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Documentation of the
Goddard Earth Observing System (GEOS)
General Circulation Model - Version 1

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Abstract

This report is a documentation of Version 1 of the Goddard Earth Observing System (GEOS) General Circulation Model (GCM) developed by NASA's Data Assimilation Office and the Climate and Radiation Branch. The report is separated into three parts. Part I describes the overall features of the GCM, including numerical schemes for the hydrodynamics, sub-grid scale physical parameterizations, and boundary conditions. Part II provides a comprehensive description of the diagnostics available within the model, including the methodology for their selection and retrieval. Part III concludes the report with a User's Guide, which includes the location of frozen versions of the GEOS system, instructions on how to create user-defined applications for running the GEOS system, and an overview of established production applications and output utilities available to the community.
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Part I

GEOS-1 GCM: A Descriptive Documentation

1 Introduction and Model Lineage

The Goddard Earth Observing System-1 (GEOS-1) General Circulation Model (GCM) was developed by the Data Assimilation Office (DAO) at the Goddard Laboratory for Atmospheres (GLA), in collaboration with the Climate and Radiation Branch, for use in the system being developed to analyze EOS data. The GEOS Data Assimilation System (DAS), currently using the GEOS-1 GCM in conjunction with an Optimal Interpolation (OI) analysis scheme, has been used to produce a multi-year global atmospheric data set for climate research (Schubert et al., 1993). The GEOS-1 GCM has also been used to produce multiple 10-year climate simulations as part of the DAO’s participation in the Atmospheric Model Intercomparison Project (AMIP) sponsored by the Program for Climate Model Diagnostics and Intercomparison (PCMDI) (see Gates, 1992). The GEOS-1 DAS has been used operationally to provide scientific flight guidance during NASA’s participation in the Measurements for Assessing the Effects of Stratospheric Aircraft (MAESA) experiment.

The earliest predecessor of the GEOS-1 GCM was developed in 1989 based on the “plug-compatible” concepts outlined in Kalnay et al. (1989), and subsequently improved in 1991 (Fox-Rabinovitz, et al., 1991, Helfand et al., 1991). The plug-compatibility of the physical parameterizations together with the plug-compatible “Dynamical Core” introduced by Suarez and Takacs (1994) facilitate the development and testing of new algorithms. Together the DAO and the Climate and Radiation Branch at GLA have produced a library of physical parameterizations and dynamical algorithms which can be utilized for various GCM applications.

1.1 Relevant Model Documentation

A description of the overall features of the GEOS Data Assimilation System (DAS) used by the DAO may be found in Schubert et al. (1993). A comprehensive documentation of the Aries/GEOS Dynamical Core incorporating the horizontal and vertical discretization and finite-difference schemes used is given in Suarez and Takacs (1994). The Relaxed Arakawa-Schubert cumulus convective parameterization and the re-evaporation of falling rain are based upon the works of Moorthi and Suarez (1992) and Sud and Molod (1988), while the radiative processes are described by Harshvardhan et al. (1987). The planetary boundary
layer (PBL) and the upper level turbulence parameterization are based on the level 2.5 closure model of Helfand and Labraga (1988) and Helfand et al. (1991). Additional postscript copies of this document may be obtained through anonymous ftp from hera.gsfc.nasa.gov. The path and filename of the document is:

\[ \text{pub/gcm/geos1.0_gcm.doc.ps} \]

## 2 Atmospheric Dynamics

The momentum equations used in the GEOS-1 GCM are written in the “vector invariant” form, as in Sadourny (1975) and Arakawa and Lamb (1981), to facilitate the derivation of the energy and potential enstrophy conserving differencing scheme. The thermodynamic (potential temperature) and moisture (specific humidity) equations are written in flux form to facilitate potential temperature and moisture conservation. A complete description of the finite-difference scheme used can be found in Suarez and Takacs (1994).

The GEOS GCM uses a \( \sigma \) coordinate defined by

\[ \sigma = \frac{p - p_T}{\pi}, \]  

where \( p \) is the pressure, \( \pi \equiv p_s - p_T \), \( p_s \) is the surface pressure, and \( p_T \) is a constant prescribed pressure at the top of the model atmosphere. With \( p_T = 0 \) this coordinate reduces to the conventional \( \sigma \) coordinate proposed by Phillips (1957).

With this vertical coordinate, the continuity equation becomes

\[ \frac{\partial \pi}{\partial t} = -\nabla_\sigma \cdot (\pi \mathbf{v}) - \frac{\partial (\pi \sigma)}{\partial \sigma}, \]  

where \( \mathbf{v} \) is the horizontal velocity vector. Integrating (2) and assuming \( \sigma = 0 \) at \( p = p_s \) and \( p = p_T \), we obtain the forms used in the model:

\[ \frac{\partial \pi}{\partial t} = -\int_0^1 \nabla_\sigma \cdot (\pi \mathbf{v}) \, d\sigma \]  

and

\[ (\pi \sigma) = -\sigma \frac{\partial \pi}{\partial t} - \int_0^\sigma \nabla_\sigma \cdot (\pi \mathbf{v}) \, d\sigma. \]  

The equation of state for an ideal gas is \( \alpha = RT/p \), where \( \alpha \) is the specific density, \( T \) is the temperature, and \( R \) is the gas constant. The following alternative forms will be used below

\[ \alpha = \frac{R \theta}{p} = \frac{c_p \theta}{p} \frac{\partial P}{\partial p} = \frac{c_p \theta}{\pi} \left( \frac{\partial P}{\partial \sigma} \right)_\pi = \frac{c_p \theta}{\sigma} \left( \frac{\partial P}{\partial \sigma} \right)_\sigma, \]  

2
where $\theta \equiv T/P$ is the potential temperature, $P \equiv (p/p_0)^\kappa$, $\kappa = R/c_p$, $c_p$ is the specific heat at constant pressure, and $p_0$ is a reference pressure. In obtaining the forms in (5) we have used $\frac{\partial P}{\partial p} = \kappa \frac{P}{p}$ and the relation

$$
\left( \frac{dP}{d\sigma} \right)_\pi = \frac{\pi}{\sigma} \left( \frac{dP}{d\pi} \right)_\sigma.
$$

(6)

For the time being virtual effects are neglected.

The hydrostatic equation is

$$
\frac{\partial \Phi}{\partial p} = -\alpha,
$$

where $\Phi$ is the geopotential. Using (5) and (6), this can be written:

$$
\frac{\partial \Phi}{\partial \sigma} = -\pi \alpha = -c_p \theta \frac{\pi}{\sigma} \left( \frac{dP}{d\pi} \right)_\sigma = -c_p \theta \left( \frac{dP}{d\sigma} \right)_\pi.
$$

(7)

From (7) we obtain

$$
\frac{\partial \Phi}{\partial P} = -c_p \theta,
$$

(8)

which, following Arakawa and Suarez (1983), is the form used in the model.

The thermodynamic equation is written in flux form to facilitate the derivation of a $\theta$-conserving differencing scheme:

$$
\frac{\partial (\pi \theta)}{\partial t} = -\nabla_\sigma \cdot (\pi \mathbf{v} \theta) - \frac{\partial (\pi \dot{\theta})}{\partial \sigma} + \frac{\pi Q}{c_p P},
$$

(9)

where $Q$ is the diabatic heating per unit mass.

In addition to the equations of motion, the Aries/GEOS Dynamical Core computes tendencies for an arbitrary number of atmospheric constituents, such as water vapor and ozone. These are also written in flux form:

$$
\frac{\partial (\pi q^{(k)})}{\partial t} = -\nabla_\sigma \cdot (\pi \mathbf{v} q^{(k)}) - \frac{\partial (\pi \dot{q}^{(k)})}{\partial \sigma} + \pi S^{(k)},
$$

(10)

where $q^{(k)}$ is the specific mass of the $k$th constituent, and $S^{(k)}$ is its source per unit mass of air.

The momentum equation is written in "vector-invariant" form, as in Sadourny (1975) and Arakawa and Lamb (1981), to facilitate the derivation of an energy- and enstrophy-conserving differencing scheme:

$$
\frac{\partial \mathbf{v}}{\partial t} = -(f + \zeta) \mathbf{k} \times \mathbf{v} - \dot{\sigma} \frac{\partial \mathbf{v}}{\partial \sigma} - \nabla_\sigma (\Phi + K) - c_p \theta \nabla_\sigma P - \frac{g}{\pi} \frac{\partial T}{\partial \sigma},
$$

(11)

$$
= -\eta \mathbf{k} \times (\pi \mathbf{v}) - \dot{\sigma} \frac{\partial \mathbf{v}}{\partial \sigma} - \nabla_\sigma (\Phi + K) - c_p \theta \left( \frac{dP}{d\pi} \right)_\sigma \nabla_\pi - \frac{g}{\pi} \frac{\partial T}{\partial \sigma},
$$

(12)
where
\[ \eta = \frac{(f + \zeta)}{\pi} \]
is an "external" potential vorticity, \( f \) is the Coriolis parameter, \( k \) is the unit vector in the vertical,
\[ \zeta \equiv \nabla_\sigma \times \mathbf{v} \]
is the vertical component of the vorticity along \( \sigma \) surfaces,
\[ K \equiv \frac{1}{2} (\mathbf{v} \cdot \mathbf{v}) \]
is the kinetic energy per unit mass, \( g \) is the acceleration of gravity, and \( T \) is the horizontal frictional stress.

2.1 Horizontal and Vertical Discretization

The GEOS-1 GCM is constructed in the horizontal using the staggered Arakawa C-grid and employs Version 1 of the Aries/GEOS Dynamical Core for the finite-differencing algorithm. The Aries/GEOS Dynamical Core is a plug-compatible dynamics module which is used at both the DAO in the GEOS GCM as well as at the Climate and Radiation Branch in the Aries GCM. The Aries GCM has been extensively used in many climate and coupled ocean/atmosphere simulations (e.g., Schubert et al. 1993, Higgins and Schubert, 1993). Version 1 of the Aries/GEOS Dynamical Core is the second-order energy and potential enstrophy conserving scheme of Sadourny described by Burridge and Haseler (1977), while Version 2, to be used in subsequent versions of the GEOS GCM, is the fourth-order formulation described in Suarez and Takacs, 1994. This scheme conserves total energy and potential enstrophy for the non-divergent component of the flow in the shallow water equations.

The Aries/GEOS Dynamical Core uses a Lorenz or unstaggered vertical grid in generalized sigma coordinates. The vertical differencing scheme is that of Arakawa and Suarez (1983) which ensures that:

- The pressure gradient force generates no circulation of vertically integrated momentum along a contour of surface topography
- The finite-difference analogues of the energy-conversion term have the same form in the kinetic energy and thermodynamic equations
- The global mass integral of the potential temperature is conserved under adiabatic processes
- The hydrostatic equation for the lowest thickness has a local form
- The hydrostatic equation is exact for vertically isentropic atmospheres
The pressure-gradient force is exact for three-dimensionally isentropic atmospheres

The GEOS-1 AMIP simulation was run using a global $4^\circ \times 5^\circ$ latitude/longitude grid in the horizontal, together with 20 sigma levels in the vertical. The GEOS-1 DAS 5-year re-analysis was run using $2^\circ \times 2.5^\circ$ horizontal resolution. The vertical distribution of the sigma levels is such as to provide enhanced resolution in the planetary boundary layer and at the jet level. The model top pressure was prescribed to be 10 mb with a rigid lid. For a surface pressure of 1000 mb, the lowest sigma level is at 994 mb while the highest sigma level is at 19 mb. There are 5 sigma levels below 800 mb, and 7 sigma levels above 200 mb. The sigma-level pressure used in the 20-level GEOS-1 GCM are shown in Figure 1.

### 2.2 Time Integration Scheme

The GEOS GCM has the ability to use the Matsuno time integration scheme or the Leapfrog time integration scheme together with an Asselin (1972) time filter. The GEOS GCM multi-year simulations have all been run using the Leapfrog time scheme together with the Asselin averaging parameter equal to 0.05. The GEOS-DAS 5-year Re-analysis used the Matsuno time scheme. The GEOS GCM employs a somewhat unique method for incorporating adjustments due to diabatic processes (i.e., moist convection, radiation and turbulence) and the analysis increments during an assimilation. At every model time step, all prognostic fields are updated due to both dynamical and sub-grid scale diabatic processes, shown schematically using the Leapfrog time scheme for an arbitrary prognostic field $q$ as:

$$ q^{n+1} = q^{n-1} + 2\Delta t \left( \frac{\partial q^n}{\partial t} \right)_{\text{total}} $$

where

$$ \left( \frac{\partial q^n}{\partial t} \right)_{\text{total}} = \left( \frac{\partial q^n}{\partial t} \right)_{\text{Dynamics}} + \left( \frac{\partial q^{n-1}}{\partial t} \right)_{\text{Shapiro Filter}} $$

$$ + \left( \frac{\partial q}{\partial t} \right)_{\text{Moist Processes}} + \left( \frac{\partial q}{\partial t} \right)_{\text{Turbulence}} $$

$$ + \left( \frac{\partial q}{\partial t} \right)_{\text{Longwave Radiation}} + \frac{S_0}{R_a^2} \cdot \cos\phi_z \left( \frac{\partial q}{\partial t} \right)_{\text{Shortwave Radiation}} $$

$$ + \left( \frac{\partial q}{\partial t} \right)_{\text{Analysis Increment}} $$

with $S_0$ defined as the solar constant, $R_a$ as the earth-sun distance in Astronomical Units, and $\cos\phi_z$ as the cosine of the zenith angle. The Dynamics and Shapiro Filter time tendencies are updated every model time step. The diabatic time tendencies are computed
Figure 1: Sigma-level pressures used in the 20-level GEOS-1 GCM.
(updated) at a time step appropriate to the physical parameterizations, and are held constant between Physics calls. The time tendency for Moist Convection is updated every 10 minutes, and for Turbulence every 30 minutes. The time tendency for Longwave Radiation is updated every 3 hours. Shortwave Radiation is updated once every three hours assuming a normalized incident solar radiation, and adjusted at every model time step by the true incident radiation. During GEOS-1 DAS assimilations, the Analysis Increment is updated every synoptic time period, or 6 hours. By gradually incorporating the diabatic adjustments during the model integration, shocks and dynamical imbalances are greatly reduced (cf. Bloom, et al. 1991).

2.3 Smoothing / Filling

An eighth- (sixteenth-) order Shapiro (1970) filter adjustment is computed for the winds, potential temperature, and specific humidity in conjunction with the 2° x 2.5° (4° x 5°) GEOS GCM in order to globally damp small-scale dispersive waves. In order to reduce the dynamical imbalances generated by intermittent use of the full filter, a Shapiro filter tendency for any quantity \( q \) is defined as:

\[
\frac{\partial q}{\partial t}^{\text{Shapiro Filter}} = \frac{q^F - q}{\tau}
\]  

where \( q^F \) is the quantity after application of the full Shapiro filter, \( q \) is the unfiltered quantity, and \( \tau \) is an adjustable time scale. Thus, a fraction \( \Delta t / \tau \) of the full Shapiro filter is incorporated at each model timestep. In the GEOS-1 GCM, the time scale \( \tau \) is equal to 1.5 hours, and has been chosen so as to remove the two-grid interval wave in about six hours. The Shapiro filter response function used in the 2° x 2.5° GEOS-1 GCM is shown in Figure 2. The GEOS-1 GCM also uses a high-latitude Fourier filter to avoid linear instability due to violation of the CFL condition for the Lamb wave and internal gravity waves near the poles. This polar filter is only applied to the time-tendencies of the prognostic fields, ie. winds, potential temperature, specific humidity, and surface pressure.

For the moisture equation, negative values computationally generated through advection are filled by “borrowing” from below while conserving the vertically integrated moisture, ie.,

\[
\left( \int_{le+1}^{le-1} \rho q \, dz \right)_{\text{final}} = \left( \int_{le+1}^{le-1} \rho q \, dz \right)_{\text{initial}}
\]  

Using the hydrostatic relation \( \rho \delta z = -\frac{\delta p}{g} = -\frac{\Delta \rho}{\Delta \sigma} \), we may write
Figure 2: Shapiro filter response function used in the 2° x 2.5° GEOS-1 GCM.
Approximating equation (17) by

\[
\left( \int_{l_{e+1}}^{l_{e+1}} \pi q d\sigma \right)_{\text{final}} = \left( \int_{l_{e+1}}^{l_{e+1}} \pi q d\sigma \right)_{\text{initial}}
\]  

an expression for the updated moisture at level \( l \) is given by

\[
\pi q_{l_{\text{final}}} = \left( \pi q_l + \pi q_{l-1} \frac{\Delta \sigma_{l-1}}{\Delta \sigma_l} \right)_{\text{initial}} - \left( \pi q_{l-1} \frac{\Delta \sigma_{l-1}}{\Delta \sigma_l} \right)_{\text{final}}
\]  

Assuming that \( \pi q_{l-1} \) is initially negative, we require that its final value is set to zero. Thus

\[
\pi q_{l-1_{\text{final}}} = 0,
\]

\[
\pi q_{l_{\text{final}}} = \left( \pi q_l + \pi q_{l-1} \frac{\Delta \sigma_{l-1}}{\Delta \sigma_l} \right)_{\text{initial}}
\]

This process is repeated until the lowest level is reached. If the resulting \( \pi q_{n_{\text{lay}}} \) is negative, the mass-weighted specific humidity in the lowest model level is simply set to zero.

3 Atmospheric Physics

3.1 Moist Convective Processes

3.1.1 Sub-grid and Large-scale Convection

Sub-grid scale penetrative and shallow cumulus convection are parameterized using the Relaxed Arakawa Schubert (RAS) scheme of Moorthi and Suarez (1992), which is a simplified Arakawa Schubert type scheme. The RAS scheme predicts mass fluxes from cloud types which have different entrainment rates and levels of neutral buoyancy, depending on the properties of the large scale environment.

The thermodynamic variables that are used in RAS to describe the grid scale vertical profile are the dry static energy, \( s = c_p T + gz \), the moist static energy, \( h = c_p T + gz + Lq \), and the saturation moist static energy, \( h^* = c_p T + gz + Lq^* \), where the superscript \( * \) refers to the saturation value.
The conceptual model behind RAS depicts the cloud as a rising plume, entraining mass from the environment during ascent, and detraining mass at the level of neutral buoyancy. RAS assumes that the cloud mass flux, $\eta$, normalized by the cloud base mass flux, is a linear function of height, expressed as:

$$\frac{\partial \eta(z)}{\partial z} = \lambda \quad \text{or} \quad \frac{\partial \eta(P^c)}{\partial P^c} = -\frac{c_p \theta \lambda}{g}$$

where we have used the hydrostatic equation written in the form:

$$\frac{\partial z}{\partial P^c} = -\frac{c_p \theta}{g}$$

The entrainment parameter, $\lambda$, characterizes a particular cloud type based on its detrainment level, and is obtained by assuming that the level of detrainment is the level of neutral buoyancy, i.e., the level at which the moist static energy of the cloud, $h_c$, is equal to the saturation moist static energy of the environment, $h^*$. Following Moorthi and Suarez (1992), $\lambda$ may be written as

$$\lambda = \frac{h_B - h^*_D}{\frac{c_p}{g} \int_{P_D}^{P_B} \theta(h^*_D - h) dP^c},$$

where the subscript $B$ refers to cloud base, and the subscript $D$ refers to the detrainment level.

The convective instability is measured in terms of the cloud work function $A$, defined as the rate of change of cumulus kinetic energy. The cloud work function is related to the buoyancy, or the difference between the moist static energy in the cloud and in the environment:

$$A = \int_{P^c}^{P_B} \eta \left(\frac{h_c - h^*}{P^c}\right) dP^c$$

where $\gamma$ is $\frac{L}{c_p \beta T}$ obtained from the Clausius Clapeyron equation, and the subscript $c$ refers to the value inside the cloud.

To determine the cloud base mass flux, the rate of change of $A$ in time due to dissipation by the clouds is assumed to approximately balance the rate of change of $A$ due to the generation by the large scale. This is the quasi-equilibrium assumption, and results in an expression for $m_B$:

$$m_B = -\frac{dA}{dt} \frac{1}{K}$$

where $K$ is the cloud kernel, defined as the rate of change of the cloud work function per unit cloud base mass flux, and is obtained by differentiating the expression for $A$ in time. The rate of change of $A$ due to the generation by the large scale is written as the difference between the current $A(t + \Delta t)$ and its value after the previous convective time step $A(t)$, divided by the time step. Since the convective parameterization is designed to nearly neutralize the instability, $A(t)$ is approximated as some (near zero) critical $A_{crit}$. 
The predicted convective mass fluxes are used to solve grid-scale temperature and moisture budget equations to determine the impact of convection on the large scale fields of temperature (through latent heating and compensating subsidence) and moisture (through precipitation and detrainment):

\[
\frac{\partial \theta}{\partial t} = \alpha R \frac{m_B}{c_p} \frac{\partial s}{\partial p}
\]

and

\[
\frac{\partial q}{\partial t} = \alpha R \frac{m_B}{L} g \eta \left( \frac{\partial h}{\partial p} - \frac{\partial s}{\partial p} \right)
\]

where \( \theta = \frac{T}{P} \), and \( P = \frac{p}{p_0} \).

As an approximation to a full interaction between the different allowable cloud types, many clouds are simulated frequently, each modifying the large scale environment some fraction \( \alpha \) of the total adjustment. The parameterization thereby "relaxes" the large scale environment towards neutrality.

In addition to the RAS cumulus convection scheme, the GEOS-1 GCM employs a Kessler-type scheme for the re-evaporation of falling rain (Sud and Molod, 1988), which correspondingly adjusts the impact on the large scale environment by the factor \( R \). The scheme accounts for the rainfall intensity, the drop size distribution, and the temperature, pressure and relative humidity of the surrounding air. The moisture deficit in any model layer into which the rain may re-evaporate is a free parameter.

Due to the increased vertical resolution in the Planetary Boundary Layer (PBL), the lowest two model layers are averaged to provide the sub-cloud layer for RAS (nominally 50 mb thick). Each time RAS is invoked (every ten simulated minutes), the possibility for shallow convection is checked for the two layers just above cloud base. RAS also randomly chooses 10 other cloud-top levels for the possibility of convection, from just above cloud base to the model top layer.

Supersaturation or large-scale convection is defined in the GEOS GCM whenever the specific humidity in any grid-box exceeds its supersaturation value. The large-scale precipitation scheme rains at supersaturation, and re-evaporates during descent to partially saturate lower layers in a process that accounts for some simple microphysics.

### 3.1.2 Cloud Formation

Convective and large-scale cloudiness which is used for cloud-radiative interactions are determined diagnostically as part of the cumulus and large-scale parameterizations. The convective and large-scale cloud fractions are combined into two separate arrays for use in
the shortwave and longwave radiation packages, one array for random overlap (CLRO) and one array for maximum overlap (CLMO) cloudiness.

If convection occurs with a cloud-top pressure less than 400 mb, a cloud fraction equal to 1.0 is assigned into CLMO at the detrainment level. These tall convective towers are said to be maximally overlapped, ie. they are radiatively correlated in the vertical. If shallow convection occurs with a detrainment level within 1 or 2 levels above the cloud base, a cloud fraction equal to 0.5 is assigned into CLRO, again at the detrainment level. These shallow clouds are not correlated in the vertical and are said to be randomly overlapped. No cloud fractions are prescribed to convection whose cloud-top is between 400 mb and 2 levels above cloud-base.

Supersaturation or large scale cloudiness is defined whenever the large scale precipitation scheme determines that the grid box at any level becomes supersaturated. In order to ensure that at any instant the total cloud fraction be less than or equal to 1.0, supersaturation clouds are only prescribed when there are no deep convective clouds. Under such conditions, the grid box is assumed to be instantaneously fully covered with randomly overlapped clouds, with a cloud fraction of 1.0 being assigned into CLRO.

3.2 Radiation

The parameterization of radiation includes both shortwave radiation and longwave radiation in the GEOS-1 GCM. A single model “sponge” layer is inserted into the GEOS-1 GCM above $P_{Top}$ with a temperature equal to the GCM’s first level temperature. This is done in order to avoid exposing the first GCM level to a zero downward thermal flux condition at $P_{Top}$. Radiative fluxes are calculated at each model edge-level in both up and down directions. The heating rates/cooling rates are then obtained from the vertical divergence of the net radiative fluxes.

The net flux is

$$F = F^u - F^d$$

where $F$ is the net flux, $F^u$ is the upward flux and $F^d$ is the downward flux.

The heating rate due to the divergence of the radiative flux is given by

$$\frac{\partial \rho c_p T}{\partial t} = -\frac{\partial F}{\partial z}$$

or

$$\frac{\partial T}{\partial t} = \frac{g}{c_p \pi} \frac{\partial F}{\partial \sigma}$$

where $g$ is the acceleration due to gravity and $c_p$ is the heat capacity of air at constant pressure.
The infrared and solar radiation parameterizations employed in the GEOS-1 GCM follow closely those described by Harshvardhan et al. (1987). The time tendency for Longwave Radiation is updated every 3 hours. The time tendency for Shortwave Radiation is updated once every three hours assuming a normalized incident solar radiation, and subsequently modified at every model time step by the true incident radiation. For the AMIP simulations, the GEOS-1 GCM used the AMIP-prescribed solar constant value of 1365 $W/m^2$ and a $CO_2$ mixing ratio of 345 ppm. For the 5-year GEOS-DAS re-analysis, a solar constant value of 1380 $W/m^2$ was used together with a $CO_2$ mixing ratio value of 330 ppm. For the ozone mixing ratio, monthly mean zonally averaged climatological values specified as a function of latitude and height (Rosenfield, et al., 1987) are linearly interpolated to the current time.

### 3.2.1 Shortwave Radiation

The parameterization of heating due to shortwave radiation is an extension of that developed by Lacis and Hansen (1974). The documentation of the computational aspects of the scheme are given in Davis (1982). Harshvardhan improved the treatment of the cloud scattering by allowing it to depend on the solar zenith angle. Two major absorbers, $H_2O$ and $O_3$, are considered in the scheme. Radiative heating by water vapor absorption is modeled in the 0.7-0.4 $\mu m$ region. The total absorption of the ozone includes the absorption of visible radiation by the Chappuis bands (450-750 nm) and the absorption of ultraviolet radiation by the Hartly (240-280 nm) and Huggins bands (280-360 nm). For a cloudy atmosphere, the Meadow-Weaver (1980) modified Eddington approximation is adopted to solve the radiative transfer equation in a scattering medium. The multiple-scattering computation with the k-distribution functions is used to calculate the cloudy-sky water absorption.

### 3.2.2 Longwave Radiation

The parameterization of longwave radiation employs a wide band model. Four broadband transmissions have been used in the model. The parameterization of the $H_2O$, containing the two water bands at 15 $\mu m$ and 9.6 $\mu m$, is calculated using the method of Chou (1984) based on the water vapor line and water vapor e-type absorption approximations. The $CO_2$ band employs the scheme of Chou and Peng (1983) which separates the band wing and band center scaled paths. The $O_3$ parameterization approach applies the formulation of Rosenfield et al. (1987), which modifies the line width used by Rodgers (1968) to include the affects of Doppler broadening.
3.2.3 Cloud-Radiation Interaction

The cloud fraction values produced by the Moist Convective Processes are used for cloud-radiation interactions as a means for evaluating clear line-of-site probabilities and effective optical thicknesses.

If we define the time-averaged random and maximum overlapped cloudiness as CLRO and CLMO respectively, then the probability of clear sky associated with random overlapped clouds at any level is \((1-CLRO)\) while the probability of clear sky associated with maximum overlapped clouds at any level is \((1-CLMO)\). The total clear sky probability is given by \((1-CLRO)(1-CLMO)\), thus the total cloud fraction at each level may be obtained by \(1-(1-CLRO)(1-CLMO)\).

At any given level, we may define the clear line-of-site probability by appropriately accounting for the maximum and random overlap cloudiness. The clear line-of-site probability is defined to be equal to the product of the clear line-of-site probabilities associated with random and maximum overlap cloudiness. The clear line-of-site probability \(C(p, p')\) associated with maximum overlap clouds, from the current pressure \(p\) to the model top pressure, \(p' = p_{top}\), or the model surface pressure, \(p' = p_{surf}\), is simply \(1.0\) minus the largest maximum overlap cloud value along the line-of-site, ie.

\[
1 - MAX^\prime_p (CLMO_p)
\]

Thus, even in the time-averaged sense it is assumed that the maximum overlap clouds are correlated in the vertical. The clear line-of-site probability associated with random overlap clouds is defined to be the product of the clear sky probabilities at each level along the line-of-site, ie.

\[
\prod_{p'} (1 - CLR^\prime_O_p)
\]

The total cloud fraction at a given level associated with a line-of-site calculation is given by

\[
1 - \left(1 - MAX^\prime_p [CLMO_p]\right) \prod_{p'} (1 - CLR^\prime_O_p)
\]

The GEOS-1 GCM cloud optical thicknesses are specified based on cloud type and temperature. The “maximum overlap” clouds are assigned an optical thickness of \(0.16 \text{ km}^{-1}\)
for the shortwave radiative feedback. The "random overlap" clouds are assigned an optical thickness based on an empirical relation between local temperature and optical thickness. The relation gives a thicker cloud when the temperature is higher and is expressed:

\[
\tau = \begin{cases} 
0 & \text{for } T < 190.66 \\
(T - 190.66)^2 \times 2 \times 10^{-6} & \text{for } 190.66 < T \leq 263.16 \\
6.95 \times 10^{-3} \times T - 1.82 & \text{for } 263.16 < T \leq 273.38 \\
0.08 & \text{for } 273.38 < T 
\end{cases}
\]

The relation for the range $190.66 < T \leq 263.16$ is from Platt and Harshvardahn, 1988. The optical thickness for the longwave radiative feedback is assumed to be 75% of these values.

The entire Moist Convective Processes Module is called with a frequency of 10 minutes. The cloud fraction values are time-averaged over the period between Radiation calls (every 3 hours). Therefore, in a time-averaged sense, both random overlap and maximum overlap cloudiness can exist in a given grid-box.

### 3.3 Turbulence

The GEOS GCM turbulence parameterization consists of elements which handle vertical diffusion (Helfand and Labraga, 1988) and surface fluxes of heat, moisture and momentum (Helfand, et al, 1991, and Helfand and Schubert, 1994). The parameterization is invoked every 30 minutes, and employs a backward-implicit iterative time scheme with an internal time step of 5 minutes. The vertical regime is divided into a free atmosphere, a surface layer, and a viscous sub-layer above the surface roughness elements. The turbulent eddy fluxes are calculated using a variety of methods depending on the vertical location in the atmosphere.

Turbulent eddy fluxes of momentum, heat and moisture in the surface layer are calculated using stability-dependant bulk formulae based on Monin-Obukov similarity functions. For an unstable surface layer, the chosen stability functions are the KEYPS function (Panofsky, 1973) for momentum, and its generalization for heat and moisture. The function for heat and moisture assures non-vanishing heat and moisture fluxes as the wind speed approaches zero. For a stable surface layer, the stability functions are those of Clarke (1970), slightly modified for the momentum flux. The moisture flux also depends on a specified evapotranspiration coefficient, set to unity over oceans and dependant on the climatological ground wetness over land. The gradients in the viscous sublayer are based on Yaglom and Kader (1974). The surface roughness length over oceans is computed as an interpolation between the functions of Large and Pond (1981) for high winds and of Kondo (1975) for weak winds, and over land is specified from the climatology of Dorman and Sellers (1989).
Above the surface layer, turbulent fluxes of momentum, heat and moisture are calculated by the Level 2.5 Mellor-Yamada type closure scheme of Helfand and Labraga (1988), which predicts turbulent kinetic energy and determines the eddy transfer coefficients used for vertical diffusion.

3.3.1 Planetary Boundary Layer

The Planetary Boundary Layer (PBL) height is diagnosed by the parameterization as the level at which the turbulent kinetic energy is reduced to a tenth of its surface value, and generally occurs within the first two to four sigma levels above the surface (sigma equal 0.994 to about sigma equal 0.875).

4 Boundary Conditions

The conditions at the lower boundary which are not explicitly predicted during execution must be specified from observed fields. Boundary condition data sets are available at present in monthly mean increments, at the model's 4° x 5° and 2° x 2.5° resolutions, for either climatological or yearly varying conditions (beginning from 1979). Monthly mean values are interpolated during each model timestep to the current time. Future model versions will incorporate boundary conditions at higher spatial (1° x 1°) and temporal resolutions.

4.1 Orography

Surface geopotential heights are provided from an averaging of the Navy 10 minute by 10 minute dataset supplied by the National Center for Atmospheric Research (NCAR) to the model's grid resolution. For the 2° x 2.5° GEOS-1 GCM, the averaged topography was then passed through a Lanczos (1966) filter in both dimensions which removes the smallest scales while inhibiting Gibbs phenomena. Negative values in the topography resulting from the filtering procedure were not filled.

In one dimension, we may define a cyclic function in $x$ as:

$$f(x) = \frac{a_0}{2} + \sum_{k=1}^{N} (a_k \cos(kx) + b_k \sin(kx))$$

where $N = \frac{IM}{2}$ and IM is the total number of points in the $x$ direction. Defining $\Delta x = \frac{2\pi}{IM}$, we may define the average of $f(x)$ over a $2\Delta x$ region as:
\[
\overline{f(x)} = \frac{1}{2\Delta x} \int_{x-\Delta x}^{x+\Delta x} f(x') \, dx'
\]  
(22)

Using equation (21) in equation (22) and integrating, we may write:

\[
\overline{f(x)} = \frac{a_0}{2} + \frac{1}{2\Delta x} \sum_{k=1}^{N} \left[ a_k \frac{\sin(kx')}{k} \right]_{x-\Delta x}^{x+\Delta x} - b_k \frac{\cos(kx')}{k} \right]_{x-\Delta x}^{x+\Delta x}
\]  
(23)

or

\[
\overline{f(x)} = \frac{a_0}{2} + \sum_{k=1}^{N} \frac{\sin(k\Delta x)}{k\Delta x} \left( a_k \cos(kx) + b_k \sin(kx) \right)
\]  
(24)

Thus, the Fourier wave amplitudes are simply modified by the Lanczos filter response function \( \sin(k\Delta x) / k\Delta x \). This may be compared with an \( m \)th-order Shapiro (1970) filter response function, defined as \( 1 - \sin^m \left( \frac{k\Delta x}{2} \right) \), shown in Figure 3.

### 4.2 Land-Sea-Ice Surface Type

The determination of the land or sea category of surface type was made from NCAR’s 10 minute by 10 minute Navy topography dataset, which includes information about the percentage of water-cover at any point. The data were averaged to the model’s 4° x 5° and 2° x 2.5° grid resolutions. Any grid-box whose averaged water percentage was \( \geq 60\% \) was defined as a water point. Permanent ice or glacier points for the 2° x 2.5° grid were identified from the climatology of Parkinson’s satellite-derived fields (NASA SP 459: Arctic Sea Ice, 1973-1976, and NASA SP 489: Antarctic Sea Ice, 1973-1976). The Land/Water masks and topography used for the 4° x 5° and 2° x 2.5° GEOS-1 GCM grids are shown in Figure 4 and Figure 5, respectively.

### 4.3 Sea Surface Temperature

Climatological sea surface temperatures for the 4° x 5° and 2° x 2.5° grid resolutions are interpolated from the data of Reynolds, 1988. Yearly varying monthly mean data for the 4° x 5° grid are from the AMIP specified monthly mean fields (Gates, 1992), and for the 2° x 2.5° grid are interpolated from Reynolds’ (1988) data. Monthly mean data are interpolated linearly to the current time at every GCM timestep.
Figure 3: Comparison between the Lanczos and $m$th-order Shapiro filter response functions for $m = 2, 4, \text{ and } 8$. 
Figure 4: GEOS-1 GCM Land/Water Mask and Topography used at 4° x 5° resolution. Light grey shading denotes land surfaces, while dark grey denotes permanent glacier.
Figure 5: GEOS-1 GCM Land/Water Mask and Topography used at 2° x 2.5° resolution. Light grey shading denotes land surfaces, while dark grey denotes permanent glacier.
4.4 Sea Surface Roughness

Ocean surface roughness is computed iteratively as a function of wind stress in the Turbu-

lence Scheme.

4.5 Albedo

Monthly mean albedos over land are data of Posey and Clapp (1964), with modifications as
described in Kitzmiller (1979). Over oceans the albedo is specified as 0.07, and over ice (sea
ice or glaciers) 0.80. Monthly mean values are linearly interpolated to the current time.

4.6 Sea Ice

Monthly mean sea ice extent were interpolated to the model's 4° x 5° and 2° x 2.5° grid res-
olutions from the AMIP monthly mean fields. Monthly mean data are linearly interpolated
to the current time. Any sea ice is assumed to be three meters thick, and the conduction
through sea ice is accounted for in the Turbulence Parameterization as part of the surface
energy budget.

4.7 Snow Cover

Snow cover is assumed to occur when the ground temperature over the land is less than
freezing in conjunction with climatological albedo values greater than 0.4. Surface conduc-
tive characteristics are adjusted accordingly.

4.8 Land Surface

Monthly varying climatological roughness lengths are specified for each land surface vege-
tation type from Dorman and Sellers (1989). A gridded surface vegetation type designation
is from Sud, et al. (1990). Monthly mean values are interpolated linearly to the model's
current time. The soil wetness climatology and yearly varying fields are from the monthly
estimates of Schemm, et al. (1992) based on the procedure developed by Mintz and Serafini
(1984) using an inverted single layer “bucket” model together with observed fields of surface
air temperature and precipitation. Values are interpolated linearly to the model's current
time. Ground temperature over land is predicted from a surface energy balance equation.
Part II

GEOS-1 GCM: Diagnostics

5 Introduction

This section of the documentation describes the Diagnostics Utilities available within the Goddard Earth Observing Systems (GEOS) General Circulation Model (GCM). In addition to a description on how to set and extract diagnostic quantities, this document also provides a comprehensive list of all available diagnostic quantities and a short description of how they are computed. It should be noted that this document is not intended to be a complete documentation of the various physical parameterization schemes used in the GEOS GCM, and the reader should refer to original publications for further insight.

6 Diagnostic Overview

A large selection of model diagnostics is available in the GEOS GCM. At the time of this writing there are 119 different diagnostic quantities which can be enabled for an experiment. As a matter of philosophy, no diagnostic is enabled as default, thus each user must specify the exact diagnostic information required for an experiment. This is accomplished by enabling the specific diagnostic of interest cataloged in the Diagnostic Menu (see Section 7). The Diagnostic Menu is a hard-wired enumeration of diagnostic quantities available within the GEOS GCM. Diagnostics are internally referred to by their associated number in the Diagnostic Menu. Once a diagnostic is enabled, the GEOS GCM will continually increment an array specifically allocated for that diagnostic whenever the associated process for the diagnostic is computed. Separate arrays are used both for the diagnostic quantity and its diagnostic counter which records how many times each diagnostic quantity has been computed.

There are several utilities within the GEOS GCM available to users to enable, disable, clear, and retrieve model diagnostics, and may be called from any user-supplied application and/or output routine. The available utilities and the CALL sequences are listed below.

SETDIAG: This subroutine enables a diagnostic from the Diagnostic Menu, meaning that space is allocated for the diagnostic and the model routines will increment the diagnostic value during execution. This routine is useful when called from either user application routines or user output routines, and is the underlying interface between the user and the desired diagnostic. The diagnostic is referenced by its diagnostic number from the menu, and its calling sequence is given by:
CALL SETDIAG (NUM)

where

NUM = Diagnostic number from menu

GETDIAG: This subroutine retrieves the value of a model diagnostic. This routine is particularly useful when called from a user output routine, although it can be called from an application routine as well. This routine returns the time-averaged value of the diagnostic by dividing the current accumulated diagnostic value by its corresponding counter. This routine does not change the value of the diagnostic itself, that is, it does not replace the diagnostic with its time-average. The calling sequence for this routine is given by:

CALL GETDIAG (LEV, NUM, QTMP, UNDEF)

where

LEV = Sigma Level at which the diagnostic is desired
NUM = Diagnostic number from menu
QTMP = Time-Averaged Diagnostic Output
UNDEF = Fill value to be used when diagnostic is undefined

CLRDIAG: This subroutine initializes the values of model diagnostics to zero, and is particularly useful when called from user output routines to re-initialize diagnostics during the run. The calling sequence is:

CALL CLRDIAG (NUM)

where

NUM = Diagnostic number from menu

ZAPDIAG: This entry into subroutine SETDIAG disables model diagnostics, meaning that the diagnostic is no longer available to the user. The memory previously allocated to the diagnostic is released when ZAPDIAG is invoked. The calling sequence is given by:

CALL ZAPDIAG (NUM)

where

NUM = Diagnostic number from menu

DIAGSIZE: We end this section with a discussion on the manner in which computer memory is allocated for GEOS diagnostics. All GEOS GCM diagnostic quantities are
stored in the single diagnostic array QDIAG which is located in the DIAG COMMON, having the form:

\[
\text{COMMON /DIAG/ NDIAG\_MAX,QDIAG(IM,JM,1)}
\]

where NDIAG\_MAX is an Integer variable which should be set equal to the number of enabled diagnostics, and QDIAG is a three-dimensional array. The first two-dimensions of QDIAG correspond to the horizontal dimension of a given diagnostic, while the third dimension of QDIAG is used to identify specific diagnostic types. In order to minimize the maximum memory requirement used by the model, the GEOS GCM libraries are compiled with room for only one horizontal diagnostic array, as shown in the above example. In order for the User to enable more than 1 two-dimensional diagnostic, the size of the DIAG COMMON must be expanded to accommodate the desired diagnostics. This can be accomplished by including the COMMON DIAG in the User's Application program, with the QDIAG array properly dimensioned. Another method, currently used by DAO production routines, involves the use of the DIAGSIZE subroutine. Using the DIAGSIZE subroutine is a convenient method to allocate just enough permanent memory during a run to be used for diagnostic storage. All production DAO applications contain a call to subroutine DIAGSIZE, which expands the size of the DIAG COMMON to include the total number of surface and upper-air diagnostics the User has enabled. An example of the DIAGSIZE subroutine is shown below:

```fortran
subroutine diagsize
  parameter ( IM = 144 )
  parameter ( JM = 91 )
  parameter ( NLAY = 20 )
  parameter ( ndiags = 11 )
  parameter ( ndiagu = 4 )
  common /diag/ ndiag_max,qdiag(im,jm,ndiags+ndiagu*nlay)
  ndiag_max = ndiags + ndiagu*nlay
  return
end
```
## 7 GEOS GCM Diagnostic Menu

<table>
<thead>
<tr>
<th>N</th>
<th>NAME</th>
<th>UNITS</th>
<th>LEVELS</th>
<th>DESCRIPTION</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>UFLUX</td>
<td>Newton/m²</td>
<td>1</td>
<td>Surface U-Wind Stress on the atmosphere</td>
</tr>
<tr>
<td>2</td>
<td>VFLUX</td>
<td>Newton/m²</td>
<td>1</td>
<td>Surface V-Wind Stress on the atmosphere</td>
</tr>
<tr>
<td>3</td>
<td>HFLUX</td>
<td>Watts/m²</td>
<td>1</td>
<td>Surface Flux of Sensible Heat</td>
</tr>
<tr>
<td>4</td>
<td>EFLUX</td>
<td>Watts/m²</td>
<td>1</td>
<td>Surface Flux of Latent Heat</td>
</tr>
<tr>
<td>5</td>
<td>QICE</td>
<td>Watts/m²</td>
<td>1</td>
<td>Heat Conduction through Sea-Ice</td>
</tr>
<tr>
<td>6</td>
<td>RADLWG</td>
<td>Watts/m²</td>
<td>1</td>
<td>Net upward LW flux at the ground</td>
</tr>
<tr>
<td>7</td>
<td>RADSWG</td>
<td>Watts/m²</td>
<td>1</td>
<td>Net downward SW flux at the ground</td>
</tr>
<tr>
<td>8</td>
<td>RI</td>
<td>deg/day</td>
<td>NLAY</td>
<td>Richardson Number</td>
</tr>
<tr>
<td>9</td>
<td>CT</td>
<td>m/sec</td>
<td>1</td>
<td>Surface Drag coefficient for T and Q</td>
</tr>
<tr>
<td>10</td>
<td>CU</td>
<td>m/sec</td>
<td>1</td>
<td>Surface Drag coefficient for U and V</td>
</tr>
<tr>
<td>11</td>
<td>ET</td>
<td>m²/sec</td>
<td>NLAY</td>
<td>Diffusivity coefficient for T and Q</td>
</tr>
<tr>
<td>12</td>
<td>EU</td>
<td>m²/sec</td>
<td>NLAY</td>
<td>Diffusivity coefficient for U and V</td>
</tr>
<tr>
<td>13</td>
<td>TURBU</td>
<td>m/sec/day</td>
<td>NLAY</td>
<td>U-Momentum Changes due to Turbulence</td>
</tr>
<tr>
<td>14</td>
<td>TURBV</td>
<td>m/sec/day</td>
<td>NLAY</td>
<td>V-Momentum Changes due to Turbulence</td>
</tr>
<tr>
<td>15</td>
<td>TURBT</td>
<td>deg/day</td>
<td>NLAY</td>
<td>Temperature Changes due to Turbulence</td>
</tr>
<tr>
<td>16</td>
<td>TURBQ</td>
<td>g/kg/day</td>
<td>NLAY</td>
<td>Specific Humidity Changes due to Turbulence</td>
</tr>
<tr>
<td>17</td>
<td>MOISTT</td>
<td>deg/day</td>
<td>NLAY</td>
<td>Temperature Changes due to Moist Processes</td>
</tr>
<tr>
<td>18</td>
<td>MOISTQ</td>
<td>g/kg/day</td>
<td>NLAY</td>
<td>Specific Humidity Changes due to Moist Processes</td>
</tr>
<tr>
<td>19</td>
<td>RADLW</td>
<td>deg/day</td>
<td>NLAY</td>
<td>Net Longwave heating rate for each level</td>
</tr>
<tr>
<td>20</td>
<td>RADSW</td>
<td>deg/day</td>
<td>NLAY</td>
<td>Net Shortwave heating rate for each level</td>
</tr>
<tr>
<td>21</td>
<td>PREACC</td>
<td>mm/day</td>
<td>1</td>
<td>Total Precipitation</td>
</tr>
<tr>
<td>22</td>
<td>PRECON</td>
<td>mm/day</td>
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<td>Convective Precipitation</td>
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<td>TUFUX</td>
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<td>NLAY</td>
<td>Turbulent Flux of U-Momentum</td>
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<td>TVFLUX</td>
<td>Newton/m²</td>
<td>NLAY</td>
<td>Turbulent Flux of V-Momentum</td>
</tr>
<tr>
<td>25</td>
<td>TTFUX</td>
<td>Watts/m²</td>
<td>NLAY</td>
<td>Turbulent Flux of Sensible Heat</td>
</tr>
<tr>
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<td>TQFLUX</td>
<td>Watts/m²</td>
<td>NLAY</td>
<td>Turbulent Flux of Latent Heat</td>
</tr>
<tr>
<td>27</td>
<td>CN</td>
<td>m/sec</td>
<td>1</td>
<td>Neutral Drag Coefficient</td>
</tr>
<tr>
<td>28</td>
<td>WINDS</td>
<td>m/sec</td>
<td>1</td>
<td>Surface Wind Speed</td>
</tr>
<tr>
<td>29</td>
<td>DTSRF</td>
<td>deg</td>
<td>1</td>
<td>Air/Surface virtual temperature difference</td>
</tr>
<tr>
<td>30</td>
<td>TG</td>
<td>deg</td>
<td>1</td>
<td>Ground temperature</td>
</tr>
<tr>
<td>31</td>
<td>TS</td>
<td>deg</td>
<td>1</td>
<td>Surface air temperature (Adiabatic from lowest model layer)</td>
</tr>
<tr>
<td>32</td>
<td>DTG</td>
<td>deg</td>
<td>1</td>
<td>Ground temperature adjustment</td>
</tr>
<tr>
<td>N</td>
<td>NAME</td>
<td>UNITS</td>
<td>LEVELS</td>
<td>DESCRIPTION</td>
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<td>---------</td>
<td>--------</td>
<td>-----------------------------------------------------------------------------</td>
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<tr>
<td>33</td>
<td>QG</td>
<td>g/kg</td>
<td>1</td>
<td>Ground specific humidity</td>
</tr>
<tr>
<td>34</td>
<td>QS</td>
<td>g/kg</td>
<td>1</td>
<td>Saturation surface specific humidity</td>
</tr>
<tr>
<td>35</td>
<td>TGRLW</td>
<td>deg</td>
<td>1</td>
<td>Instantaneous ground temperature used as input to the Longwave radiation subroutine</td>
</tr>
<tr>
<td>36</td>
<td>ST4</td>
<td>Watts/m²</td>
<td>1</td>
<td>Upward Longwave flux at the ground (σT⁴)</td>
</tr>
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<td>OLR</td>
<td>Watts/m²</td>
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<td>Net upward Longwave flux at the top of the model</td>
</tr>
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<td>OLRCLR</td>
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<td>1</td>
<td>Net upward clearsky Longwave flux at the top of the model</td>
</tr>
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<td>LWGCLR</td>
<td>Watts/m²</td>
<td>1</td>
<td>Net upward clearsky Longwave flux at the ground</td>
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<td>LWCLR</td>
<td>deg/day</td>
<td>NLAY</td>
<td>Net clearsky Longwave heating rate for each level</td>
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<td>TLW</td>
<td>deg</td>
<td>NLAY</td>
<td>Instantaneous temperature used as input to the Longwave radiation subroutine</td>
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<td>SHLW</td>
<td>g/g</td>
<td>NLAY</td>
<td>Instantaneous specific humidity used as input to the Longwave radiation subroutine</td>
</tr>
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<td>OZLW</td>
<td>g/g</td>
<td>NLAY</td>
<td>Instantaneous ozone used as input to the Longwave radiation subroutine</td>
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<td>CLMOLW</td>
<td>0–1</td>
<td>NLAY</td>
<td>Maximum overlap cloud fraction used in the Longwave radiation subroutine</td>
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<td>CLROILW</td>
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<td>NLAY</td>
<td>Random overlap cloud fraction used in the Longwave radiation subroutine</td>
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<td>CLMOSW</td>
<td>0–1</td>
<td>NLAY</td>
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</tr>
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<td>CLROSW</td>
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<td>NLAY</td>
<td>Random overlap cloud fraction used in the Shortwave radiation subroutine</td>
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<td>RADSWT</td>
<td>Watts/m²</td>
<td>1</td>
<td>Incident Shortwave radiation at the top of the atmosphere</td>
</tr>
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<td>EVAP</td>
<td>mm/day</td>
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<td>Surface evaporation</td>
</tr>
<tr>
<td>50</td>
<td>DPDT</td>
<td>mm/day</td>
<td>1</td>
<td>Total surface pressure tendency</td>
</tr>
<tr>
<td>51</td>
<td>ANALU</td>
<td>m/sec/sec</td>
<td>NLAY</td>
<td>Analysis u – Wind increment</td>
</tr>
<tr>
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<td>ANALV</td>
<td>m/sec/sec</td>
<td>NLAY</td>
<td>Analysis v – Wind increment</td>
</tr>
<tr>
<td>53</td>
<td>ANALT</td>
<td>mb deg/mb²/sec</td>
<td>NLAY</td>
<td>Analysis θ increment, θ = $\frac{\hat{T}}{\hat{p}}$</td>
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<td>54</td>
<td>ANALQ</td>
<td>mb + g/g/sec</td>
<td>NLAY</td>
<td>Analysis θq increment</td>
</tr>
<tr>
<td>55</td>
<td>OMEGA</td>
<td>mb/day</td>
<td>NLAY</td>
<td>Vertical (Omega) velocity</td>
</tr>
<tr>
<td>N</td>
<td>NAME</td>
<td>UNITS</td>
<td>LEVELS</td>
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</tr>
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<td>NLAY</td>
<td>Total U-Wind tendency</td>
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<td>NLAY</td>
<td>Total Temperature tendency</td>
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<td>DQDT</td>
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<td>Total Specific Humidity tendency</td>
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<td>Frequency of Turbulence</td>
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<td>70</td>
<td>PBL</td>
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<td>Planetary Boundary Layer depth</td>
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<td>71</td>
<td>SWCLR</td>
<td>deg/day</td>
<td>NLAY</td>
<td>Net clearsky Shortwave heating rate for each level</td>
</tr>
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<td>Convective cloud mass flux</td>
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<tr>
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<td>UAVE</td>
<td>m/sec</td>
<td>NLAY</td>
<td>Time-averaged u – Wind</td>
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<td>NLAY</td>
<td>Time-averaged v – Wind</td>
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<td>NLAY</td>
<td>Time-averaged Temperature</td>
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<td>QAVE</td>
<td>g/g</td>
<td>NLAY</td>
<td>Time-averaged Specific Humidity</td>
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<td>DIABQ</td>
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<td>NLAY</td>
<td>Total Diabatic forcing on Specific Humidity</td>
</tr>
<tr>
<td>N</td>
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<td>U-Wind at 2 meters</td>
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<td>V-Wind at 2 meters</td>
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<td>VINTUQ</td>
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<td>CLDFRC</td>
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</tr>
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<td>QINT</td>
<td>gm/cm²</td>
<td>1</td>
<td>Precipitable water</td>
</tr>
<tr>
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<td>U2M</td>
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</tr>
<tr>
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<td>V2M</td>
<td>m/sec</td>
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<td>V-Wind at 2 meters</td>
</tr>
<tr>
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<td>deg</td>
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<td>Temperature at 2 meters</td>
</tr>
<tr>
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<td>Q2M</td>
<td>g/kg</td>
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<td>Specific Humidity at 2 meters</td>
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<td>m/sec</td>
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<td>V-Wind at 10 meters</td>
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<td>115</td>
<td>Q10M</td>
<td>g/kg</td>
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<td>Specific Humidity at 10 meters</td>
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<tr>
<td>116</td>
<td>DTRAIN</td>
<td>kg·m/sec</td>
<td>NLAY</td>
<td>Convective Cloud Detrainment</td>
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<td>117</td>
<td>QFILL</td>
<td>g/kg/day</td>
<td>NLAY</td>
<td>Filling of negative specific humidity</td>
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<tr>
<td>118</td>
<td>VINTQANA</td>
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<td>Vertically integrated analysis specific humidity increment</td>
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<tr>
<td>119</td>
<td>VINTQFIL</td>
<td>mm/day</td>
<td>1</td>
<td>Vertically integrated QFILL</td>
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</table>
8 Diagnostic Description

In this section we list and describe the diagnostic quantities available within the GEOS GCM. The diagnostics are listed in the order that they appear in the Diagnostic Menu, Section 7. In all cases, each diagnostic as currently archived on DAO production output datasets is time-averaged over its diagnostic output frequency:

\[ \text{DIAGNOSTIC} = \frac{1}{T T O T} \sum_{t=1}^{T T O T} \text{diag}(t) \]

where \( T T O T = \frac{N Q D I A G}{\Delta t} \), \( N Q D I A G \) is the output frequency of the diagnostic, and \( \Delta t \) is the timestep over which the diagnostic is updated. For further information on how to set the diagnostic output frequency \( N Q D I A G \), please see Part III, A User's Guide. Also, note that several Diagnostic Index locations are currently disabled due to changes which have occurred from previous GCM versions.

A word of caution: The Arakawa C-grid stencil used in the GEOS GCM employs a "stretched" coordinate near the poles due to conservation requirements. As a result, a well-defined vorticity is located at the poles, while the first mass point is located one \( \Delta \phi \) away from the poles. Since all Physical (diabatic) processes of the model are computed at mass points, all Physics diagnostics are only truly defined at latitudes \(-90 + \Delta \phi \) to \(90 - \Delta \phi \). While an attempt was made to "diagnose" some pole value diagnostics through interpolation over the poles, others have been left unfilled.

1) \textit{UFLUX} Surface Zonal Wind Stress on the Atmosphere (\textit{Newton/m}^2)

The zonal wind stress is the turbulent flux of zonal momentum from the surface. See section 3.3 for a description of the surface layer parameterization.

\[ \text{UFLUX} = -\rho C_D W_s u \quad \text{where:} \quad C_D = C_u^2 \]

where \( \rho \) = the atmospheric density at the surface, \( C_D \) is the surface drag coefficient, \( C_u \) is the dimensionless surface exchange coefficient for momentum (see diagnostic number 10), \( W_s \) is the magnitude of the surface layer wind, and \( u \) is the zonal wind in the lowest model layer.

2) \textit{VFLUX} Surface Meridional Wind Stress on the Atmosphere (\textit{Newton/m}^2)

The meridional wind stress is the turbulent flux of meridional momentum from the surface. See section 3.3 for a description of the surface layer parameterization.

\[ \text{VFLUX} = -\rho C_D W_s v \quad \text{where:} \quad C_D = C_u^2 \]
where $\rho$ = the atmospheric density at the surface, $C_D$ is the surface drag coefficient, $C_u$ is the dimensionless surface exchange coefficient for momentum (see diagnostic number 10), $W_s$ is the magnitude of the surface layer wind, and $v$ is the meridional wind in the lowest model layer.

3) **HFLUX** Surface Flux of Sensible Heat ($\text{Watts/m}^2$)

The turbulent flux of sensible heat from the surface to the atmosphere is a function of the gradient of virtual potential temperature and the eddy exchange coefficient:

$$
\text{HFLUX} = P^* \rho c_p C_H W_s (\theta_{\text{surface}} - \theta_{\text{NLAY}})
$$

where $\rho$ = the atmospheric density at the surface, $c_p$ is the specific heat of air, $C_H$ is the dimensionless surface heat transfer coefficient, $W_s$ is the magnitude of the surface layer wind, $C_u$ is the dimensionless surface exchange coefficient for momentum (see diagnostic number 10), $C_t$ is the dimensionless surface exchange coefficient for heat and moisture (see diagnostic number 9), and $\theta$ is the potential temperature at the surface and at the bottom $\sigma$-level.

4) **EFLUX** Surface Flux of Latent Heat ($\text{Watts/m}^2$)

The turbulent flux of latent heat from the surface to the atmosphere is a function of the gradient of moisture, the potential evapotranspiration fraction and the eddy exchange coefficient:

$$
\text{EFLUX} = \rho \beta L C_H W_s (q_{\text{surface}} - q_{\text{NLAY}})
$$

where $\rho$ = the atmospheric density at the surface, $\beta$ is the fraction of the potential evapotranspiration actually evaporated, $L$ is the latent heat of evaporation, $C_H$ is the dimensionless surface heat transfer coefficient, $W_s$ is the magnitude of the surface layer wind, $C_u$ is the dimensionless surface exchange coefficient for momentum (see diagnostic number 10), $C_t$ is the dimensionless surface exchange coefficient for heat and moisture (see diagnostic number 9), and $q_{\text{surface}}$ and $q_{\text{NLAY}}$ are the specific humidity at the surface and at the bottom $\sigma$-level, respectively.

5) **QICE** Heat Conduction Through Sea Ice ($\text{Watts/m}^2$)

Over sea ice there is an additional source of energy at the surface due to the heat conduction from the relatively warm ocean through the sea ice. The heat conduction through sea ice represents an additional energy source term for the ground temperature equation.

$$
\text{QICE} = \frac{C_{ti}}{H_i} (T_i - T_g)
$$
where $C_{ti}$ is the thermal conductivity of ice, $H_i$ is the ice thickness, assumed to be 3 m where sea ice is present, $T_i$ is 273 degrees Kelvin, and $T_g$ is the temperature of the sea ice.

NOTE: QICE is not available through model version 5.3, but is available in subsequent versions.

6) **RADLWG** Net upward Longwave Flux at the surface (Watts/m$^2$)

\[
\text{RADLWG} = F_{\text{LW}, \text{lay+1}}^\text{Net} = F_{\text{LW}, \text{lay+1}}^1 - F_{\text{LW}, \text{lay+1}}^1
\]

where lay+1 indicates the lowest model edge-level, or $p = p_{\text{surf}}$. $F_{\text{LW}}^1$ is the upward Longwave flux and $F_{\text{LW}}^1$ is the downward Longwave flux.

7) **RADSWG** Net downward shortwave Flux at the surface (Watts/m$^2$)

\[
\text{RADSWG} = F_{\text{SW}, \text{lay+1}}^\text{Net} = F_{\text{SW}, \text{lay+1}}^1 - F_{\text{SW}, \text{lay+1}}^1
\]

where lay+1 indicates the lowest model edge-level, or $p = p_{\text{surf}}$. $F_{\text{SW}}^1$ is the downward Shortwave flux and $F_{\text{SW}}^1$ is the upward Shortwave flux.

8) **RI** Richardson Number (dimensionless)

The non-dimensional stability indicator is the ratio of the buoyancy to the shear:

\[
\text{RI} = \frac{\frac{\partial}{\partial z} \frac{\partial \theta}{\partial z}}{(\frac{\partial u}{\partial z})^2 + (\frac{\partial v}{\partial z})^2} = \frac{C_p \frac{\partial \theta}{\partial z} \frac{\partial P}{\partial z}}{(\frac{\partial u}{\partial z})^2 + (\frac{\partial v}{\partial z})^2}
\]

where we used the hydrostatic equation:

\[
\frac{\partial \Phi}{\partial P} = C_p \theta_v
\]

Negative values indicate unstable buoyancy AND shear, small positive values (< 0.4) indicate dominantly unstable shear, and large positive values indicate dominantly stable stratification.
9) **CT Surface Exchange Coefficient for Temperature and Moisture** (*dimensionless*)

The surface exchange coefficient is obtained from the similarity functions for the stability dependant flux profile relationships:

\[
CT = -\frac{(w'\theta')}{u_* \Delta \theta} = -\frac{(w'q')}{u_* \Delta q} = \frac{k}{(\psi_h + \psi_g)}
\]

where \(\psi_h\) is the surface layer non-dimensional temperature change and \(\psi_g\) is the viscous sublayer non-dimensional temperature or moisture change:

\[
\psi_h = \int_{z_0}^{z} \frac{\phi_h}{\zeta} d\zeta \quad \text{and} \quad \psi_g = \frac{0.55(Pr^{2/3} - 0.2)}{\nu^{1/2}} (h_0 u_* - h_{0,ref} u_{*,ref})^{1/2}
\]

and: \(h_0 = 30z_0\) with a maximum value over land of 0.01

\(\phi_h\) is the similarity function of \(\zeta\), which expresses the stability dependance of the temperature and moisture gradients, specified differently for stable and unstable layers according to Helfand and Schubert, 1993. \(k\) is the Von Karman constant, \(\zeta\) is the non-dimensional stability parameter, \(Pr\) is the Prandtl number for air, \(\nu\) is the molecular viscosity, \(z_0\) is the surface roughness length, \(u_*\) is the surface stress velocity (see diagnostic number 67), and the subscript ref refers to a reference value.

10) **CU Surface Exchange Coefficient for Momentum** (*dimensionless*)

The surface exchange coefficient is obtained from the similarity functions for the stability dependant flux profile relationships:

\[
CU = \frac{u_*}{W_s} = \frac{k}{\psi_m}
\]

where \(\psi_m\) is the surface layer non-dimensional wind shear:

\[
\psi_m = \int_{z_0}^{z} \frac{\phi_m}{\zeta} d\zeta
\]

\(\phi_m\) is the similarity function of \(\zeta\), which expresses the stability dependance of the temperature and moisture gradients, specified differently for stable and unstable layers according to Helfand and Schubert, 1993. \(k\) is the Von Karman constant, \(\zeta\) is the non-dimensional stability parameter, \(u_*\) is the surface stress velocity (see diagnostic number 67), and \(W_s\) is the magnitude of the surface layer wind.

11) **ET Diffusivity Coefficient for Temperature and Moisture** (*m^2/sec*)
In the level 2.5 version of the Mellor-Yamada (1974) hierarchy, the turbulent heat or moisture flux for the atmosphere above the surface layer can be expressed as a turbulent diffusion coefficient \( K_h \) times the negative of the gradient of potential temperature or moisture. In the Helfand and Labraga (1988) adaptation of this closure, \( K_h \) takes the form:

\[
ET = K_h = -\varepsilon \frac{\partial \theta}{\partial z} = \begin{cases} 
q \ell S_H(G_M, G_H) & \text{for decaying turbulence} \\
\frac{q^2}{q_e} \ell S_H(G_{M_e}, G_{H_e}) & \text{for growing turbulence}
\end{cases}
\]

where \( q \) is the turbulent velocity, or \( \sqrt{2 \times \text{turbulent kinetic energy}} \), \( q_e \) is the turbulence velocity derived from the more simple level 2.0 model, which describes equilibrium turbulence, \( \ell \) is the master length scale related to the layer depth, \( S_H \) is a function of \( G_H \) and \( G_M \), the dimensionless buoyancy and wind shear parameters, respectively, or a function of \( G_{H_e} \) and \( G_{M_e} \), the equilibrium dimensionless buoyancy and wind shear parameters. Both \( G_H \) and \( G_M \), and their equilibrium values \( G_{H_e} \) and \( G_{M_e} \), are functions of the Richardson number.

For the detailed equations and derivations of the modified level 2.5 closure scheme, see Helfand and Labraga, 1988.

In the surface layer, \( ET \) is the exchange coefficient for heat and moisture, in units of \( m/sec \), given by:

\[
ET_{NLAY} = C_t \times u_* = C_H W_s
\]

where \( C_t \) is the dimensionless exchange coefficient for heat and moisture from the surface layer similarity functions (see diagnostic number 9), \( u_* \) is the surface friction velocity (see diagnostic number 67), \( C_H \) is the heat transfer coefficient, and \( W_s \) is the magnitude of the surface layer wind.

### 12) EU Diffusivity Coefficient for Momentum (\( m^2/sec \))

In the level 2.5 version of the Mellor-Yamada (1974) hierarchy, the turbulent heat momentum flux for the atmosphere above the surface layer can be expressed as a turbulent diffusion coefficient \( K_m \) times the negative of the gradient of the u-wind. In the Helfand and Labraga (1988) adaptation of this closure, \( K_m \) takes the form:

\[
EU = K_m = -\varepsilon \frac{u'w'}{\partial z} = \begin{cases} 
q \ell S_M(G_M, G_H) & \text{for decaying turbulence} \\
\frac{q^2}{q_e} \ell S_M(G_{M_e}, G_{H_e}) & \text{for growing turbulence}
\end{cases}
\]

where \( q \) is the turbulent velocity, or \( \sqrt{2 \times \text{turbulent kinetic energy}} \), \( q_e \) is the turbulence velocity derived from the more simple level 2.0 model, which describes equilibrium turbulence, \( \ell \) is the master length scale related to the layer depth, \( S_M \) is a function of \( G_H \) and \( G_M \), the dimensionless buoyancy and wind shear parameters, respectively, or a function of \( G_{H_e} \) and \( G_{M_e} \), the equilibrium dimensionless buoyancy and wind shear parameters. Both
$G_H$ and $G_M$, and their equilibrium values $G_{H*}$ and $G_{M*}$, are functions of the Richardson number.

For the detailed equations and derivations of the modified level 2.5 closure scheme, see Helfand and Labraga, 1988.

In the surface layer, $EU$ is the exchange coefficient for momentum, in units of $m/sec$, given by:

$$ EU_{NLAY} = C_u * u_* = C_D W_s $$

where $C_u$ is the dimensionless exchange coefficient for momentum from the surface layer similarity functions (see diagnostic number 10), $u_*$ is the surface friction velocity (see diagnostic number 67), $C_D$ is the surface drag coefficient, and $W_s$ is the magnitude of the surface layer wind.

13) **TURBU** Zonal U-Momentum changes due to Turbulence ($m/sec/day$)

The tendency of U-Momentum due to turbulence is written:

$$ TURBU = \frac{\partial u}{\partial t_{turb}} = \frac{\partial}{\partial z}(-u'w') = \frac{\partial}{\partial z}(K_m \frac{\partial u}{\partial z}) $$

The Helfand and Labraga level 2.5 scheme models the turbulent flux of u-momentum in terms of $K_m$, and the equation has the form of a diffusion equation.

14) **TURBV** Meridional V-Momentum changes due to Turbulence ($m/sec/day$)

The tendency of V-Momentum due to turbulence is written:

$$ TURBV = \frac{\partial v}{\partial t_{turb}} = \frac{\partial}{\partial z}(-v'w') = \frac{\partial}{\partial z}(K_m \frac{\partial v}{\partial z}) $$

The Helfand and Labraga level 2.5 scheme models the turbulent flux of v-momentum in terms of $K_m$, and the equation has the form of a diffusion equation.

15) **TURBT** Temperature changes due to Turbulence ($deg/day$)

The tendency of temperature due to turbulence is written:

$$ TURBT = \frac{\partial T}{\partial t} = P^\kappa \frac{\partial \theta}{\partial t_{turb}} = P^\kappa \frac{\partial}{\partial z}(-w' \theta') = P^\kappa \frac{\partial}{\partial z}(K_h \frac{\partial \theta}{\partial z}) $$

The Helfand and Labraga level 2.5 scheme models the turbulent flux of temperature in terms of $K_h$, and the equation has the form of a diffusion equation.
16) **TURBQ** Specific Humidity changes due to Turbulence (g/kg/day)

The tendency of specific humidity due to turbulence is written:

\[ TURBQ = \frac{\partial q}{\partial t_{turb}} = \frac{\partial}{\partial z}(-w'q') = \frac{\partial}{\partial z}(K_h \frac{\partial q}{\partial z}) \]

The Helfand and Labraga level 2.5 scheme models the turbulent flux of temperature in terms of \( K_h \), and the equation has the form of a diffusion equation.

17) **MOISTT** Temperature Changes Due to Moist Processes (deg/day)

\[ MOISTT = \frac{\partial T}{\partial t} |_c + \frac{\partial T}{\partial t} |_{ls} \]

where:

\[ \frac{\partial T}{\partial t} |_c = R \sum_i \left( \frac{\alpha m_B}{c_p} \Gamma_s \right)_i \quad \text{and} \quad \frac{\partial T}{\partial t} |_{ls} = \frac{L}{c_p} (q^* - q) \]

and

\[ \Gamma_s = g \eta \frac{\partial s}{\partial p} \]

The subscript \( c \) refers to convective processes, while the subscript \( ls \) refers to large scale precipitation processes, or supersaturation rain. The summation refers to contributions from each cloud type called by RAS. The dry static energy is given as \( s \), the convective cloud base mass flux is given as \( m_B \), and the cloud entrainment is given as \( \eta \), which are explicitly defined in Section 3.1.1, the description of the convective parameterization. The fractional adjustment, or relaxation parameter, for each cloud type is given as \( \alpha \), while \( R \) is the rain re-evaporation adjustment.

18) **MOISTQ** Specific Humidity Changes Due to Moist Processes (g/kg/day)

\[ MOISTQ = \frac{\partial q}{\partial t} |_c + \frac{\partial q}{\partial t} |_{ls} \]

where:

\[ \frac{\partial q}{\partial t} |_c = R \sum_i \left( \frac{\alpha_m}{L} (\Gamma_h - \Gamma_s) \right)_i \quad \text{and} \quad \frac{\partial q}{\partial t} |_{ls} = (q^* - q) \]

and

\[ \Gamma_s = g \eta \frac{\partial s}{\partial p} \quad \text{and} \quad \Gamma_h = g \eta \frac{\partial h}{\partial p} \]

The subscript \( c \) refers to convective processes, while the subscript \( ls \) refers to large scale precipitation processes, or supersaturation rain. The summation refers to contributions
from each cloud type called by RAS. The dry static energy is given as $s$, the moist static energy is given as $h$, the convective cloud base mass flux is given as $m_B$, and the cloud entrainment is given as $\eta$, which are explicitly defined in Section 3.1.1, the description of the convective parameterization. The fractional adjustment, or relaxation parameter, for each cloud type is given as $\alpha$, while $R$ is the rain re-evaporation adjustment.

19) **RADLW Heating Rate due to Longwave Radiation** *(deg/day)*

The net longwave heating rate is calculated as the vertical divergence of the net terrestrial radiative fluxes. Both the clear-sky and cloudy-sky longwave fluxes are computed within the longwave routine. The subroutine calculates the clear-sky flux, $F_{LW}^{\text{clearsky}}$, first. For a given cloud fraction, the clear line-of-sight probability $C(p, p')$ is computed from the current level pressure $p$ to the model top pressure, $p' = p_{\text{top}}$, and the model surface pressure, $p' = p_{\text{surf}}$, for the upward and downward radiative fluxes. (see Section 3.2.3). The cloudy-sky flux is then obtained as:

$$F_{LW} = C(p, p') \cdot F_{LW}^{\text{clearsky}},$$

Finally, the net longwave heating rate is calculated as the vertical divergence of the net terrestrial radiative fluxes:

$$\frac{\partial pc_p T}{\partial t} = -\frac{\partial}{\partial z} F_{LW}^{\text{NET}},$$

or

$$\text{RADLW} = \frac{g}{c_p \pi} \frac{\partial}{\partial \sigma} F_{LW}^{\text{NET}}.$$

where $g$ is the acceleration due to gravity, $c_p$ is the heat capacity of air at constant pressure, and

$$F_{LW}^{\text{NET}} = F_{LW}^1 - F_{LW}^1$$

20) **RADSW Heating Rate due to Shortwave Radiation** *(deg/day)*

The net Shortwave heating rate is calculated as the vertical divergence of the net solar radiative fluxes. The clear-sky and cloudy-sky shortwave fluxes are calculated separately. For the clear-sky case, the shortwave fluxes and heating rates are computed with both CLMO (maximum overlap cloud fraction) and CLRO (random overlap cloud fraction) set to zero (see Section 3.2.3). The shortwave routine is then called a second time, for the cloudy-sky case, with the true time-averaged cloud fractions CLMO and CLRO being used. In all cases, a normalized incident shortwave flux is used as input at the top of the atmosphere.
The heating rate due to Shortwave Radiation under cloudy skies is defined as:
\[
\frac{\partial p c_p T}{\partial t} = - \frac{\partial}{\partial z} F(\text{cloudy})_{SW}^{NET} \cdot \text{RADSWT},
\]
or
\[
\text{RADSW} = \frac{g}{c_p \pi} \frac{\partial}{\partial \sigma} F(\text{cloudy})_{SW}^{NET} \cdot \text{RADSWT}.
\]

where \( g \) is the acceleration due to gravity, \( c_p \) is the heat capacity of air at constant pressure, \( \text{RADSWT} \) is the true incident shortwave radiation at the top of the atmosphere (See Diagnostic #48), and

\[
F(\text{cloudy})_{SW}^{Net} = F(\text{cloudy})_{SW}^{1} - F(\text{cloudy})_{SW}^{1}
\]

21) **PREACC** Total (Large-scale + Convective) Accumulated Precipitation \((mm/day)\)

For a change in specific humidity due to moist processes, \( \Delta q_{moist} \), the vertical integral or total precipitable amount is given by:

\[
\text{PREACC} = \int_{\text{surf}}^{\text{top}} \rho \Delta q_{moist} dz = - \int_{\text{surf}}^{\text{top}} \Delta q_{moist} \frac{dp}{g} = \frac{\pi}{g} \int_{0}^{1} \Delta q_{moist} d\sigma
\]

A precipitation rate is defined as the vertically integrated moisture adjustment per Moist Processes time step, scaled to \( mm/day \).

22) **PRECON** Convective Precipitation \((mm/day)\)

For a change in specific humidity due to sub-grid scale cumulus convective processes, \( \Delta q_{cum} \), the vertical integral or total precipitable amount is given by:

\[
\text{PRECON} = \int_{\text{surf}}^{\text{top}} \rho \Delta q_{cum} dz = - \int_{\text{surf}}^{\text{top}} \Delta q_{cum} \frac{dp}{g} = \frac{\pi}{g} \int_{0}^{1} \Delta q_{cum} d\sigma
\]

A precipitation rate is defined as the vertically integrated moisture adjustment per Moist Processes time step, scaled to \( mm/day \).
23) **TUFLUX** Turbulent Flux of U-Momentum \((Newton/m^2)\)

The turbulent flux of u-momentum is calculated for *diagnostic purposes only* from the eddy coefficient for momentum:

\[
TUFLUX = \rho \langle w'w' \rangle = \rho \left( -K_u \frac{\partial U}{\partial z} \right)
\]

where \(\rho\) is the air density, and \(K_u\) is the eddy coefficient.

24) **TVFLUX** Turbulent Flux of V-Momentum \((Newton/m^2)\)

The turbulent flux of v-momentum is calculated for *diagnostic purposes only* from the eddy coefficient for momentum:

\[
TVFLUX = \rho \langle v'w' \rangle = \rho \left( -K_v \frac{\partial V}{\partial z} \right)
\]

where \(\rho\) is the air density, and \(K_v\) is the eddy coefficient.

25) **TTFLUX** Turbulent Flux of Sensible Heat \((Watts/m^2)\)

The turbulent flux of sensible heat is calculated for *diagnostic purposes only* from the eddy coefficient for heat and moisture:

\[
TTFLUX = c_p \rho P^s \langle w'\theta' \rangle = c_p \rho \rho P^s \left( -K_h \frac{\partial \theta}{\partial z} \right)
\]

where \(\rho\) is the air density, and \(K_h\) is the eddy coefficient.

26) **TQFLUX** Turbulent Flux of Latent Heat \((Watts/m^2)\)

The turbulent flux of latent heat is calculated for *diagnostic purposes only* from the eddy coefficient for heat and moisture:

\[
TQFLUX = L \rho \langle w'q' \rangle = L \rho \left( -K_h \frac{\partial q}{\partial z} \right)
\]

where \(\rho\) is the air density, and \(K_h\) is the eddy coefficient.
27) **CN Neutral Drag Coefficient** *(dimensionless)*

The drag coefficient for momentum obtained by assuming a neutrally stable surface layer:

\[ CN = \frac{k}{\ln \left( \frac{h}{z_0} \right)} \]

where \( k \) is the Von Karman constant, \( h \) is the height of the surface layer, and \( z_0 \) is the surface roughness.

NOTE: CN is not available through model version 5.3, but is available in subsequent versions.

28) **WINDS Surface Wind Speed** *(meter/sec)*

The surface wind speed is calculated for the last internal turbulence time step:

\[ \text{WINDS} = \sqrt{u_{NLAY}^2 + v_{NLAY}^2} \]

where the subscript \( NLAY \) refers to the lowest model \( \sigma \)-level.

29) **DTSRF Air/Surface Virtual Temperature Difference** *(deg K)*

The air/surface virtual temperature difference measures the stability of the surface layer:

\[ \text{DTSRF} = (\theta_{NLAY+1} - \theta_{NLAY})P_{surf}^\kappa \]

where

\[ \theta_{NLAY+1} = \frac{T_g}{P_{surf}^\kappa} \left( 1 + 0.609 q_{NLAY+1} \right) \quad \text{and} \quad q_{NLAY+1} = q_{NLAY} + \beta (q^*(T_g, P_s) - q_{NLAY}) \]

\( \beta \) is the surface potential evapotranspiration coefficient \((\beta = 1 \text{ over oceans})\), \( q^*(T_g, P_s) \) is the saturation specific humidity at the ground temperature and surface pressure, level \( NLAY \) refers to the lowest model level and level \( NLAY + 1 \) refers to the surface.

30) **TG Ground Temperature** *(deg K)*

The ground temperature equation is solved as part of the turbulence package using a backward implicit time differencing scheme:

\[ TG \text{ is obtained from: } C_g \frac{\partial T_g}{\partial t} = R_{sw} - R_{lw} + Q_{ice} - H - LE \]
where $R_{sw}$ is the net surface downward shortwave radiative flux, $R_{lw}$ is the net surface upward longwave radiative flux, $Q_{ice}$ is the heat conduction through sea ice, $H$ is the upward sensible heat flux, $LE$ is the upward latent heat flux, and $C_g$ is the total heat capacity of the ground. $C_g$ is obtained by solving a heat diffusion equation for the penetration of the diurnal cycle into the ground (Blackadar, 1977), and is given by:

$$C_g = \sqrt{\frac{\lambda C_s}{2\omega}} = \sqrt{(0.386 + 0.536W + 0.15W^2)2x10^{-3}} \cdot \frac{86400}{2\pi}.$$  

Here, the thermal conductivity, $\lambda$, is equal to $2x10^{-3} \frac{\text{ly cm}}{\text{sec} \cdot \text{K}}$, the angular velocity of the earth, $\omega$, is written as $86400 \text{ sec/day}$ divided by $2\pi \text{ radians/day}$, and the expression for $C_s$, the heat capacity per unit volume at the surface, is a function of the ground wetness, $W$.

31) **TS Surface Temperature (deg K)**

The surface temperature estimate is made by assuming that the model's lowest layer is well-mixed, and therefore that $\theta$ is constant in that layer. The surface temperature is therefore:

$$TS = \theta_{NLAY} P_{surf}$$

32) **DTG Surface Temperature Adjustment (deg K)**

The change in surface temperature from one turbulence time step to the next, solved using the Ground Temperature Equation (see diagnostic number 30) is calculated:

$$DTG = T_g^n - T_g^{n-1}$$

where superscript $n$ refers to the new, updated time level, and the superscript $n - 1$ refers to the value at the previous turbulence time level.

33) **QG Ground Specific Humidity (g/kg)**

The ground specific humidity is obtained by interpolating between the specific humidity at the lowest model level and the specific humidity of a saturated ground. The interpolation is performed using the potential evapotranspiration function:

$$QG = q_{NLAY+1} = q_{NLAY} + \beta(q^*(T_g, P_s) - q_{NLAY})$$

where $\beta$ is the surface potential evapotranspiration coefficient ($\beta = 1$ over oceans), and $q^*(T_g, P_s)$ is the saturation specific humidity at the ground temperature and surface pres-
34) **QS** Saturation Surface Specific Humidity \((g/kg)\)

The surface saturation specific humidity is the saturation specific humidity at the ground temperature and surface pressure:

\[
QS = q^*(T_g, P_s)
\]

35) **TGRLW** Instantaneous ground temperature used as input to the Longwave radiation subroutine (deg)

\[
TGRLW = T_g(\lambda, \phi, n)
\]

where \(T_g\) is the model ground temperature at the current time step \(n\).

36) **ST4** Upward Longwave flux at the surface \((Watts/m^2)\)

\[
ST4 = \sigma T^4
\]

where \(\sigma\) is the Stefan-Boltzmann constant and \(T\) is the temperature.

37) **OLR** Net upward Longwave flux at \(p = p_{top}\) \((Watts/m^2)\)

\[
OLR = \frac{F^NET_{LW,top}}{F_{LW,top}}
\]

where \(top\) indicates the top of the first model layer. In the GEOS-1 GCM, \(p_{top} = 10\) mb.

38) **OLRCLR** Net upward clearsky Longwave flux at \(p = p_{top}\) \((Watts/m^2)\)

\[
OLRCLR = F(clearsky)^{NET}_{LW,top}
\]

where \(top\) indicates the top of the first model layer. In the GEOS-1 GCM, \(p_{top} = 10\) mb.

39) **LWGCLR** Net upward clearsky Longwave flux at the surface \((Watts/m^2)\)

\[
LWGCLR = F(clearsky)^{NET}_{LW,nlay+1}
\]

\[
= F(clearsky)^{1}_{LW,nlay+1} - F(clearsky)^{1}_{LW,nlay+1}
\]
where \( n_{lay+1} \) indicates the lowest model edge-level, or \( p = p_{surf} \). \( F(\text{clearsky})_{LW}^{1} \) is the upward clearsky Longwave flux and the \( F(\text{clearsky})_{LW}^{-1} \) is the downward clearsky Longwave flux.

40) \textbf{LWCLR} Heating Rate due to Clearsky Longwave Radiation (deg/day)

The net longwave heating rate is calculated as the vertical divergence of the net terrestrial radiative fluxes. Both the clear-sky and cloudy-sky longwave fluxes are computed within the longwave routine. The subroutine calculates the clear-sky flux, \( F_{LW}^{\text{clearsky}} \), first. For a given cloud fraction, the clear line-of-sight probability \( C(p, p') \) is computed from the current level pressure \( p \) to the model top pressure, \( p' = p_{top} \), and the model surface pressure, \( p' = p_{surf} \), for the upward and downward radiative fluxes. (see Section 3.2.3). The cloudy-sky flux is then obtained as:

\[
F_{LW} = C(p, p') \cdot F_{LW}^{\text{clearsky}},
\]

Thus, \textbf{LWCLR} is defined as the net longwave heating rate due to the vertical divergence of the clear-sky longwave radiative flux:

\[
\frac{\partial \rho c_{p} T}{\partial t}_{\text{clearsky}} = -\frac{\partial}{\partial z} F(\text{clearsky})_{LW}^{\text{NET}},
\]

or

\[
\text{LWCLR} = g \frac{\partial}{\partial \sigma} F(\text{clearsky})_{LW}^{\text{NET}}.
\]

where \( g \) is the acceleration due to gravity, \( c_{p} \) is the heat capacity of air at constant pressure, and

\[
F(\text{clearsky})_{LW}^{\text{Net}} = F(\text{clearsky})_{LW}^{1} - F(\text{clearsky})_{LW}^{-1}
\]

41) \textbf{TLW} Instantaneous temperature used as input to the Longwave radiation subroutine (deg)

\[
\text{TLW} = T(\lambda, \phi, \sigma, n)
\]

where \( T \) is the model temperature at the current time step \( n \).

42) \textbf{SHLW} Instantaneous specific humidity used as input to the Longwave radiation subroutine (kg/kg)

\[
\text{SHLW} = q(\lambda, \phi, \sigma, n)
\]
where $q$ is the model specific humidity at the current time step $n$.

43) **OZLW** Instantaneous ozone used as input to the Longwave radiation subroutine (kg/kg)

$$OZLW = OZ(\lambda, \phi, \sigma, n)$$

where $OZ$ is the interpolated ozone data set from the climatological monthly mean zonally averaged ozone data set.

44) **CLMOLW** Maximum Overlap cloud fraction used in LW Radiation (0 – 1)

CLMOLW is the time-averaged maximum overlap cloud fraction that has been filled by the Relaxed Arakawa/Schubert Convection scheme and will be used in the Longwave Radiation algorithm. These are convective clouds whose radiative characteristics are assumed to be correlated in the vertical. For a complete description of cloud/radiative interactions, see Section 3.2.3.

$$CLMOLW = CLMORAs, LW(\lambda, \phi, \sigma)$$

Note, in Version 1 of the GEOS GCM, shortwave and longwave cloud fields are computed identically, ie.

$$CLMOLW = CLMOSW$$

45) **CLROLW** Random Overlap cloud fraction used in LW Radiation (0 – 1)

CLROLW is the time-averaged random overlap cloud fraction that has been filled by the Relaxed Arakawa/Schubert and Large-scale Convection schemes and will be used in the Longwave Radiation algorithm. These are convective and large-scale clouds whose radiative characteristics are not assumed to be correlated in the vertical. For a complete description of cloud/radiative interactions, see Section 3.2.3.

$$CLROLW = CLRORAS, LargeScale, LW(\lambda, \phi, \sigma)$$

Note, in Version 1 of the GEOS GCM, shortwave and longwave cloud fields are computed identically, ie.

$$CLROLW = CLROSW$$

46) **CLMOSW** Maximum Overlap cloud fraction used in SW Radiation (0 – 1)
**CLMOSW** is the time-averaged maximum overlap cloud fraction that has been filled by the Relaxed Arakawa/Schubert Convection scheme and will be used in the Shortwave Radiation algorithm. These are convective clouds whose radiative characteristics are assumed to be correlated in the vertical. For a complete description of cloud/radiative interactions, see Section 3.2.3.

\[
\text{CLMOSW} = CLO_{RAS,SW}(\lambda, \phi, \sigma)
\]

Note, in Version 1 of the GEOS GCM, shortwave and longwave cloud fields are computed identically, ie.

\[
\text{CLMOSW} = \text{CLMOLW}
\]

47) **CLROSW** Random Overlap cloud fraction used in SW Radiation \((0 - 1)\)

**CLROSW** is the time-averaged random overlap cloud fraction that has been filled by the Relaxed Arakawa/Schubert and Large-scale Convection schemes and will be used in the Shortwave Radiation algorithm. These are convective and large-scale clouds whose radiative characteristics are not assumed to be correlated in the vertical. For a complete description of cloud/radiative interactions, see Section 3.2.3.

\[
\text{CLROSW} = CLO_{RAS,\text{LargeScale,SW}}(\lambda, \phi, \sigma)
\]

Note, in Version 1 of the GEOS GCM, shortwave and longwave cloud fields are computed identically, ie.

\[
\text{CLROSW} = \text{CLROLW}
\]

48) **RADSWT** Incident Shortwave radiation at the top of the atmosphere \((Watts/m^2)\)

\[
\text{RADSWT} = \frac{S_0}{R_a^2} \cdot \cos \phi_z
\]

where \(S_0\), is the extra-terrestrial solar constant, \(R_a\) is the earth-sun distance in Astronomical Units, and \(\cos \phi_z\) is the cosine of the zenith angle. It should be noted that **RADSWT**, as well as **OSR** and **OSRCLR**, are calculated at the top of the atmosphere (\(p=0\) mb). However, the **OLR** and **OLRCLR** diagnostics are currently calculated at \(p = p_{top}\) (10 mb for the 20-level GEOS-1 GCM).

49) Disabled
50) Disabled

51) Disabled

52) EVAP Surface Evaporation \( (\text{mm/day}) \)

The surface evaporation is a function of the gradient of moisture, the potential evapotranspiration fraction and the eddy exchange coefficient:

\[
\text{EVAP} = \rho \beta K_h (q_{\text{surface}} - q_{NLAY})
\]

where \( \rho \) = the atmospheric density at the surface, \( \beta \) is the fraction of the potential evapotranspiration actually evaporated (\( \beta = 1 \) over oceans), \( K_h \) is the turbulent eddy exchange coefficient for heat and moisture at the surface in \( \text{m/sec} \) and \( q_{\text{surface}} \) and \( q_{NLAY} \) are the specific humidity at the surface (see diagnostic number 34) and at the bottom \( \sigma \)-level, respectively.

53) DPDT Total Surface Pressure Tendency \( (\text{mb/day}) \)

DPDT is the total time-tendency of the surface pressure due to Hydrodynamic and Analysis forcing. There are no surface pressure changes due to physical diabatic processes, such as changes in total precipitable water amounts.

\[
\text{DPDT} = \frac{\partial \pi}{\partial t_{\text{Dynamics}}} + \frac{\partial \pi}{\partial t_{\text{Analysis}}},
\]

where \( \pi = p_{\text{surf}} - p_{\text{top}} \). In forecast simulations \( \frac{\partial \pi}{\partial t_{\text{Analysis}}} = 0 \).

54) ANALU Analysis Zonal U-Wind Tendency \( (\text{m/sec/sec}) \)

ANALU is the Zonal U-Wind tendency created during the assimilation cycle which is used in the Incremental Analysis Updating (IAU) procedure.

\[
\text{ANALU} = \left( u^A(\lambda, \phi, \sigma, t) - u^{FG}(\lambda, \phi, \sigma, t) \right)/\Delta t_{\text{Analysis}}
\]

where \( u^A \) corresponds to the u-wind after the analysis, \( u^{FG} \) corresponds to the u-wind first guess, and \( \Delta t_{\text{Analysis}} \) is the time-step between analyses (6 hours). Note, ANALU is stored on the staggered C-grid.

55) ANALV Analysis Meridional V-Wind Tendency \( (\text{m/sec/sec}) \)
ANALV is the Meridional V-Wind tendency created during the assimilation cycle which is used in the Incremental Analysis Updating (IAU) procedure.

\[
\text{ANALV} = \left( v^A(\lambda, \phi, \sigma, t) - v^{FG}(\lambda, \phi, \sigma, t) \right) / \Delta t_{\text{Analysis}}
\]

where \( v^A \) corresponds to the v-wind after the analysis, \( v^{FG} \) corresponds to the v-wind first guess, and \( \Delta t_{\text{Analysis}} \) is the time-step between analyses (6 hours). Note, ANALV is stored on the staggered C-grid.

56) **ANALT Analysis Temperature Tendency** (\( mb \text{deg} / mb^\sigma / \text{sec} \))

ANALT is the pressure weighted Potential Temperature tendency created during the assimilation cycle which is used in the Incremental Analysis Updating (IAU) procedure.

\[
\text{ANALT} = \left( (\pi \theta)^A(\lambda, \phi, \sigma, t) - (\pi \theta)^{FG}(\lambda, \phi, \sigma, t) \right) / \Delta t_{\text{Analysis}}
\]

where \((\pi \theta)^A\) corresponds to the pressure weighted Potential Temperature after the analysis, \((\pi \theta)^{FG}\) corresponds to the pressure weighted Potential Temperature first guess, \(\theta = T / p^\sigma\), and \(\Delta t_{\text{Analysis}}\) is the time-step between analyses (6 hours).

57) **ANALQ Analysis Temperature Tendency** (\( mb \cdot g / g / \text{sec} \))

ANALQ is the pressure weighted Specific Humidity tendency created during the assimilation cycle which is used in the Incremental Analysis Updating (IAU) procedure.

\[
\text{ANALQ} = \left( (\pi q)^A(\lambda, \phi, \sigma, t) - (\pi q)^{FG}(\lambda, \phi, \sigma, t) \right) / \Delta t_{\text{Analysis}}
\]

where \((\pi q)^A\) corresponds to the pressure weighted Specific Humidity after the analysis, \((\pi q)^{FG}\) corresponds to the pressure weighted Specific Humidity first guess, and \(\Delta t_{\text{Analysis}}\) is the time-step between analyses (6 hours).

58) **OMEGA Pressure Coordinate Vertical Velocity** (\( mb / \text{day} \))

OMEGA, which is the change in pressure in time following the parcel, is diagnostically obtained by integrating the GCM’s continuity equation. We have

\[
\text{OMEGA} \equiv \frac{dp}{dt} = (\pi \dot{\sigma} + \dot{\pi} \sigma)
\]

since \( p = \pi \sigma + p_{Top} \). Now,

\[
\begin{align*}
\dot{\pi} &= \frac{\partial \pi}{\partial t} + \mathbf{V} \cdot \nabla \pi \\
\frac{\partial \pi}{\partial t} &= -\nabla \cdot (\pi \mathbf{V}) - \frac{\partial \pi \dot{\sigma}}{\partial \sigma}
\end{align*}
\]
Integrating the continuity equation from the top of the model to a given model edge-level, \( \sigma_e \), and using the boundary condition \( \dot{\sigma} \big|_{\sigma=0} = 0 \), we have

\[
\pi \dot{\sigma}_e = - \int_0^{\sigma_e} \nabla \cdot (\pi \mathbf{V}) d\sigma - \sigma_e \frac{\partial \pi}{\partial t}
\]

Combining the above equations, we have

\[
\Omega = - \int_0^{\sigma_e} \nabla \cdot (\pi \mathbf{V}) d\sigma + \sigma_e \mathbf{V} \cdot \nabla \pi
\]

Note that in this formulation, \( \Omega \) is computed at GCM model edge-levels. \( \Omega \) at the model top edge is taken to be 0, thus the first computed value of \( \Omega \) is at sigma edge 2, or the bottom edge of the top model sigma layer.

59) \( \text{DUDT} \) Total Zonal U-Wind Tendency \( (m/sec/day) \)

\( \text{DUDT} \) is the total time-tendency of the Zonal U-Wind due to Hydrodynamic, Diabatic, and Analysis forcing.

\[
\text{DUDT} = \frac{\partial u}{\partial t_{\text{Dynamics}}} + \frac{\partial u}{\partial t_{\text{Moist}}} + \frac{\partial u}{\partial t_{\text{Turbulence}}} + \frac{\partial u}{\partial t_{\text{Analysis}}}
\]

Note, in Version 1 of the GEOS GCM there are no u-wind tendencies due to cumulus friction, (ie. \( \frac{\partial u}{\partial t_{\text{Moist}}} = 0 \)). Also, in forecast simulations \( \frac{\partial u}{\partial t_{\text{Analysis}}} = 0 \).

60) \( \text{DVDT} \) Total Zonal V-Wind Tendency \( (m/sec/day) \)

\( \text{DVDT} \) is the total time-tendency of the Meridional V-Wind due to Hydrodynamic, Diabatic, and Analysis forcing.

\[
\text{DVDT} = \frac{\partial v}{\partial t_{\text{Dynamics}}} + \frac{\partial v}{\partial t_{\text{Moist}}} + \frac{\partial v}{\partial t_{\text{Turbulence}}} + \frac{\partial v}{\partial t_{\text{Analysis}}}
\]

Note, in Version 1 of the GEOS GCM there are no v-wind tendencies due to cumulus friction, (ie. \( \frac{\partial v}{\partial t_{\text{Moist}}} = 0 \)). Also, in forecast simulations \( \frac{\partial v}{\partial t_{\text{Analysis}}} = 0 \).

61) \( \text{DTDT} \) Total Temperature Tendency \( (\text{deg/day}) \)

\( \text{DTDT} \) is the total time-tendency of Temperature due to Hydrodynamic, Diabatic, and Analysis forcing.

\[
\text{DTDT} = \frac{\partial T}{\partial t_{\text{Dynamics}}} + \frac{\partial T}{\partial t_{\text{MoistProcesses}}} + \frac{\partial T}{\partial t_{\text{ShortwaveRadiation}}} + \frac{\partial T}{\partial t_{\text{LongwaveRadiation}}} + \frac{\partial T}{\partial t_{\text{Turbulence}}} + \frac{\partial T}{\partial t_{\text{Analysis}}}
\]
Note, this diagnostic is obtained through differentiation by parts and re-arranging the model prognostic equations. Since the thermodynamic equation is written in the mass-weighted flux form, we have

$$\frac{\partial T}{\partial t} = \frac{p^e}{\pi} \left( \frac{\partial \pi \theta}{\partial t} + \theta \frac{\partial \pi}{\partial t} \left[ \frac{\sigma \pi \kappa}{p} - 1 \right] \right)$$

where $\theta = T/p^e$. Also, in forecast simulations, $\frac{\partial T}{\partial t \ Analysis} = 0$.

62) **DQDT Total Specific Humidity Tendency** ($g/kg/day$)

**DQDT** is the total time-tendency of Specific Humidity due to Hydrodynamic, Diabatic, and Analysis forcing.

$$DQDT = \frac{\partial q}{\partial t \ Dynamics} + \frac{\partial q}{\partial t \ MoistProcesses} + \frac{\partial q}{\partial t \ Turbulence} + \frac{\partial q}{\partial t \ Analysis}$$

Note, this diagnostic is obtained through differentiation by parts and re-arranging the model prognostic equations. Since the moisture equation is written in the mass-weighted flux form, we have

$$\frac{\partial q}{\partial t} = \left( \frac{\partial \pi q}{\partial t} - q \frac{\partial \pi}{\partial t} \right) / \pi$$

Also, in forecast simulations, $\frac{\partial q}{\partial t \ Analysis} = 0$.

63) **VORT Relative Vorticity** ($sec^{-1}$)

The relative vorticity used in the formulation of the potential energy and enstrophy conserving scheme described in Section 2 is stored as the diagnostic **VORT**. The relative vorticity is given as:

$$VORT = \frac{1}{a \ cos \ \phi} \left[ \frac{\partial v}{\partial \lambda} - \frac{\partial (u \ cos \ \phi)}{\partial \phi} \right]$$

This diagnostic is stored on the Arakawa C-grid used in the GEOS GCM at the “vorticity-point” locations, ie., centered between the u-points and v-points, including values well defined at both poles.

64) Disabled

65) Disabled
67) **USTAR** Surface-Stress Velocity \((m/sec)\)

The surface stress velocity, or the friction velocity, is the wind speed at the surface layer top impeded by the surface drag:

\[
USTAR = C_u W_s \quad \text{where} \quad C_u = \frac{k}{\psi_m}
\]

\(C_u\) is the non-dimensional surface drag coefficient (see diagnostic number 10), and \(W_s\) is the surface wind speed (see diagnostic number 28).

68) **Z0** Surface Roughness Length \((m)\)

Over the land surface, the surface roughness length is interpolated to the local time from the monthly mean data of Dorman and Sellers (1989). Over the ocean, the roughness length is a function of the surface-stress velocity, \(u_*\).

\[
Z0 = c_1 u_*^3 + c_2 u_*^2 + c_3 u_* + c_4 + \frac{c_5}{u_*}
\]

where the constants are chosen to interpolate between the reciprocal relation of Kondo (1975) for weak winds, and the piecewise linear relation of Large and Pond (1981) for moderate to large winds.

69) **FRQTRB** Frequency of Turbulence \((0 - 1)\)

The fraction of time when turbulence is present is defined as the fraction of time when the turbulent kinetic energy exceeds some minimum value, defined here to be \(0.005 \text{ m}^2/\text{sec}^2\). When this criterion is met, a counter is incremented. The fraction over the averaging interval is reported.

70) **PBL** Planetary Boundary Layer Depth \((mb)\)

The depth of the PBL is defined by the turbulence parameterization to be the depth at which the turbulent kinetic energy reduces to ten percent of its surface value.

\[
PBL = P_{PBL} - P_{surface}
\]

where \(P_{PBL}\) is the pressure in \(mb\) at which the turbulent kinetic energy reaches one tenth
of its surface value, and \( P_s \) is the surface pressure.

71) **SWCLR** Clear sky Heating Rate due to Shortwave Radiation (deg/day)

The net Shortwave heating rate is calculated as the vertical divergence of the net solar radiative fluxes. The clear-sky and cloudy-sky shortwave fluxes are calculated separately. For the clear-sky case, the shortwave fluxes and heating rates are computed with both CLMO (maximum overlap cloud fraction) and CLRO (random overlap cloud fraction) set to zero (see Section 3.2.3). The shortwave routine is then called a second time, for the cloudy-sky case, with the true time-averaged cloud fractions CLMO and CLRO being used. In all cases, a normalized incident shortwave flux is used as input at the top of the atmosphere.

The heating rate due to Shortwave Radiation under clear skies is defined as:

\[
\frac{\partial p c_p T}{\partial t} = -\frac{\partial}{\partial z} F(\text{clear})_{SW}^{\text{NET}} \cdot \text{RADSWT},
\]

or

\[
\text{SWCLR} = \frac{g}{c_p \pi} \frac{\partial}{\partial \sigma} F(\text{clear})_{SW}^{\text{NET}} \cdot \text{RADSWT}.
\]

where \( g \) is the acceleration due to gravity, \( c_p \) is the heat capacity of air at constant pressure, RADSWT is the true incident shortwave radiation at the top of the atmosphere (See Diagnostic #48), and

\[
F(\text{clear})_{SW}^{\text{Net}} = F(\text{clear})_{SW}^{1} - F(\text{clear})_{SW}^{1/2}
\]

72) **OSR** Net upward Shortwave flux at the top of the model (Watts/m²)

\[
\text{OSR} = F_{SW,\text{top}}^{\text{NET}}
\]

where top indicates the top of the first model layer used in the shortwave radiation routine. In the GEOS-1 GCM, \( p_{SW,\text{top}} = 0 \) mb.

73) **OSRCLR** Net upward clearsky Shortwave flux at the top of the model (Watts/m²)

\[
\text{OSRCLR} = F(\text{clearsky})_{SW,\text{top}}^{\text{NET}}
\]

where top indicates the top of the first model layer used in the shortwave radiation routine. In the GEOS-1 GCM, \( p_{SW,\text{top}} = 0 \) mb.
74) **CLDMAS** Convective Cloud Mass Flux \((kg/m/sec)\)

The cumulative cloud mass flux per unit time, from all convective clouds generated during the output interval, is written:

\[
CLDMAS = \eta m_B
\]

where \(\eta\) is the entrainment, normalized by the cloud base mass flux, and \(m_B\) is the cloud base mass flux. \(m_B\) and \(\eta\) are defined explicitly in Section 3.1.1, the description of the convective parameterization.

75) **UAVE** Time-Averaged Zonal U-Wind \((m/sec)\)

The diagnostic **UAVE** is simply the time-averaged Zonal U-Wind over the NUAVE output frequency. This is contrasted to the instantaneous Zonal U-Wind which is archived on the Prognostic Output data stream.

\[
UAVE = u(\lambda, \phi, \sigma, t)
\]

Note, **UAVE** is computed and stored on the staggered C-grid.

76) **VAVE** Time-Averaged Meridional V-Wind \((m/sec)\)

The diagnostic **VAVE** is simply the time-averaged Meridional V-Wind over the NVAVE output frequency. This is contrasted to the instantaneous Meridional V-Wind which is archived on the Prognostic Output data stream.

\[
VAVE = v(\lambda, \phi, \sigma, t)
\]

Note, **VAVE** is computed and stored on the staggered C-grid.

77) **TAVE** Time-Averaged Temperature \((Kelvin)\)

The diagnostic **TAVE** is simply the time-averaged Temperature over the NTAVE output frequency. This is contrasted to the instantaneous Temperature which is archived on the Prognostic Output data stream.

\[
TAVE = T(\lambda, \phi, \sigma, t)
\]

78) **QAVE** Time-Averaged Specific Humidity \((g/kg)\)
The diagnostic QAVE is simply the time-averaged Specific Humidity over the NQAVE output frequency. This is contrasted to the instantaneous Specific Humidity which is archived on the Prognostic Output data stream.

\[ \text{QAVE} = q(\lambda, \phi, \sigma, t) \]

79) Disabled

80) \textbf{PAVE} Time-Averaged Surface Pressure - PTOP (mb)

The diagnostic PAVE is simply the time-averaged Surface Pressure - PTOP over the NPAVE output frequency. This is contrasted to the instantaneous Surface Pressure - PTOP which is archived on the Prognostic Output data stream.

\[ \text{PAVE} = \pi(\lambda, \phi, \sigma, t) \]
\[ = p_s(\lambda, \phi, \sigma, t) - p_T \]

81) \textbf{QQAVE} Time-Averaged Turbulent Kinetic Energy \((m/sec)^2\)

The diagnostic QQAVE is simply the time-averaged prognostic Turbulent Kinetic Energy produced by the GEOS-1 GCM Turbulence parameterization over the NQQAVE output frequency. This is contrasted to the instantaneous Turbulent Kinetic Energy which is archived on the Prognostic Output data stream.

\[ \text{QQAVE} = qq(\lambda, \phi, \sigma, t) \]

Note, QQAVE is computed and stored at the “mass-point” locations on the staggered C-grid.

82) \textbf{SWGCLR} Net downward clearsky Shortwave flux at the surface \((Watts/m^2)\)

\[ \text{SWGCLR} = F(\text{clearsky})^{\text{Net}}_{SW, nlay+1} \]
\[ = F(\text{clearsky})^{\uparrow}_{SW, nlay+1} - F(\text{clearsky})^{\downarrow}_{SW, nlay+1} \]

where nlay+1 indicates the lowest model edge-level, or \(p = p_{surf}\). \(F(\text{clearsky})SW^{\uparrow}\) is the
downward clearsky Shortwave flux and $F(\text{clearsky})_{SW}^T$ is the upward clearsky Shortwave flux.

83) **ANALP** Analysis Surface Pressure Tendency (mb/sec)

**ANALP** is the surface pressure tendency created during the assimilation cycle which is used in the Incremental Analysis Updating (IAU) procedure.

$$\text{ANALP} = \left( \pi^A(\lambda, \phi, \sigma, t) - \pi^{FG}(\lambda, \phi, \sigma, t) \right) / \Delta t_{\text{Analysis}}$$

where $\pi^A$ corresponds to the surface pressure after the analysis, $\pi^{FG}$ corresponds to the surface pressure first guess, and $\Delta t_{\text{Analysis}}$ is the time-step between analyses (6 hours).

84) **SDIAG1** User-Defined Surface Diagnostic-1

The GEOS GCM provides Users with a built-in mechanism for archiving user-defined diagnostics. The generic diagnostic array QDIAG located in COMMON /DIAG/, and the associated diagnostic counters and pointers located in COMMON /DIAGP/, must be accessible in order to use the user-defined diagnostics (see Section 6). A convenient method for incorporating all necessary COMMON files is to include the GEOS GCM vstate.com file in the routine which employs the user-defined diagnostics.

In addition to enabling the user-defined diagnostic (ie., CALL SETDIAG(84)), the User must fill the QDIAG array with the desired quantity within the User's application program or within modified GEOS GCM subroutines, as well as increment the diagnostic counter at the time when the diagnostic is updated. The QDIAG location index for **SDIAG1** and its corresponding counter is automatically defined as ISDIAG1 and NSDIAG1, respectively, after the diagnostic has been enabled. The syntax for its use is given by

\begin{verbatim}
do j=1,jm
do i=1,im
   qdiag(i,j,ISDIAG1) = qdiag(i,j,ISDIAG1) + ... 
enddo
enddo

NSDIAG1 = NSDIAG1 + 1
\end{verbatim}

The diagnostics defined in this manner will automatically be archived by the production output programs (see Section 11).

85) **SDIAG2** User-Defined Surface Diagnostic-2
The GEOS GCM provides Users with a built-in mechanism for archiving user-defined diagnostics. For a complete description refer to Diagnostic #84. The syntax for using the surface SDIAG2 diagnostic is given by

\[
\begin{align*}
\text{do } & j=1,jm \\
\text{do } & i=1,im \\
& qdiag(i,j,ISDIAG2) = qdiag(i,j,ISDIAG2) + \ldots \\
& \text{enddo} \\
& \text{enddo} \\
& \text{NSDIAG2} = \text{NSDIAG2} + 1
\end{align*}
\]

The diagnostics defined in this manner will automatically be archived by the production output programs (see Section 11).

86) **UDIAG1** User-Defined Upper-Air Diagnostic-1

The GEOS GCM provides Users with a built-in mechanism for archiving user-defined diagnostics. For a complete description refer to Diagnostic #84. The syntax for using the upper-air UDIAG1 diagnostic is given by

\[
\begin{align*}
\text{do } & L=1,nlay \\
\text{do } & j=1,jm \\
\text{do } & i=1,im \\
& qdiag(i,j,IUDIAG1+L-1) = qdiag(i,j,IUDIAG1+L-1) + \ldots \\
& \text{enddo} \\
& \text{enddo} \\
& \text{enddo} \\
& \text{NUDIAG1} = \text{NUDIAG1} + 1
\end{align*}
\]

The diagnostics defined in this manner will automatically be archived by the production output programs (see Section 11).

87) **UDIAG2** User-Defined Upper-Air Diagnostic-2

The GEOS GCM provides Users with a built-in mechanism for archiving user-defined diagnostics. For a complete description refer to Diagnostic #84. The syntax for using the upper-air UDIAG2 diagnostic is given by
The diagnostics defined in this manner will automatically be archived by the production output programs (see Section 11).

88) DIABU Total Diabatic Zonal U-Wind Tendency \((m/sec/day)\)

DIABU is the total time-tendency of the Zonal U-Wind due to Diabatic processes and the Analysis forcing.

\[
DIABU = \frac{\partial u}{\partial t_{Moist}} + \frac{\partial u}{\partial t_{Turbulence}} + \frac{\partial u}{\partial t_{Analysis}}
\]

Note, in Version 1 of the GEOS GCM there are no u-wind tendencies due to cumulus friction, (ie. \(\frac{\partial u}{\partial t_{Moist}} = 0\)). Also, in forecast simulations \(\frac{\partial u}{\partial t_{Analysis}} = 0\).

89) DIABV Total Diabatic Meridional V-Wind Tendency \((m/sec/day)\)

DIABV is the total time-tendency of the Meridional V-Wind due to Diabatic processes and the Analysis forcing.

\[
DIABV = \frac{\partial v}{\partial t_{Moist}} + \frac{\partial v}{\partial t_{Turbulence}} + \frac{\partial v}{\partial t_{Analysis}}
\]

Note, in Version 1 of the GEOS GCM there are no v-wind tendencies due to cumulus friction, (ie. \(\frac{\partial v}{\partial t_{Moist}} = 0\)). Also, in forecast simulations \(\frac{\partial v}{\partial t_{Analysis}} = 0\).

90) DIABT Total Diabatic Temperature Tendency \((deg/day)\)

DIABT is the total time-tendency of Temperature due to Diabatic processes and the Analysis forcing.

\[
DIABT = \frac{\partial T}{\partial t_{MoistProcesses}} + \frac{\partial T}{\partial t_{ShortwaveRadiation}} + \frac{\partial T}{\partial t_{LongwaveRadiation}} + \frac{\partial T}{\partial t_{Turbulence}} + \frac{\partial T}{\partial t_{Analysis}}
\]
If we define the time-tendency of Temperature due to Diabatic processes as

\[
\frac{\partial T}{\partial t} \text{ Diabatic} = \frac{\partial T}{\partial t} \text{ Moist Processes} + \frac{\partial T}{\partial t} \text{ Shortwave Radiation} + \frac{\partial T}{\partial t} \text{ Longwave Radiation} + \frac{\partial T}{\partial t} \text{ Turbulence}
\]

then, since there are no surface pressure changes due to Diabatic processes, we may write

\[
\frac{\partial T}{\partial t} \text{ Diabatic} = \frac{p^e}{\pi} \frac{\partial \pi \theta}{\partial t} \text{ Diabatic}
\]

where \( \theta = T/p^e \). Thus, \( \text{DIABT} \) may be written as

\[
\text{DIABT} = \frac{p^e}{\pi} \left( \frac{\partial \pi \theta}{\partial t} \text{ Diabatic} + \frac{\partial \pi \theta}{\partial t} \text{ Analysis} \right)
\]

Note, this diagnostic is only approximate since we have neglected the affect of the analysis on the surface pressure, ie.

\[
\frac{\partial T}{\partial t} \text{ Analysis} = \frac{p^e}{\pi} \left( \frac{\partial \pi \theta}{\partial t} \text{ Analysis} + \theta \frac{\partial \pi}{\partial t} \text{ Analysis} \left[ \frac{\sigma \pi \kappa}{p} - 1 \right] \right)
\]

This term will be added in subsequent versions of the GEOS GCM. In forecast simulations there is no error since \( \frac{\partial T}{\partial t} \text{ Analysis} = 0 \).

91) \( \text{DIABQ} \) Total Diabatic Specific Humidity Tendency \((g/kg/day)\)

\( \text{DIABQ} \) is the total time-tendency of Specific Humidity due to Diabatic processes and the Analysis forcing.

\[
\text{DIABQ} = \frac{\partial q}{\partial t} \text{ Moist Processes} + \frac{\partial q}{\partial t} \text{ Turbulence} + \frac{\partial q}{\partial t} \text{ Analysis}
\]

If we define the time-tendency of Specific Humidity due to Diabatic processes as

\[
\frac{\partial q}{\partial t} \text{ Diabatic} = \frac{\partial q}{\partial t} \text{ Moist Processes} + \frac{\partial q}{\partial t} \text{ Turbulence}
\]

then, since there are no surface pressure changes due to Diabatic processes, we may write

\[
\frac{\partial q}{\partial t} \text{ Diabatic} = \frac{1}{\pi} \frac{\partial \pi q}{\partial t} \text{ Diabatic}
\]

Thus, \( \text{DIABQ} \) may be written as

\[
\text{DIABQ} = \frac{1}{\pi} \left( \frac{\partial \pi q}{\partial t} \text{ Diabatic} + \frac{\partial \pi q}{\partial t} \text{ Analysis} \right)
\]
Note, this diagnostic is only approximate since we have neglected the affect of the analysis on the surface pressure, ie.

\[
\frac{\partial q}{\partial t_{\text{Analysis}}} = \frac{1}{\pi} \left( \frac{\partial \pi q}{\partial t_{\text{Analysis}}} - q \frac{\partial \pi}{\partial t_{\text{Analysis}}} \right)
\]

This term will be added in subsequent versions of the GEOS GCM. In forecast simulations there is no error since \( \frac{\partial q}{\partial t_{\text{Analysis}}} = 0 \).

92) Disabled

93) Disabled

94) Disabled

95) Disabled

96) Disabled

97) Disabled

98) Disabled

99) Disabled

100) Disabled

101) Disabled

102) *VINTUQ* Vertically Integrated Moisture Flux \((m/sec \cdot g/kg)\)

The vertically integrated moisture flux due to the zonal u-wind is obtained by integrating \(uq\) over the depth of the atmosphere at each model timestep, and dividing by the total mass.
of the column.

\[ \text{VINTUQ} = \frac{\int_{\text{surf}}^{\text{top}} u q \rho dz}{\int_{\text{surf}}^{\text{top}} \rho dz} \]

Using \( \rho \delta z = -\frac{\delta P}{g} = -\frac{\pi}{g} \delta \sigma \), we have

\[ \text{VINTUQ} = \int_{0}^{1} u q d\sigma \]

103) **VINTVQ Vertically Integrated Moisture Flux** \((m/sec \cdot g/kg)\)

The vertically integrated moisture flux due to the meridional \( v \)-wind is obtained by integrating \( vq \) over the depth of the atmosphere at each model timestep, and dividing by the total mass of the column.

\[ \text{VINTVQ} = \frac{\int_{\text{surf}}^{\text{top}} v q \rho dz}{\int_{\text{surf}}^{\text{top}} \rho dz} \]

Using \( \rho \delta z = -\frac{\delta P}{g} = -\frac{\pi}{g} \delta \sigma \), we have

\[ \text{VINTVQ} = \int_{0}^{1} v q d\sigma \]

104) **VINTUT Vertically Integrated Heat Flux** \((m/sec \cdot deg)\)

The vertically integrated heat flux due to the zonal \( u \)-wind is obtained by integrating \( uT \) over the depth of the atmosphere at each model timestep, and dividing by the total mass of the column.

\[ \text{VINTUT} = \frac{\int_{\text{surf}}^{\text{top}} u T \rho dz}{\int_{\text{surf}}^{\text{top}} \rho dz} \]

Using \( \rho \delta z = -\frac{\delta P}{g} = -\frac{\pi}{g} \delta \sigma \), we have

\[ \text{VINTUT} = \int_{0}^{1} u T d\sigma \]

105) **VINTVT Vertically Integrated Heat Flux** \((m/sec \cdot deg)\)
The vertically integrated heat flux due to the meridional $v$-wind is obtained by integrating $vT$ over the depth of the atmosphere at each model timestep, and dividing by the total mass of the column.

$$VINTVT = \frac{\int_{surf}^{top} vT \rho dz}{\int_{surf}^{top} \rho dz}$$

Using $\rho \delta z = -\frac{\delta p}{g} = -\frac{\pi}{g} \delta \sigma$, we have

$$VINTVT = \int_0^1 vT d\sigma$$

106 **CLDFRC** Total 2-Dimensional Cloud Fraction ($0 - 1$)

The 2-dimensional net cloud fraction as seen from the top of the atmosphere is given by

$$CLDFRC = 1 - \left(1 - MAX_{i=l_1}^{NLAY}[CLMO_i]\right) \prod_{i=l_1}^{NLAY}(1 - CLRO_i)$$

For a complete description of cloud/radiative interactions, see Section 3.2.3.

107) **QINT** Total Precipitable Water ($gm/cm^2$)

The Total Precipitable Water is defined as the vertical integral of the specific humidity, given by:

$$QINT = \int_{surf}^{top} \rho q dz$$

$$= \frac{\pi}{g} \int_0^1 q d\sigma$$

where we have used the hydrostatic relation $\rho \delta z = -\frac{\delta p}{g} = -\frac{\pi}{g} \delta \sigma$.

108) **U2M** Zonal U-Wind at 2 Meter Depth ($m/sec$)

The $u$-wind at the 2-meter depth is determined from the similarity theory:

$$U2M = \frac{u_*}{k} \frac{\psi_{m2m}}{\psi_{m2l}} \frac{u_{sl}}{W_s} = \frac{\psi_{m2m}}{\psi_{m2l}} \frac{u_{sl}}{W_s}$$

where $\psi_{m}(2m)$ is the non-dimensional wind shear at two meters, and the subscript $sl$ refers to the height of the top of the surface layer. If the roughness height is above two meters,
109) **V2M Meridional V-Wind at 2 Meter Depth (m/sec)**

The v-wind at the 2-meter depth is determined from the similarity theory:

\[
V2M = \frac{u^*}{k} \psi_{m2m} \frac{v_{sl}}{W_z} = \frac{\psi_{m2m}}{\psi_{m,st}} v_{sl}
\]

where \(\psi_m(2m)\) is the non-dimensional wind shear at two meters, and the subscript \(sl\) refers to the height of the top of the surface layer. If the roughness height is above two meters, \(V2M\) is undefined.

110) **T2M Temperature at 2 Meter Depth (deg K)**

The temperature at the 2-meter depth is determined from the similarity theory:

\[
T2M = P^\kappa \left( \frac{\theta^*}{k} (\psi_{h2m} + \psi_g) + \theta_{surf} \right) = P^\kappa \left( \theta_{surf} + \frac{\psi_{h2m} + \psi_g}{\psi_{sli} + \psi_g} (\theta_{sl} - \theta_{surf}) \right)
\]

where:

\[
\theta_* = \frac{\left( w'\theta' \right)}{u_*}
\]

where \(\psi_h(2m)\) is the non-dimensional temperature gradient at two meters, \(\psi_g\) is the non-dimensional temperature gradient in the viscous sublayer, and the subscript \(sl\) refers to the height of the top of the surface layer. If the roughness height is above two meters, \(T2M\) is undefined.

111) **Q2M Specific Humidity at 2 Meter Depth (g/kg)**

The specific humidity at the 2-meter depth is determined from the similarity theory:

\[
Q2M = P^\kappa \left( \frac{q^*}{k} (\psi_{h2m} + \psi_g) + q_{surf} \right) = P^\kappa \left( q_{surf} + \frac{\psi_{h2m} + \psi_g}{\psi_{sli} + \psi_g} (q_{sl} - q_{surf}) \right)
\]

where:

\[
q_* = \frac{\left( w'q' \right)}{u_*}
\]

where \(\psi_h(2m)\) is the non-dimensional temperature gradient at two meters, \(\psi_g\) is the non-dimensional temperature gradient in the viscous sublayer, and the subscript \(sl\) refers to the
height of the top of the surface layer. If the roughness height is above two meters, Q2M is undefined.

112) U10M Zonal U-Wind at 10 Meter Depth (m/sec)

The u-wind at the 10-meter depth is an interpolation between the surface wind and the model lowest level wind using the ratio of the non-dimensional wind shear at the two levels:

\[ U_{10M} = \frac{u_*}{k} \frac{u_{sl}}{W_s} = \frac{\psi_{m10m}}{\psi_{m_{sl}}} u_{sl} \]

where \( \psi_m(10m) \) is the non-dimensional wind shear at ten meters, and the subscript sl refers to the height of the top of the surface layer.

113) V10M Meridional V-Wind at 10 Meter Depth (m/sec)

The v-wind at the 10-meter depth is an interpolation between the surface wind and the model lowest level wind using the ratio of the non-dimensional wind shear at the two levels:

\[ V_{10M} = \frac{u_*}{k} \frac{v_{sl}}{W_s} = \frac{\psi_{m10m}}{\psi_{m_{sl}}} v_{sl} \]

where \( \psi_m(10m) \) is the non-dimensional wind shear at ten meters, and the subscript sl refers to the height of the top of the surface layer.

114) T10M Temperature at 10 Meter Depth (deg K)

The temperature at the 10-meter depth is an interpolation between the surface potential temperature and the model lowest level potential temperature using the ratio of the non-dimensional temperature gradient at the two levels:

\[ T_{10M} = P^* \left( \frac{\theta_*}{k} \left( \psi_{h10m} + \psi_{g} \right) + \theta_{surf} \right) = P^* \left( \theta_{surf} + \frac{\psi_{h10m} + \psi_{g}(\theta_{sl} - \theta_{surf})}{\psi_{h_{sl}} + \psi_{g}} \right) \]

where:

\[ \theta_* = -\left( \frac{w'\theta'}{u_*} \right) \]

where \( \psi_h(10m) \) is the non-dimensional temperature gradient at two meters, \( \psi_g \) is the non-dimensional temperature gradient in the viscous sublayer, and the subscript sl refers to the height of the top of the surface layer.
115) **Q10M** Specific Humidity at 10 Meter Depth (g/kg)

The specific humidity at the 10-meter depth is an interpolation between the surface specific humidity and the model lowest level specific humidity using the ratio of the non-dimensional temperature gradient at the two levels:

\[
Q_{10M} = P^\kappa \left( \frac{q_*}{k} (\psi_{h10m} + \psi_g) + q_{surf} \right) = P^\kappa (q_{surf} + \frac{\psi_{h10m}}{\psi_{hsl}} \psi_g (q_{sl} - q_{surf}))
\]

where:

\[
q_* = -\frac{(w'q)}{u_*}
\]

where \(\psi_h(10m)\) is the non-dimensional temperature gradient at two meters, \(\psi_g\) is the non-dimensional temperature gradient in the viscous sublayer, and the subscript \(sl\) refers to the height of the top of the surface layer.

116) **DTRAIN** Convective Cloud Detrainment (kg m/sec)

The cloud mass flux at the detrainment level per unit time, from all convective clouds generated during the output interval, is written:

\[
D\text{TRAIN} = \eta_{\sigma_D} m_B
\]

where \(\sigma_D\) is the detrainment \(\sigma\) level, \(m_B\) is the cloud base mass flux, and \(\eta\) is the entrainment, defined in Section 3.1.1.

117) **QFILL** Filling of negative Specific Humidity (g/kg/day)

Due to computational errors associated with the numerical scheme used for the advection of moisture, negative values of specific humidity may be generated. The specific humidity is checked for negative values after every dynamics timestep. If negative values have been produced, a filling algorithm is invoked which redistributes moisture from below (see Section 2.3). Diagnostic **QFILL** is equal to the net filling needed to eliminate negative specific humidity, scaled to a per-day rate:

\[
Q\text{FILL} = q^{n+1}_{final} - q^{n+1}_{initial}
\]

where

\[
q^{n+1} = (\pi q)^{n+1}/\pi^{n+1}
\]

118) **VINTQANA** Precipitable Moisture Adjustment from Analysis (mm/day)
For the time change in specific humidity due to the analysis increment, \( \frac{\partial q}{\partial t} \), the vertical integral or total precipitable amount is given by:

\[
VINTQANA = \int_{surf}^{top} \rho \frac{\partial q}{\partial t} \, dz = - \int_{surf}^{top} \frac{\partial q}{\partial t} \frac{dp}{g} = \frac{\pi}{g} \int_{0}^{1} \frac{\partial q}{\partial t} \, d\sigma
\]

Defining the specific humidity analysis increment, \( \frac{\partial q}{\partial t} \), as

\[
\frac{\partial q}{\partial t} = \frac{1}{\pi} \left( \frac{\partial \pi q}{\partial t} - \frac{\partial \pi}{\partial t} \right)
\]

we may write

\[
VINTQANA = \frac{1}{g} \int_{0}^{1} \left( \frac{\partial \pi q}{\partial t} - \frac{\partial \pi}{\partial t} \right) \, d\sigma
\]

119) \textbf{VINTQFIL} Precipitable Moisture Adjustment from Filling (mm/day)

When advective or other processes create spurious negative specific humidity, total moisture is conserved by filling the negative amounts by “borrowing” moisture from below. When there is not enough moisture in a column to adequately fill the negative values, the specific humidities are set to zero, creating an artificial source of moisture.

For a change in specific humidity due to filling, \( \Delta q_{fill} \), the vertical integral or total precipitable amount is given by:

\[
VINTQFIL = \int_{surf}^{top} \rho \Delta q_{fill} \, dz = - \int_{surf}^{top} \Delta q_{fill} \frac{dp}{g} = \frac{\pi}{g} \int_{0}^{1} \Delta q_{fill} \, d\sigma
\]

A precipitable adjustment rate is defined as the vertically integrated moisture adjustment per Dynamics time step, scaled to mm/day.
Part III

GEOS-1 GCM: User’s Guide

9 Introduction

In this section we discuss the general philosophy used in the construct of the GEOS GCM system, the necessary components of the system, and a basic User’s Guide on how to set-up and run experiments.

10 GEOS Superstructure

The superstructure of the GEOS GCM is designed so that the entire model may be invoked by a simple subroutine call. With each call of the main GCM driver, all model prognostic and diagnostic fields are updated by one timestep. A variety of applications may be built around the main GCM driver (GCMDRV), which allow for straight forecast simulations, assimilation cycles, and dynamic initialization procedures. It should be emphasized that what is meant by the GEOS System is simply the library of GCM subroutines which include GCMDRV and those that are subsequently called. These subroutines will be denoted as below GCMDRV. Those routines which are above GCMDRV include the Main Program, or the Application, and any model Output routines.

10.1 GEOS Applications

The GEOS Application can be defined as the Main Program from which the GEOS GCM and/or the GEOS Utility subroutines are called. All Applications are User-supplied, that is, they do not belong to the GEOS GCM compiled library. The purpose of the GEOS Application is a pre-determined experiment, or application, which the User wants to perform using the GEOS GCM library.

For example, let us create an Application which performs the following functions:

- Read a model restart from Unit1
- Integrate the model forward in time for NMAX iterations
- Write the resulting model restart to Unit2
This could be accomplished with the following FORTRAN code:

```
CALL RESTART (UNIT1, IREAD, ...)
DO    N = 1, NMAX
   CALL GCMDRV (Arg1, Arg2, Arg3, ...)
ENDDO
CALL RESTART (UNIT2, IWRITE, ...)
```

Here IREAD and IWRITE are read/write IO flags used by the GEOS Utility Subroutine RESTART to read/write GEOS Model restarts, and Arg1, Arg2, Arg3, ... are the arguments to GCMDRV. Both of these argument lists are described fully in the next section.

We can easily modify this application to run forecasts from the same initial state but using different GCMDRV model parameters:

```
CALL RESTART (UNIT1, IREAD, ...)
DO    N = 1, NMAX
   CALL GCMDRV (Arg1.1, Arg1.2, Arg1.3, ...)
ENDDO
REWIND UNIT1
CALL RESTART (UNIT1, IREAD, ...)
DO    N = 1, NMAX
   CALL GCMDRV (Arg2.1, Arg2.2, Arg2.3, ...)
ENDDO
CALL RESTART (UNIT2, IWRITE, ...)
```

Finally, we could run the model with the same set of GCMDRV model parameters but starting from two different initial states:

```
CALL RESTART (UNIT1, IREAD, ...)
```
DO N = 1,NMAX
CALL GCMDRV (Arg1, Arg2, Arg3,...)
ENDDO

CALL RESTART (UNIT2,IWRITE,...)
CALL RESTART (UNIT3,IREAD,...)

DO N = 1,NMAX
CALL GCMDRV (Arg1, Arg2, Arg3,...)
ENDDO

CALL RESTART (UNIT4,IWRITE,...)

These simple examples show the ease and flexibility of creating applications to perform experiments with the GEOS GCM. In the next section we describe fully the argument lists for subroutines GCMDRV and RESTART, and briefly describe GCMDRV’s flow structure.

10.2 Subroutines GCMDRV & RESTART

Subroutine GCMDRV is the single interface between User Application routines and the actual updating of model prognostic fields. Its flow structure can be separated into three essential parts:

- Read and refresh Boundary Conditions
- Compute time-tendencies of atmospheric state variables due to physical and hydrodynamical processes
- Integrate the atmospheric state variables forward in time by one timestep

Every call to GCMDRV updates the boundary conditions to the current date and time. In Version 1 of the GEOS GCM this is accomplished by a linear interpolation between monthly mean data sets. Subsequent versions of the GCM will allow for boundary condition data sets with varying frequency intervals. The boundary conditions which are currently updated are:

- Sea Surface Temperature
• Ground Wetness
• Surface Albedo
• Surface Roughness
• Sea-Ice Thickness

In addition, data is provided describing the model topography, topography variance, and the Land/Water/Glacier flags. These data are not interpolated to the current date and time since they are fixed quantities. Finally, the incident solar radiation at the top of the atmosphere is computed for the current date and time.

At prescribed time intervals, GCMDRV calls separate mini-drivers to update the time-tendencies of the prognostic state variables due to particular physical or hydrodynamical processes. Each mini-driver has its own timescale of operation. The mini-drivers associated with physical and hydrodynamical processes are:

• Moist Processes (Updated every 10 minutes)
• Shortwave Radiation (Updated every 3 hours using a normalized incident radiation, and adjusted every timestep using the true incident radiation)
• Longwave Radiation (Updated every 3 hours)
• Turbulence (Updated every 30 minutes)
• Eulerian Hydrodynamics (Updated every timestep)

The time-tendencies for the processes stated above are always kept in core during the GEOS GCM simulation. Thus, all prognostic fields are time-continuous while the time-tendencies of the prognostic fields due to the various processes are discontinuous at the above stated frequencies. These time-tendencies are summed together to provide a total time-tendency to be used in the time-integration scheme. Currently, the GEOS GCM can be run with either the Matsuno or the Leapfrog time scheme. The Leapfrog scheme has associated with it an Asselin (1972) time filter.
The calling sequence associated with GCMDRV and a description of the argument list is provided below:

```fortran
CALL GCMDRV ( KBC,N,NYMD,NHMS,SCHEME,NDT,ALPHA,QQINIT,QBUD )
```

<table>
<thead>
<tr>
<th>PARAMETER</th>
<th>TYPE</th>
<th>DESCRIPTION</th>
</tr>
</thead>
<tbody>
<tr>
<td>KBC</td>
<td>Integer</td>
<td>KBC is the unit from which Boundary Conditions are read. Every call to GCMDRV will initiate an update of the boundary conditions. Boundary conditions are updated by a linear interpolation between the two months kept in memory which strap the current date and time. Each time a new mid-month is passed or a new KBC value is encountered, two new months are read in. Setting KBC equal to zero disables all boundary condition updates.</td>
</tr>
<tr>
<td>N</td>
<td>Integer</td>
<td>N refers to the current time index, and is equal to 1 or 2. Its value always points to the updated fields and is modified by GCMDRV.</td>
</tr>
<tr>
<td>NYMD</td>
<td>Integer</td>
<td>NYMD is the current year/month/day in YYMMDD format. Its value is modified by GCMDRV.</td>
</tr>
<tr>
<td>NHMS</td>
<td>Integer</td>
<td>NHMS is the current hour/minute/second in HHMMSS format. Its value is modified by GCMDRV.</td>
</tr>
<tr>
<td>SCHEME</td>
<td>Character*4</td>
<td>SCHEME is the variable indicating the time scheme to be used for the integration. SCHEME may be set to 'MATS' for the Matsuno time scheme, or 'LEAP' for the Leapfrog time scheme. The SCHEME variable may be changed at any time during the forecast.</td>
</tr>
<tr>
<td>NDT</td>
<td>Integer</td>
<td>NDT is the timestep (in seconds) used for the forecast. NDT may be positive, zero, or negative, and may be changed at any time during the forecast.</td>
</tr>
<tr>
<td>ALPHA</td>
<td>Real</td>
<td>Asselin Filter Coefficient, ( \alpha ), to be used with the Leapfrog Time Scheme, where ( q_n^{\text{Filt}} = q_n (1 - \alpha) + \alpha \left( \frac{q_{n+1} + q_{n-1}}{2} \right) ).</td>
</tr>
<tr>
<td>QQINIT</td>
<td>Logical</td>
<td>QQINIT is a Logical flag parameter used by the Turbulence parameterization scheme. QQINIT should be set to TRUE if the GEOS GCM is starting from an initial condition which does not contain Turbulent Kinetic Energy as a prognostic state variable. After one successful call to GCMDRV, QQINIT should be set to FALSE.</td>
</tr>
<tr>
<td>QBUD</td>
<td>Logical</td>
<td>Logical flag to compute budget diagnostics.</td>
</tr>
</tbody>
</table>
The GEOS GCM Utility subroutine **RESTART** may be used within User-Defined Applications to read and/or write a GEOS GCM Restart dataset. The calling sequence associated with **RESTART** and a description of the argument list is provided below:

```
CALL RESTART ( KU,IOFLAG,NYMD,NHMS,N,NYMD0,NHMS0,JOB,NTYPE )
```

<table>
<thead>
<tr>
<th>PARAMETER</th>
<th>TYPE</th>
<th>DESCRIPTION</th>
</tr>
</thead>
<tbody>
<tr>
<td>KU</td>
<td>Integer</td>
<td>Unit number from which to read or to write the GEOS Restart</td>
</tr>
<tr>
<td>IOFLAG</td>
<td>Integer</td>
<td>IO Flag which is equal to: 1 to Read a GEOS Restart, 2 to Write a GEOS Restart.</td>
</tr>
<tr>
<td>NYMD</td>
<td>Integer</td>
<td>NYMD is the current year/month/day in YYMMDD format.</td>
</tr>
<tr>
<td>NHMS</td>
<td>Integer</td>
<td>NHMS is the current hour/minute/second in HHMMSS format.</td>
</tr>
<tr>
<td>N</td>
<td>Integer</td>
<td>N refers to the current time index, and is equal to 1 or 2. Its value always points to the most recently updated fields.</td>
</tr>
<tr>
<td>NYMD0</td>
<td>Integer</td>
<td>NYMD0 is the beginning year/month/day of the current experiment in YYMMDD format. NYMD0 is used in current Production Applications to control Output frequencies.</td>
</tr>
<tr>
<td>NHMS0</td>
<td>Integer</td>
<td>NHMS0 is the beginning hour/minute/second of the current experiment in HHMMSS format. NYMD0 is used in current Production Applications to control Output frequencies.</td>
</tr>
<tr>
<td>JOB</td>
<td>Character*8</td>
<td>An Experiment Identifier (usually a name or experiment number).</td>
</tr>
</tbody>
</table>
| NTYPE     | Integer  | NTYPE is the GEOS Restart Identification index.  
|           |          | NTYPE = -1 refers to a “First Guess” restart which was used for an analysis. NTYPE = 0 refers to the “After Analysis” or straight forecast restart. NTYPE = 1 refers to a restart from an IAU assimilation. |

It should be noted that if IOFLAG = 1, the **RESTART** parameters NYMD, NHMS, N, NYMD0, NHMS0, JOB, and NTYPE are read from the GEOS Restart dataset and passed back to the calling program. If IOFLAG = 2, the parameters NYMD, NHMS, N, NYMD0, NHMS0, JOB, and NTYPE passed into subroutine **RESTART** will be written out to the GEOS Restart.
11 GEOS Production

Although the GEOS GCM system is designed to work with a User supplied "Applications" program, several application programs have been developed by the DAO which are used in standard production forecast simulations and analyses, and are available as templates. These applications provide examples of convenient ways to control the input and output streams, the frequency and selection of diagnostic and prognostic fields for output, and the simulation start and end times. The application routine examples reside on the SGI workstation hera.gsfc.nasa.gov, in the directory /production/geos_das/geos1.1/applications. Some of them are described here in detail:

- **gcmprod.f** - A simple application routine designed to read and follow namelist-supplied directions to simulate a specific period and direct all "activated" diagnostics to an output stream. The application reads one namelist to establish the initial and final time and date of the simulation, the frequency of the output stream, and other run control parameters such as timestep and assim filter strength. Another namelist is read in which logical flags determine the selection of diagnostics to be activated and directed to the output stream.

- **gcmprodn.f** - An application routine which is designed to handle a somewhat more complex output stream for diagnostic and prognostic fields. The namelist controlled interface for the diagnostics involves the specification of an HHMMSS format variable for each activated diagnostic separately. Diagnostics may be averaged for any time period up to the length of the simulation, and the averaging period may be centered about the current time or may extend from the previous time to the present. The application determines how many different averaging periods and types are present, and directs the output to the appropriate number of separate output streams. Separate output routines are called to properly direct the diagnostic output. Control over the begin and end times of the simulation are also governed by namelist parameters.

- **amip.f** - An experiment-specific application routine that was designed to handle the control of an extended multi-year simulation with complex diagnostic output requirements. The application was written to direct diagnostic output to several output data streams each with a pre-specified set of diagnostics. Separate output routines are called for each data stream, providing them with the required list of diagnostic quantities.

In addition to flexible application routines, the GEOS GCM system also incorporates user defined output routines that can be called by the application. The output routines developed at the DAO write a combination of prognostic and/or diagnostic quantities, on model sigma levels or interpolated to pressure levels, in the Phoenix Output Format (see Section 11.2).
Just like the application routines, several output routines have been developed by the DAO and are available as examples. Some of them are described here:

diagprs.f - An output routine which accepts as arguments a specific list of diagnostic quantities and interpolates them to an argument specified list of pressure levels. The routine uses the model provided GETDIAG subprogram to obtain the proper time-averaged quantities. The code is written to distinguish between different quantities and attempts to use an appropriate interpolation algorithm. Output is written in Phoenix Format.

diagsig.f - Designed to accept an argument-specified list of diagnostics, and direct output in Phoenix Format to one output stream. The GETDIAG routine is used to retrieve diagnostic quantities in the proper units from the model.

glaoutc.f - When this output routine is called, all prognostic quantities and all activated diagnostic quantities are interpolated to an argument-list specified set of pressure levels. All output is directed to a Phoenix Format data stream.

sigoutc.f - All model prognostic and activated diagnostic quantities are directed to a single sigma-level Phoenix Output Format data set.

progprs.f - Designed to direct model prognostic quantities to a single output stream. Quantities are interpolated to a list of pressure levels.

progsig.f - Model prognostic quantities are simply directed to a single sigma-level Phoenix format data set.
11.1 Production Namelists

Two general namelists have been designed to setup and control the Production runs of the GEOS GCM. The namelists include information about the beginning date and time of the forecast, the ending date and time of the forecast, the duration of each job segment, and output characteristics.

The first namelist is called INPUT, and contains the parameters to control the beginning and ending times of each experiment, as well as some output information. The following INPUT namelist and subsequent description can be used as is or as a template for Users to build upon within their own Applications:

```
&INPUT
  JOB = ' 1234',
  XLAB = 'E1234 GEOS-GCM CLIMATE SIMULATION EXPERIMENT',
  RIMENT = 850101, 000000, 860101, 000000,
  NYMDB = .TRUE.,
  NDT = 225,
  LEAP = .FALSE.,
  MATS = .FALSE.,
  QQINIT = .FALSE.,
  ALPHA = 0.05,
  NDSEG = 2400000,
  NDOUT = 060000,
  PLEVS = 1000.0, 950.0, 900.0, 850.0, 800.0,
         700.0, 600.0, 500.0, 400.0, 300.0, 250.0,
         200.0, 150.0, 100.0, 70.0, 50.0, 30.0, 20.0
&END
```
<table>
<thead>
<tr>
<th>PARAMETER</th>
<th>TYPE</th>
<th>DESCRIPTION</th>
</tr>
</thead>
<tbody>
<tr>
<td>JOB</td>
<td>Character*8</td>
<td>Experiment Identifier (Usually a name or experiment number)</td>
</tr>
<tr>
<td>XLAB</td>
<td>Character*80</td>
<td>User Assigned Description of Current Experiment</td>
</tr>
<tr>
<td>NYMDB</td>
<td>Integer</td>
<td>Beginning Date of Experiment in YYMMDD format</td>
</tr>
<tr>
<td>NHMSB</td>
<td>Integer</td>
<td>Beginning Time of Experiment in HHMMSS format</td>
</tr>
<tr>
<td>NYMDE</td>
<td>Integer</td>
<td>Ending Date of Experiment in YYMMDD format</td>
</tr>
<tr>
<td>NHMSE</td>
<td>Integer</td>
<td>Ending Time of Experiment in HHMMSS format</td>
</tr>
<tr>
<td>NDT</td>
<td>Integer</td>
<td>Time step (in seconds) of the hydrodynamics</td>
</tr>
<tr>
<td>LEAP</td>
<td>Logical</td>
<td>Logical Flag to use the Leapfrog Time Scheme</td>
</tr>
<tr>
<td>MATS</td>
<td>Logical</td>
<td>Logical Flag to use the Matsuno Time Scheme</td>
</tr>
<tr>
<td>QQINIT</td>
<td>Logical</td>
<td>QQINIT is a Logical flag parameter used by the Turbulence parameterization scheme. QQINIT should only be set to TRUE if the GEOS GCM is starting from an initial condition which does not contain Turbulent Kinetic Energy as a prognostic state variable. After one successful call to GCMDRV, QQINIT should be set to FALSE. Asselin Filter Coefficient, ( \alpha ), to be used with the Leapfrog Time Scheme, where ( q_n^{\text{Filt}} = q_n(1 - \alpha) + \alpha \left( \frac{ q_{n+1} + q_{n-1}}{2} \right) ).</td>
</tr>
<tr>
<td>ALPHA</td>
<td>Real</td>
<td></td>
</tr>
<tr>
<td>NDSEG</td>
<td>Integer</td>
<td>Current job segment time interval in HHMMSS format</td>
</tr>
<tr>
<td>NDOUOUT</td>
<td>Integer</td>
<td></td>
</tr>
<tr>
<td>PLEVS</td>
<td>Real</td>
<td>Levels used for Sigma to Pressure Interpolation to create Pressure Level Output. (Maximum number of levels = 50)</td>
</tr>
</tbody>
</table>

When the GEOS GCM is run during a particular experiment, no diagnostic quantities will be produced or saved unless the User has explicitly enabled the desired diagnostics from the Applications program. Using the DAO supplied applications directly or as templates, a convenient NAMELIST method has been adopted for diagnostic control. Diagnostic quantities which are of interest to the User may be enabled by the application by including...
the Diagnostic Name from the Diagnostic Menu (Section 7) in the namelist INDIAG. For those Users employing the gcmprod.f application described in Section 11, simply preface the Diagnostic Name with the letter L. This variable will then be treated as a logical parameter. Those parameters set to TRUE will be enabled for the GEOS GCM simulation. An example of the INDIAG namelist using the Logical variable syntax is shown below:

```
&INDIAG
LUFLUX = .TRUE.,
LVFLUX = .TRUE.,
LHFLUX = .TRUE.,
LMOISTT = .TRUE.,
LMOISTQ = .TRUE.,
LRADLW = .TRUE.,
LRADSW = .TRUE.,
LPREACC = .TRUE.,
LPRECON = .TRUE.,
LOLR = .FALSE.,
LEVAP = .FALSE.,
LU2M = .FALSE.,
LV2M = .FALSE.,
LT2M = .FALSE.,
LQ2M = .FALSE.,
&END
```

In this example, the diagnostic quantities UFLUX, VFLUX, HFLUX, MOISTT, MOISTQ, RADIW, RADSW, PREACC, and PRECON will all be enabled during the simulation. The remaining diagnostics OLR, EVAP, U2M, V2M, T2M, and Q2m will be disabled during the run. It should be noted that explicitly defining the logical diagnostic parameters to FALSE in the INDIAG namelist is not necessary since this is the default. Using this method in conjunction with the application gcmprod.f will produce diagnostic quantities whose output frequency is governed by the INPUT namelist parameter NDOUT.

For those Users interested in a more sophisticated diagnostic output frequency algorithm, each diagnostic may be written out at its own unique frequency by using the DAO application gcmprodn.f. In this case, a particular diagnostic is turned on, or enabled, by specifying a non-zero diagnostic frequency, in HHMMSS format, for that diagnostic. The diagnostic frequency name is the same as the diagnostic name (from the Diagnostic Menu, Section 7) preceded by the letter N (eg, NUFLUX = 030000). In this example, the diagnostic UFLUX will be accumulated and written out every 3 hours. Every diagnostics has its own frequency, thus certain diagnostics may be written out more frequently than others. A separate file is defined for each unique diagnostic frequency. Thus, if the user turns on a total of 24 diagnostics, 12 with a 6 hour frequency, 10 with a 3 hour frequency, and 2 with a half-hour frequency, a total of 3 diagnostic datasets will be created. Finally, diagnostics are accumulated and written out with the time-stamp at the end of the accumulation period.
A time-stamp which occurs at the middle of the accumulating period may be obtained by making the diagnostic frequency negative. An example of the INDIAG namelist using the frequency variable syntax is shown below:

```
&INDIAG
  NUFLUX  = 030000,
  NVFLUX  = 030000,
  NHFLUX  = 030000,
  NMOISTT =-060000,
  NMOISTQ =-060000,
  NNRADLW =-060000,
  NNRADSW =-060000,
  NPREACC = 060000,
  NPRECON = 060000,
  NOLR    = 060000,
  NEVAP   = 060000,
  NU2M    = 010000,
  NV2M    = 010000,
  NT2M    = 010000,
  NQ2M    = 010000,
&END
```

In this example, four output data streams will be produced. Diagnostics U2M, V2M, T2M, and Q2M will be averaged over a one-hour frequency and written out with a time-stamp at the end of each hour. Diagnostics UFLUX, VFLUX, and HFLUX will be averaged over a three-hour time-period. Diagnostics PREACC, PRECON, OLR, and EVAP will be averaged over a six-hour period and written out with a time-stamp at the end of the averaging period. Diagnostics MOISTT, MOISTQ, RADLW, and RADSW will also be averaged over a six-hour period but will be written out with a time-stamp associated with the middle of the averaging period. All other diagnostics are disabled.
11.2 Phoenix Format

The Phoenix Format used in the archiving of the GEOS AMIP and the GEOS-DAS 5-Year Re-analysis employs a self-documenting output stream which describes each field and its attributes for each time period on the output dataset. At each output time period, Header information is given which describes the horizontal and vertical dimensions of the output, the time period of the output, and variable descriptions of the data type. The number and type of fields are User Defined.

The Phoenix format allows for writing data in three different ways (or configurations). The first, configuration #1, is simply a sequence of distinct 2-dimensional fields (such as sea-level pressure, topography, or selected individual levels of specific quantities such as wind or moisture). The second method, configuration #2, is 3-dimensional data which are ordered by quantity for each level. It is assumed that all quantities written in this manner are defined for all levels. The third, configuration #3, is 3-dimensional data which are ordered by level for each quantity. Here, the number of levels associated with each quantity may vary. The specific configurations used within any dataset are determined by the User. The following sample code may be used to read any Phoenix Format Output stream:

```c
PARAMETER ( IDIM = 144 ) ! IDIM=144 for 2.5 degree Longitude
PARAMETER ( JDIM = 091 ) ! JDIM=091 for 2.0 degree Latitude
DIMENSION Q( IDIM,JDIM )
CHARACTER*8 XLABEL(IO), JOB, NAME
DIMENSION SIGE (50), NDLEV(100), ZLEV(50)
DIMENSION IDUM(100), RDUM(100)
CHARACTER*8 NAMES(100), NAMEU(100), NAMED(100), CDUM(100)
CHARACTER*40 DESCS(100), DESCU(100), DESCDD(100), DDUM(100)

DATA KU /10/
READ(KU,END=500) JOB, NYMD, NHMS, NYMDO, NHMSO,
      XLABEL, IM, JM,
      NSFLD, NUFID, NDFLD,
      PTOP, NULEV, (ZLEV(K),K=1,NULEV),
      NLAY, (SIGE(K),K=1,NLAY+1),
      NDUM, (DDUM(N),IDUM(N),RDUM(N),CDUM(N),N=1,NDUM)
READ(KU) ( NAMES(N), DESCS(N), N=1,NSFLD )
```

77
DO N=I,NSFLD
  READ(KU) ZLEVID,NAME,Q
ENDDO

C *********************************************************
C **** READ 2-DIMENSIONAL FIELDS  ****
C *********************************************************

DO L=I,NLEVEL
  DO N=I,NSFLD
    READ(KU) ZLEVID,NAME,Q
  ENDDO
ENDDO

C *********************************************************
C **** READ 3-DIMENSIONAL FIELDS (ORDERED ALL LEVELS FOR EACH LEVEL)  ****
C *********************************************************

DO N=I,NSFLD
  DO L=I,NLEVEL
    READ(KU) ZLEVID,NAME,Q
  ENDDO
ENDDO

C *********************************************************
C **** READ 3-DIMENSIONAL FIELDS (ORDERED ALL LEVELS FOR EACH FIELD)  ****
C *********************************************************

DO N=1,NDFLD
  DO L=1,NDLEV(N)
    READ(KU) ZLEVID,NAME,Q
  ENDDO
ENDDO

500 CONTINUE
  STOP
  END
<table>
<thead>
<tr>
<th>PARAMETER</th>
<th>TYPE</th>
<th>DESCRIPTION</th>
</tr>
</thead>
<tbody>
<tr>
<td>JOB</td>
<td>Character*8</td>
<td>Experiment Identifier (Usually a name or experiment number)</td>
</tr>
<tr>
<td>XLAB</td>
<td>Character*80</td>
<td>User Assigned Description of Current Experiment</td>
</tr>
<tr>
<td>NYMD</td>
<td>Integer</td>
<td>Current Date of Experiment in YYMMDD format</td>
</tr>
<tr>
<td>NHMS</td>
<td>Integer</td>
<td>Current Time of Experiment in HHMMSS format</td>
</tr>
<tr>
<td>NYMD0</td>
<td>Integer</td>
<td>Beginning Date of Experiment in YYMMDD format</td>
</tr>
<tr>
<td>NHMS0</td>
<td>Integer</td>
<td>Beginning Time of Experiment in HHMMSS format</td>
</tr>
<tr>
<td>IM</td>
<td>Integer</td>
<td>Longitudinal dimension of data</td>
</tr>
<tr>
<td>JM</td>
<td>Integer</td>
<td>Latitudinal dimension of data</td>
</tr>
<tr>
<td>NLAY</td>
<td>Integer</td>
<td>Number of GCM vertical levels used in Experiment</td>
</tr>
<tr>
<td>SIGE</td>
<td>Real</td>
<td>GCM Sigma-Edge values</td>
</tr>
<tr>
<td>PTOP</td>
<td>Real</td>
<td>Model Top Pressure used in Experiment</td>
</tr>
<tr>
<td>ZLEV</td>
<td>Real</td>
<td>Upper-Air Level Values archived on data (Pressure or Sigma Level)</td>
</tr>
<tr>
<td>NSFLD</td>
<td>Integer</td>
<td>Number of 2-Dimensional Fields written in Config. #1</td>
</tr>
<tr>
<td>NUFLD</td>
<td>Integer</td>
<td>Number of 3-Dimensional Fields ordered by Level written in Config. #2</td>
</tr>
<tr>
<td>NDFLD</td>
<td>Integer</td>
<td>Number of 3-Dimensional Fields ordered by Quantity written in Config. #3</td>
</tr>
<tr>
<td>NAMES</td>
<td>Character*8</td>
<td>Names of 2-Dimensional Fields defined by NSFLD</td>
</tr>
<tr>
<td>DESCS</td>
<td>Character*40</td>
<td>Description of 2-Dimensional Fields defined by NSFLD</td>
</tr>
<tr>
<td>NAMEU</td>
<td>Character*8</td>
<td>Names of 3-Dimensional Fields ordered by Level defined by NUFLD</td>
</tr>
<tr>
<td>DESCU</td>
<td>Character*40</td>
<td>Description of 3-Dimensional Fields ordered by Level defined by NUFLD</td>
</tr>
<tr>
<td>NULEV</td>
<td>Integer</td>
<td>Number of Upper-Air Levels associated with all fields in Config. #2</td>
</tr>
<tr>
<td>NAMED</td>
<td>Character*8</td>
<td>Names of 3-Dimensional Fields ordered by Quantity defined by NDFLD</td>
</tr>
<tr>
<td>DESCD</td>
<td>Character*40</td>
<td>Description of 3-Dimensional Fields ordered by Quantity defined by NDFLD</td>
</tr>
<tr>
<td>NDLEV</td>
<td>Integer</td>
<td>Number of Upper-Air Levels associated with each field in Config. #3</td>
</tr>
<tr>
<td>NDUM</td>
<td>Integer</td>
<td>Number of User-Defined Header variables. NDUM, DDUM, IDUM, RDUM, and CDUM may be used by Users to add REAL, INTEGER, or CHARACTER information in the Header concerning the User's experiment.</td>
</tr>
<tr>
<td>DDUM</td>
<td>Character*40</td>
<td>User-Defined Description of Header variables</td>
</tr>
<tr>
<td>IDUM</td>
<td>Integer</td>
<td>User-Defined Integer Header variables</td>
</tr>
<tr>
<td>RDUM</td>
<td>Real</td>
<td>User-Defined Real Header variables</td>
</tr>
<tr>
<td>CDUM</td>
<td>Character*8</td>
<td>User-Defined Character Header variables</td>
</tr>
<tr>
<td>PLEVS</td>
<td>Real</td>
<td>Levels used for Sigma to Pressure Interpolation to create Pressure Level Output. (Maximum number of levels = 50)</td>
</tr>
<tr>
<td>ZLEVID</td>
<td>Real</td>
<td>Level Identifier (Pressure or Sigma Level) of Quantity being read</td>
</tr>
<tr>
<td>NAME</td>
<td>Character*8</td>
<td>Character Name of Quantity being read</td>
</tr>
<tr>
<td>Q</td>
<td>Real</td>
<td>Two-Dimensional Quantity being read</td>
</tr>
</tbody>
</table>
It should be noted that the Datasets produced for the GEOS-1 DAS 5-year re-analysis and the GEOS-1 GCM AMIP simulation have the following characteristics and User-Defined Header information:

- The horizontal resolution for all data is constant for all time periods.
- All horizontal data, $q(i, j)$, is written with $i = 1$ corresponding to $Longitude = -180$ (International Dateline), and $j = 1$ corresponding to $Latitude = -90$ (South Pole).
- The data written in Configuration # 1 are generally surface data (such as Sea-Level Pressure, Surface Topography, Land/Water Masks, etc.).
- The data written in Configuration # 2 are generally Upper-Air prognostic quantities (such as wind, temperature, moisture, etc.).
- The data written in Configuration # 3 are generally GCM Model Diagnostic quantities (such as Diabatic Heating). Some Diagnostics are only defined at one level (such as Precipitation, Outgoing Longwave Radiation, etc.).

The User-Defined Header information has been defined as:

```
NDUM       =   2
DDUM(1)    =   'OUTPUT RECORD TYPE (INTEGER)'
IDUM(1)    =   -1 FIRST GUESS
              0 AFTER ANALYSIS (or STRAIGHT FORECAST)
              1 INITIALIZED ANALYSIS
RDUM(1)    =   0.100000E+16
CDUM(1)    =   'LOG8R'
DDUM(2)    =   'UNDEFINED FILL VALUE (REAL)'
IDUM(2)    =   0
RDUM(2)    =   0.100000E+16
CDUM(2)    =   'UNDEP'
```
11.3 Geopotential Heights on Sigma Surfaces

In the GEOS-1 GCM, the pressure gradient computed on sigma surfaces is derived by removing the large-scale component due to a mean potential temperature and surface pressure. Therefore, the perturbation geopotential heights have been archived on the GEOS-1 GCM sigma level datasets.

The pressure gradient on a constant pressure surface is related to the gradient on a constant sigma surface by

\[ \nabla_p \Phi = \nabla_\sigma \Phi + c_p \theta \nabla_\sigma p^\kappa, \]

(25)

while the hydrostatic equation is given as

\[ \frac{\partial \Phi}{\partial p^\kappa} = -c_p \theta. \]

(26)

Defining \( \Phi = \Phi + \Phi' \) and \( \theta = \theta + \theta' \), we may write

\[ \frac{\partial (\Phi + \Phi')}{\partial p^\kappa} = -c_p (\theta + \theta') \]

(27)

or

\[ \frac{\partial \Phi}{\partial p^\kappa} = -c_p \theta \quad \Rightarrow \quad \Phi = -c_p \theta (p^\kappa - p_0^\kappa) \]

(28)

\[ \frac{\partial \Phi'}{\partial p^\kappa} = -c_p \theta'. \]

(29)

Thus, we may express the pressure gradient on sigma surfaces as

\[
\begin{align*}
\nabla_\sigma \Phi + c_p \theta \nabla_\sigma p^\kappa &= \nabla \left( \Phi + \Phi' \right) + c_p \left( \theta + \theta' \right) \nabla p^\kappa \\
&= \nabla \Phi + \nabla \Phi' + c_p \theta \nabla p^\kappa + c_p \theta' \nabla p^\kappa \\
&= \nabla \Phi + c_p \theta \nabla p^\kappa + \nabla \Phi' + c_p \theta \nabla p^\kappa + c_p \theta' \nabla p^\kappa \\
&= \nabla \Phi' + c_p \theta' \nabla p^\kappa 
\end{align*}
\]

(30)
where

\[
\Phi'_\ell - \Phi_\ell = \Phi_\ell - c_p\bar{\theta} (p^*_0 - p^*_\ell)
\]

\[
\theta'_\ell = \theta_\ell - \bar{\theta}; \quad \bar{\theta} = 280/1000c
\]

\[
p_0 = 1000.
\]

(31)
11.4  Interpolation from Sigma to Pressure Surfaces

While the GEOS GCM uses the sigma coordinate in the vertical for the discritization of
the primitive equations, the current analysis system requires “first guess” fields to be on
prescribed pressure surfaces. In addition, it is often desirable to process model output on
pressure surfaces for scientific studies and comparisons to observations. The method used
for the sigma coordinate to pressure coordinate interpolation is consistant with the form of
the hydrostatic equation used in the GEOS GCM.

The vertical grid and index notation for sigma levels used in the GEOS GCM is given by

\[
\begin{align*}
\Phi^{P_0} & \theta \; q & \ell - 1 \\
\Phi & \; q_* & \ell e \\
\Phi^{P_\ell} & \theta \; q & \ell
\end{align*}
\]

Figure 6: Vertical placement and index notation for sigma levels in the GEOS GCM

The form of the hydrostatic equation used in the GEOS GCM is given by

\[
\frac{\partial \Phi}{\partial P_\kappa} = -c_p \theta
\]

where \( \Phi \) is the geopotential height, and \( \theta = T/P_\kappa \). Similarly, for an arbitrary quantity, \( q \),
we may define

\[
\frac{\partial q}{\partial P_\kappa} = c
\]

where \( c \) is a constant. Thus, we may write

\[
\frac{\partial q}{\partial P_\kappa} = c \quad \Rightarrow \quad \frac{q_\ell - q_\ell - 1}{P_\ell - P_\ell - 1} = \frac{q_\ell - q_*}{P_\ell - P_*}
\]

or

83
This method is used for the winds, temperature, and relative humidity. Specific humidity at pressure levels is obtained from the previously interpolated temperature and relative humidity. For target pressures above the first model level, a linear extrapolation in $P^\kappa$ from levels 1 and 2 is used. For target pressures between the surface pressure and the lowest model level, the quantity is assumed constant and is equal to its value in the lowest model level. For target pressures greater than the surface pressure (below topography), the quantity is undefined. It should also be noted that all GEOS GCM diagnostic quantities on pressure surfaces have been similarly linearly interpolated in $P^\kappa$.

The method used in the interpolation of geopotential heights, in particular, must be intimately linked to the finite-differencing scheme used in the background model in order to minimize computational error and prevent spurious height gradients (or geostrophic imbalances). The technique described here, formulated at UCLA by Arakawa and Suarez (personal communication), uses two separate interpolation algorithms depending on whether you are above or below the nearest sigma-level interface, $\ell e$, and has the following properties:

1) $\Phi_{P_1} = \Phi_{\ell}$ for $P_1 = P_\ell$
2) $\Phi_{P_2} = \Phi_{\ell-1}$ for $P_2 = P_{\ell-1}$
3) $\Phi_{P_1} = \Phi_{P_2}$ for $P_1 = 2 P_{\ell e}$
4) $\nabla_p \Phi_{P_1} = \nabla_\sigma \Phi_{\ell} + c_p \theta_\ell \nabla_\sigma P^\kappa_{\ell}$ for $P_1 = P_{\ell}$
5) $\nabla_p \Phi_{P_2} = \nabla_\sigma \Phi_{\ell-1} + c_p \theta_{\ell-1} \nabla_\sigma P^\kappa_{\ell-1}$ for $P_2 = P_{\ell-1}$

Figure 7: Index notation for vertical interpolation of Geopotential Heights

\[
q^* = q_{\ell-1} \left[ 1 - \frac{P^\kappa_{\ell-1} - P^\kappa_*}{P^\kappa_{\ell-1} - P^\kappa_{\ell}} \right] + q_{\ell} \left[ \frac{P^\kappa_{\ell-1} - P^\kappa_*}{P^\kappa_{\ell-1} - P^\kappa_{\ell}} \right] \tag{35}
\]
Thus, the interpolation not only preserves the heights on sigma surfaces whose target pressure coincides with the sigma level pressure, but also the horizontal gradient of the height on pressure surfaces is defined accurately as the horizontal gradient of the sigma level height plus the term corresponding to the horizontal gradient of the surface pressure.

We begin by writing the hydrostatic equation in the form

\[
\frac{\partial \Phi}{\partial P^\kappa} = -c_p \theta .
\]  

Following Arakawa and Suarez (1983), we may write

\[
\Phi_{\ell-1} = \Phi_\ell + c_p \hat{\theta}_{\ell e}(P_\ell^{\kappa} - P_{\ell-1}^{\kappa}),
\]

\[
\hat{\theta}_{\ell e} = \frac{(\hat{P}_{\ell e}^{\kappa} - P_{\ell-1}^{\kappa})\theta_{\ell-1} + (P_\ell^{\kappa} - \hat{P}_{\ell e}^{\kappa})\theta_\ell}{(P_\ell^{\kappa} - P_{\ell-1}^{\kappa})}.
\]

Thus, for the two regions in Figure 7, we have:

1) For \( P_\ell \geq P_1 \geq \hat{P}_{\ell e} \)

\[
\Phi_{P_1} = \Phi_\ell + c_p \int_{P_1}^{P_\ell} \Theta_1 dP^\kappa
\]

2) For \( \hat{P}_{\ell e} \geq P_2 \geq P_{\ell-1} \)

\[
\Phi_{P_2} = \Phi_{\ell-1} + c_p \int_{P_{\ell-1}}^{P_2} \Theta_2 dP^\kappa
\]

where

\[
\Theta_1(P^\kappa) = \frac{(P_\ell^{\kappa} - P^\kappa) \hat{\theta}_{\ell e} + (P^\kappa - \hat{P}_{\ell e}^{\kappa}) \theta_\ell}{(P_\ell^{\kappa} - \hat{P}_{\ell e}^{\kappa})}
\]

\[
\Theta_2(P^\kappa) = \frac{(\hat{P}_{\ell e}^{\kappa} - P^\kappa) \theta_{\ell-1} + (P^\kappa - P_{\ell-1}^{\kappa}) \hat{\theta}_{\ell e}}{(\hat{P}_{\ell e}^{\kappa} - P_{\ell-1}^{\kappa})}
\]

Here \( \Theta_1(P^\kappa) \) and \( \Theta_2(P^\kappa) \) are the interpolated \( \theta \) values from levels \( \ell \) and \( \ell - 1 \) to the pressure \( P \). Using (42) in (39), we may define
\[ I_1 = \int_{P_1}^{P_t} \Theta_1 dP^\kappa \]
\[ = \int_{P_1}^{P_t} \left( \frac{P_\ell - P^\kappa}{P_\ell - \tilde{P}_\ell^\kappa} \right) \hat{\theta}_\ell e + (P^\kappa - \tilde{P}_\ell^\kappa) \frac{\theta_\ell}{dP^\kappa} \]
\[ = \frac{1}{(P_\ell^\kappa - P_\ell^{-1})} \right] \left[ \hat{\theta}_\ell e \left( P_\ell^\kappa \left[ P_\ell^\kappa - P_\ell^{-1} \right] - \frac{1}{2} \left[ P_\ell^\kappa^2 - P_1^\kappa^2 \right] \right) + \hat{\theta}_\ell \left( \frac{1}{2} \left[ P_\ell^\kappa^2 - P_1^\kappa^2 \right] - \tilde{P}_\ell^\kappa \left[ P_\ell^\kappa - P_1^\kappa \right] \right) \]
\[ = \frac{1}{2} \left( \frac{P_\ell^\kappa - P_\ell^{-1}}{P_\ell^\kappa - \tilde{P}_\ell^\kappa} \right) \left[ \hat{\theta}_\ell e \left( P_\ell^\kappa - P_\ell^{-1} \right) - \theta_\ell \left( P_\ell^\kappa - P_\ell^{-1} \right) \right] . \]

Similarly, we may define

\[ I_2 = \int_{P_{t-1}}^{P_t} \Theta_2 dP^\kappa \]
\[ = \int_{P_{t-1}}^{P_t} \left( \frac{\tilde{P}_\ell^\kappa - P^\kappa}{\tilde{P}_\ell^\kappa - P_{t-1}^\kappa} \right) \tilde{\theta}_{\ell-1} + (P^\kappa - P_{t-1}^\kappa) \frac{\tilde{\theta}_\ell}{dP^\kappa} \]
\[ = \frac{1}{(P_\ell^\kappa - P_{t-1}^\kappa)} \right] \left[ \tilde{\theta}_{\ell-1} \left( \tilde{P}_\ell^\kappa \left[ P_2^\kappa - P_{t-1}^\kappa \right] - \frac{1}{2} \left[ P_2^\kappa^2 - P_{t-1}^\kappa^2 \right] \right) + \tilde{\theta}_\ell \left( \frac{1}{2} \left[ P_2^\kappa^2 - P_{t-1}^\kappa^2 \right] - P_{t-1}^\kappa \left[ P_2^\kappa - P_{t-1}^\kappa \right] \right) \]
\[ = \frac{1}{2} \left( \frac{P_\ell^\kappa - P_{t-1}^\kappa}{P_\ell^\kappa - \tilde{P}_{t-1}^\kappa} \right) \left[ \tilde{\theta}_{\ell-1} \left( P_\ell^\kappa - P_{t-1}^\kappa \right) - \tilde{\theta}_\ell \left( P_\ell^\kappa - P_{t-1}^\kappa \right) \right] . \]
Therefore, for the interpolation of heights within regions (1) and (2) from Figure 7, we may write:

1) For \( P_1 \geq P_t \geq \dot{P}_{\ell e} \)

\[
\Phi_{P_1} = \Phi_t + \frac{c_p}{2} \left( \frac{P_t^c - P_1^c}{\dot{P}_{\ell e}^c - P_{\ell e}^c} \right) \left[ \theta_{\ell e} (P_t^e - P_1^e) - \theta_t (2\dot{P}_{\ell e}^e - P_t^e - P_1^e) \right] \tag{51}
\]

2) For \( \dot{P}_{\ell e} \geq P_2 \geq P_{\ell e} \)

\[
\Phi_{P_2} = \Phi_{\ell e - 1} + \frac{c_p}{2} \left( \frac{P_{\ell e}^c - P_{\ell e - 1}^c}{\dot{P}_{\ell e}^c - P_{\ell e}^c} \right) \left[ \theta_{\ell e - 1} (2\dot{P}_{\ell e}^e - P_2^e - P_{\ell e - 1}^e) + \theta_{\ell e} (P_2^e - P_{\ell e - 1}^e) \right] \tag{52}
\]
11.5 Sea-Level Pressure

In addition to the archiving of surface pressure, which is a prognostic variable in the GEOS-1 GCM, on all GEOS Production output streams, the sea-level pressure is also computed and archived. Also, during the GEOS data assimilation cycle, sea-level pressure rather than surface pressure is analyzed. The method used to compute sea-level pressure from surface pressure involves integrating the hydrostatic equation from sea-level to the surface, and an extrapolation of the temperature profile below topography.

The hydrostatic equation may be written as:

$$\frac{\partial p}{\partial z} = -\rho g = -\frac{pg}{RT}.$$  \hspace{1cm} (53)

Integrating (53) from sea-level, $sl$, to the surface, $s$, we may write

$$\int_{sl}^{s} \frac{dp}{p} = -\int_{sl}^{s} \frac{d\Phi}{RT},$$ \hspace{1cm} (54)

where we have used $d\Phi = g dz$. Assuming a mean temperature, $\overline{T}$, between the sea-level and surface, (54) may be written as

$$\ln \left( \frac{p_s}{p_{sl}} \right) = -\frac{\Phi_s}{RT},$$ \hspace{1cm} (55)

or

$$p_{sl} = p_s e^{\Phi_s/RT}.$$ \hspace{1cm} (56)

The mean temperature $\overline{T}$ is defined as the average of the surface and sea-level temperatures:

$$\overline{T} = \frac{1}{2} (T_s + T_{sl}),$$ \hspace{1cm} (57)

where $T_{sl}$ is obtained by assuming the temperature follows the moist adiabatic lapse rate

$$\frac{dT}{dz} = -\beta \quad \Rightarrow \quad T_{sl} = T_s + \frac{\beta \Phi_s}{g}$$ \hspace{1cm} (58)

where $\beta = 6.5^o K/km$. The surface temperature used for the sea-level pressure calculation is obtained by averaging the potential temperature, $\theta$, in the lowest 100 mb of the model.
This allows variable resolution changes in the PBL while maintaining consistent sea-level pressure values between resolutions. We have

\[ T_s \equiv \bar{\theta} p_s^* \quad ; \quad \theta = T/p^* \]  

(59)

where

\[ \bar{\theta} = \frac{\sum \theta \Delta \sigma}{\sum \Delta \sigma} \]  

(60)

such that

\[ \sum \Delta \sigma = \frac{\sum \Delta p}{\pi} \approx \frac{100 \, mb}{\pi}. \]  

(61)

Here \( \pi = p_s - p_{top} \).
12 GEOS Unix Script

The GEOS GCM is currently being run on NASA/GSFC's Cray C-90 computer system under the Cray UNICOS environment. There are five essential components to the basic script which are needed in order to perform experiments with the GEOS GCM:

- Create GEOS GCM NAMELIST for model parameters and diagnostics
- Create DIAGSIZE Subroutine for enabled diagnostics
- Remote copy frozen GEOS Libray, Application and Output Routines from geos_das@dao
- Compile User Application and Output Routines
- Load and Run GEOS simulation

Each of these elements are briefly discussed in the following subsections.

12.1 Creating the GEOS GCM Namelist

The following example shows the creation of a typical GCM namelist to be used during a GEOS simulation:

```
# Setup GCM Namelist

/bin/rm gcmrun.namelist
cat > gcmrun.namelist <<'EOF'
&INPUT
  JOB = '1234',
  XLAB = 'E1234 GEOS-GCM CLIMATE SIMULATION EXPE',
        'REIMENT',
  NYMDB = 850101,
  NHMSB = 000000,
  NYMDE = 860101,
  NHMSE = 000000,
  NDT = 225,
  LEAP = .TRUE.,
  MATS = .FALSE.,

EOF
```
For a complete description of GCM Namelist parameters, see Section 11.1.

12.2 Creating the DIAGSIZE Subroutine

Using the DIAGSIZE subroutine is a convenient method to allocate just enough permanent memory during a run to be used for diagnostic storage. All GEOS GCM diagnostic quantities are stored in the single diagnostic array QDIAG which is located in the DIAG COMMON. In order to minimize the maximum memory requirement used by the model, the GEOS GCM libraries are compiled with room for only one horizontal diagnostic array. All production DAO applications contain a call to subroutine DIAGSIZE, which expands the size of the DIAG COMMON to include the total number of surface and upper-air diagnostics the User has enabled. For the diagnostics enabled in the preceding section, the appropriate DIAGSIZE subroutine should be defined as:

```
#########################################################################
###### Create DIAGSIZE Routine to control Total Diagnostics
#########################################################################
```
We see that the User needs to simply assign the appropriate values for the number of surface diagnostics (NDIAGS) and the number of upper-air diagnostics (NDIAGU) which are enabled in the User's namelist.

12.3 Remote copy frozen GEOS Library

In order to gain access to the GEOS GCM, the User must remote copy the frozen production system from the DAO production workstation, geos_das@dao. In addition to the GEOS GCM library, the User may also remote copy production application programs and output routines. The COMMON files associated with these programs must also be copied to the User's local directory. Finally, the GEOS GCM boundary conditions and other GEOS input datasets must be obtained from the production workstation. An example of the syntax necessary for these remote copies is as follows:

```
rcp geos_das@dao:/production/geos_das/geosl.1/gcm/vc5.2b20.macro vc5.2b20
rcp geos_das@dao:/production/geos_das/geosl.1/applications/gcmprodn.f.
rcp geos_das@dao:/production/geos_das/geosl.1/applications/progprs.f.
rcp geos_das@dao:/production/geos_das/geosl.1/applications/progsig.f.
```
12.4 Compile User Application and Output Routines

Since the GEOS GCM library only contains subroutines from GCMDRV and below, the User must provide compiled Main Programs, or Applications, to perform specific experiments. For complete consistancy with the GEOS library, the Application and Output routines should be compiled in the same manner as the background model. An inline compile routine to accomplish this is shown below:

```
# Create COMPILE Routine and Compile Userappl & Output
RM compile
cat > compile <<'EOF'
/lib/cpp -P -I . $1.f $1.cpp
fpp -dec $1.cpp > $1.fpp
cft77 -i 64 -a static -V -exm $1.fpp
mv $1.fpp.o $1.o
'EOF'
chmod +x compile
./compile gcmprodn 
./compile progprs
./compile progsig
./compile diagprs
./compile diagsize
```
Here we have compiled those routines which were remote copied in Section 12.3.

12.5 Load and Run GEOS Simulation

Following successful compilations, the User must load the Application and Output routines together with the GEOS GCM library, creating an executable `a.out`. Then, assuming that an appropriate GEOS GCM Restart has been brought to the current local directory as `gcmrst`, the GEOS simulation may be performed:

```
# Load GCM, Applications and Output Routines, and Run

segldr -M,s -l ./vc5.2b20 -- gcmprodn.o progsig.o progprs.o diagprs.o diagsize.o

assign -a gcmrun.namelist         fort.11
assign -a gcmrst                  fort.12
assign -a prog.sig               -N ieee -F f77 fort.34
assign -a prog.prs               -N ieee -F f77 fort.35
assign -a diag.prs.a             -N ieee -F f77 fort.41
assign -a diag.prs.b             -N ieee -F f77 fort.42
assign -a diag.prs.c             -N ieee -F f77 fort.43
assign -a gcmbcs                 -N ibm -F COS fort.16
assign -a gcmo3.data             fort.20
assign -a zonalT.data            fort.52

setenv NCPUS 4
a.out
```

Note that `NCPUS` has been set as an environmental variable equal to 4 in order to take advantage of the macro-tasking properties of the GEOS GCM. In this example, file units 34, 35, 41, 42, 43 are associated with Output data streams, and file units 16, 20, 52 are input datasets and boundary conditions for the GEOS GCM. Once the GEOS GCM simulation has been completed, the resulting output data streams should be moved to permanent storage.
Acknowledgements

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This technical report documents Version 1 of the Goddard Earth Observing System (GEOS) General Circulation Model (GCM). The GEOS-1 GCM is being used by NASA's Data Assimilation Office (DAO) to produce multiyear data sets for climate research. This report provides a documentation of the model components used in the GEOS-1 GCM, a complete description of model diagnostics available and a User's Guide to facilitate GEOS-1 GCM experiments.