace{1cm} (B.3.7)

\[ B(m,j) = 2b(m,j) \]  \hspace{1cm} (B.3.8)
and $f(\xi, j)$ is the right hand side of (B.1.3) or (B.1.4) for stream function or scalar potential, respectively.

After we solve (B.3.3) and (B.3.4) by Lindzen and Kuo's method, we can obtain $\psi$ and $\chi$ through (B.3.1).
The hydrologic cycle includes exchange of water among the atmosphere, hydrosphere and lithosphere through changes of physical states and dynamic processes of transports. One such exchange is between continents and oceans. The main mechanism is the runoff which transports water mostly via rivers into the oceans. In the present model the runoff rate is

\[ R = w(P_r + S_m) \]  \hspace{1cm} (C.1)

where \( w \) is the ground wetness, \( P_r \) is the rainfall, and \( S_m \) is the rate of snow melt. If the ground wetness exceeds unity, \( w \) is set equal to unity and the excess ground water is taken as additional runoff. The sum of the large-scale and convective precipitation is taken as rainfall \( (P_r) \) if \( T_4 > T_{ice} = 273.1^\circ K \). For a snow-covered surface the excess thermal energy

\[ Q_e = \Gamma_{snow}(T_g - T_{ice}) \]  \hspace{1cm} (C.2)

is explicitly utilized to melt existing snow mass \( S \) where \( \Gamma_{snow} \) is the bulk heat capacity of snow and \( T_g \) is ground temperature. If \( Q_e \leq L_f S \), where \( L_f \) is the latent heat of fusion, \( T_g \) is set equal to \( T_{ice} \) and the rate of snow melt is

\[ S_m = \frac{Q_e}{L_f \Delta t} \]  \hspace{1cm} (C.3)

where \( \Delta t \) is the time step of the model integration. If \( Q_e > L_f S \), the snow melt is given by

\[ S_m = \frac{S}{\Delta t} \]  \hspace{1cm} (C.4)

Further detailed information about the ground hydrology and thermodynamics of the model can be seen in Schlesinger and Gates (1979).