A Comparison of the Climate Simulated by the NCAR Community Climate Model (CCM1:R15) with ECMWF Analyses

WILLIAM J. RANDEL AND DAVID L. WILLIAMSON

National Center for Atmospheric Research, * Boulder, Colorado

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ABSTRACT

Detailed comparisons are made between the climate simulated by a seasonal version of the NCAR Community Climate Model (CCM1) at 12 level, R15 spectral resolution, and that revealed by ECMWF operational analyses over 1980–86 truncated to a similar resolution. A variety of circulation statistics are presented to reveal the spatial character and seasonality of CCM1 biases in temperatures, winds, and wave flux quantities. CCM1 biases are typical of current climate models run at similar resolution. Interrelationships between the above biases are a focus of this study, in particular using wave–mean flow interaction diagnostics.

CCM1 exhibits a westerly zonal wind bias in the tropics and a lack of westerlies in the high latitude Southern Hemisphere (SH). The tropical zonal mean meridional circulation (Hadley cell) in the model is approximately a factor of two too weak. The poleward eddy heat flux is accurately simulated, but the poleward eddy momentum flux is severely underestimated, particularly in the SH. There is a resulting excessive large-scale wave drag in the model extratropical upper troposphere, in qualitative agreement with the weak model high latitude westerlies (and temperature bias patterns). Conversely, the model tropical zonal wind bias does not appear to be related to influences by large-scale waves. Wave flux biases are compared for stationary and transient statistics; model stationary waves are in good agreement with observations, while the largest relative momentum flux error is found for higher frequency transient waves.

1. Introduction

In the past decade global atmospheric models have improved to the point where they simulate many properties of the general circulation of the atmosphere very well. Because of the quality of their simulations, they have become a useful tool for scientists to explore global atmospheric phenomena. They provide a basis for controlled experiments, which are impossible to perform in the atmosphere. In addition, they allow complete time–space sampling of state variables. Such complete sampling is unlikely to ever be obtainable from the atmosphere.

The Community Climate Model (CCM) is a comprehensive, three-dimensional global atmospheric model that has been assembled at the National Center for Atmospheric Research (NCAR). The model is applied by many researchers from NCAR and the university community to a wide variety of problems. The more recent applications over the last few years are summarized in Williamson (1988b). Periodically, as the model evolves, versions are frozen and documented to provide controls for experimental application and further development. This documentation consists of both algorithm and code descriptions as well as summaries of circulation statistics of the control simulations. This paper contributes to the documentation of a recent version of the CCM, namely CCM1 which was frozen during June 1987.

We present a variety of circulation statistics to reveal the spatial character and seasonality of CCM1 biases in temperature, winds, and wave flux quantities. These statistics are compared to those from the European Centre for Medium Range Weather Forecasts (ECMWF) analyses truncated to the same resolution. The biases revealed here for CCM1 are similar in many respects to those found in other climate models run at similar resolution (WMO 1988). Two currently used models for which circulation statistics are well documented are the Canadian Climate Center (CCC) model run at (triangular truncation) T20 and 11 vertical sigma levels (Boer et al. 1984; WMO 1988) and the ECMWF model run at T21 and 16 vertical levels, used for climate simulations at the University of Hamburg (Fischer 1987; WMO 1988); we discuss some detailed comparisons of CCM biases with these models below. Our findings here, in particular the interrelationships between wave flux and mean flow biases, are likely applicable beyond the NCAR CCM. Some aspects of

* The National Center for Atmospheric Research is sponsored by the National Science Foundation.

Corresponding author address: Dr. William J. Randel, National Center for Atmospheric Research, P.O. Box 3000, Boulder, CO 80307-3000.

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the life cycles of midlatitude baroclinic waves in CCM1 are discussed in Randel (1990), including comparison with such features seen in the ECMWF analyses.

With the improvement of models in simulating the atmosphere it has become customary to concentrate on their remaining weaknesses when discussing model circulation statistics. Such knowledge provides the basis for future developments and is vital for experimental design. It is important for researchers to recognize the deficiencies of a model and to consider how the deficiencies might influence the results in a particular application both during the experimental design phase and when interpreting the experimental results. We have followed this approach here by providing less emphasis on the strong points of the model simulation.

In the next section we summarize the algorithms of the CCM and provide references where complete details may be found. We then describe the ECMWF analyses used for comparison, and point out a few aspects of these analyses which are less reliable for model validation. In section 3 we present the results of this study in the following order: temperature biases and thermodynamic balances, wind fields, zonal mean wave diagnostics, stationary waves, and local eddy statistics. In the last section we summarize the major findings.

2. Data and analyses

a. CCM1 simulation

The version of the NCAR CCM primarily analyzed here (CCM1) is based on modifications to an earlier version denoted CCM0. Circulation statistics and model documentation for CCM0 are found in Pitchev et al. (1983), Ramanathan et al. (1983), Malone et al. (1984), Williamson (1983), and Williamson and Williamson (1984). This model existed in two forms, denoted CCM0A and CCM0B, which differed in Fortran code but not in algorithms. Because the algorithms properly define a model, we will denote it here as CCM0, and the discussion applies to results produced by either code. Algorithm changes to CCM0 which resulted in CCM1 are detailed in Williamson et al. (1987); the most important differences are summarized here. The radiation scheme has been substantially modified; a new solar albedo parameterization is used, and parameterizations of H2O, O2, and O3 absorption of solar and terrestrial radiation have been improved. An energy-conserving vertical finite difference formulation is adopted, eliminating a 15 W m$^{-2}$ energy imbalance of CCM0. The horizontal diffusion has been made more scale selective by changing to a $V^4$ form throughout the troposphere. The vertical diffusion and surface energy exchange have been made stability dependent. The possibility of negative moisture pockets has been eliminated, resulting in a significant effect on cloud forecasts in polar regions. The general formulation of precipitation processes themselves and cloud forecast scheme have been only slightly altered, except that precipitation and thus clouds are assumed to occur at 100% relative humidity in CCM1 rather than at 80% as in CCM0. Hoerling et al. (1990) show the effect of these changes on diabatic heating.

One motivation for this study is to compare CCM1 and CCM0 climates in light of known model changes; some comparisons of temperature biases and thermodynamic balances in the two models are shown below. An important result is that although there are clear differences in the simulations, the changes in equilibrium temperatures are not simply related to differences in the forcing fields, because of the complex nonlinearities inherent to the climate system. This is a general result which hampers the search for causes of observed model biases.

The modifications introduced into CCM1 have resulted in mainly local, relatively minor climate changes, although the details of the balances producing the climates may be significantly different. The overall global climates simulated by CCM1 and CCM0 are found to be similar, in particular as measured by the mean wind patterns and stationary and transient wave flux quantities. Although differences exist between CCM1 and CCM0, the differences between the respective models and observations (climate biases) are typically much larger than differences between the two models. Accordingly, the results here are valid overall for CCM0. Circulation statistics for CCM0 are available from the references cited above and are not repeated here.

CCM1 can be used in a perpetual January or July simulation mode, or in a seasonally varying mode in which the specified surface conditions and solar forcing vary with time. It is mainly the seasonal mode that is documented here, for direct comparison with the seasonality revealed in ECMWF analyses. We have made some comparisons with perpetual January and July simulations, and a general result is that model minus observed biases are similar for seasonal and perpetual runs. The seasonal simulation analyzed here was integrated for 3800 days, and the last seven years (or subsets thereof) of twice-daily samples are used for the comparisons. A horizontal spectral truncation at rhomboidal wave 15 (denoted R15—approximately 4.5° latitude by 7.5° longitude) is used, and there are 12 vertical sigma levels (.991, .926, .811, .664, .500, .355, .245, .165, .110, .060, .025, and .009). Data on these sigma levels are interpolated to seven standard pressure surfaces (1000, 850, 700, 500, 300, 200, and 100 mb) for comparison with the ECMWF analyses. It should be noted that CCM0 used only nine vertical $\sigma$-levels (.991, .926, .811, .664, .500, .356, .189, .074, .009) and that some of the upper level differences between the models may be due to the different location of model levels above $\sigma = .500$.

Numerical calculations were performed as follows. Time series of zonal mean winds, temperatures, and wave flux quantities were generated for both ECMWF
and CCM1 data using the CCM Processor (Wolski 1987) and saved in a history tape format. During this procedure CCM1 data on sigma surfaces are interpolated to the standard pressure levels. These were then used to calculate all the time-average zonal mean meridional sections and latitude–time plots. Meridional derivatives for zonal mean quantities were calculated using a $0.95$, $-0.90$, $0.90$, $-0.95$ centered difference formula, and only latitudes up to $\pm 80^\circ$ are plotted in the diagrams. Vertical derivatives were approximated by finite differences between adjacent pressure levels, interpolated back to the standard levels. Horizontally local statistics (sections 3d–e) were calculated entirely with horizontal spectral transforms using the CCM Processor (Wolski 1987).

An extensive compilation of circulation statistics for CCM1.R15 is available in Williamson and Williamson (1987), for both seasonal and perpetual January and July simulations. A discrepancy in the model surface temperature calculation was found subsequent to the compilation of those statistics; new simulations were made with the discrepancy removed, and those form the basis for the study here.1

b. ECMWF analyses

The ECMWF analyses examined here are twice-daily operational products covering the seven years 1980–86. These products are produced via an assimilation of available observations, using model produced forecast fields as a first guess. These operational analyses are referred to throughout this paper as “observations.” Fields available include temperature, three-dimensional velocity, geopotential height, relative and specific humidity, and surface pressure and geopotential. These products were originally archived on 2.5 $\times$ 2.5 degree latitude–longitude grids but have been interpolated to the R15 gaussian grid to be compatible with the history tape format used in the CCM modular processor (Trenberth and Olson 1988a; Wolski 1987). Seven standard pressure levels are available (as listed above). The CCM1 simulation and ECMWF analyses compared here thus have identical spatial and temporal sampling.

The quality of the ECMWF analyses and evolution of the operational system over 1980–86 (resulting in clear changes in some analysis products) are documented in detail in Trenberth and Olson (1988a, b); of the aspects considered here, the analyses of tropical divergent circulations was most dramatically affected (Trenberth and Olson 1988a; note their Figs. 12–13).

c. Observed–model comparisons

The main focus here is to analyze the overall climate simulated by CCM1 and to compare balances in the model to those indicated by the atmospheric analyses. Much of the analysis here documents zonal mean structure and variability, because many of the model biases have an axisymmetric character. January means are used to analyze temperature biases and to allow comparisons with available CCM0 January simulations. Annual means are used for wind and wave flux comparisons, and seasonality is documented via a series of latitude–time diagrams. Comparisons between observed and modeled stationary wave structure in the Northern Hemisphere (NH) winter are also presented. Additionally, horizontally local eddy statistics of the model are compared with observations during NH winter to complement the zonal mean analyses.

Comparisons are made here based upon the sampled climate ensemble framework, i.e., each year of observations or model simulation is treated as an independent estimate of the observed or modeled climate. This methodology takes into account the inherent interannual variabilities when searching for significant model biases. Following Chervin (1981), the null hypothesis that there is no difference between the observed and simulated ensemble averages is tested via the test variate

$$r_1 = \frac{\langle m_{\text{OBS}} \rangle - \langle m_{\text{MOD}} \rangle}{\left( \sigma^2_{(m_{\text{OBS}})} + \sigma^2_{(m_{\text{MOD}})} \right)^{1/2}} \quad (1)$$

Here $\langle m_{\text{OBS}} \rangle$ and $\langle m_{\text{MOD}} \rangle$ are the respective ensemble
averages of observed and modeled variable $m$, and the denominator represents a composite estimate of variance (note that ensemble variances are, in general, not equal for observed and modeled statistics). Chervin (1981) discusses and includes a table of acceptable regions for the null hypothesis. The test variate $t_i$ was calculated along with all the difference maps shown here; the model biases highlighted and discussed here are highly significant according to this test (and are thus true model biases).

d. Wave–mean flow interaction diagnostics

Circulation statistics presented here include diagnostics based on the Eliassen-Palm (EP) flux $\vec{F}$, using both the zonal mean equations and their three-dimensional extension. The EP flux has received widespread use as a diagnostic tool in recent years, following the motivations presented in Edmon et al. (1980). Briefly, the important points of the EP flux $\vec{F}$ are (Plumb 1986):

1) For almost-plane waves on a zonal flow, $\vec{F}$ is a measure of the flux of wave activity parallel to the group velocity of the waves;
2) The EP flux divergence $\nabla \cdot \vec{F}$ is zero for conservative, wavelike disturbances; the observed divergence thus provides a measure of the effects of wave transience and nonconservative effects such as diabatic heating and friction;
3) For quasi-geostrophic flow, $\nabla \cdot \vec{F}$ is proportional to the northward flux of quasigeostrophic potential vorticity; and
4) The quasi-geostrophic momentum and thermodynamic equations may be written in such a way that the only term describing wave-mean flow interactions is an effective zonal force, proportional to $\nabla \cdot \vec{F}$; $\vec{F}$ may thus be interpreted as an effective flux of easterly momentum.

The zonal mean equations used here follow the notation of Dunkerton et al. (1981), using a log-pressure vertical coordinate $z = H \ln(p_0/p)$, with $H = 7$ km. Overbars denote zonal means and primes deviations therefrom. The transformed Eulerian-mean momentum equation is written

$$\frac{\partial \vec{u}}{\partial t} + \vec{F} = D_F + S. \tag{2}$$

Here

$$\vec{F} = 2\Omega \sin \phi \frac{\partial}{\partial \phi} (\cos \phi \vec{u}),$$

$S$ represents momentum sources or sinks, and $D_F$ is the scaled Eliassen-Palm (EP) flux divergence:

$$D_F = \frac{1}{\rho_o a \cos \phi} \left[ \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (\cos \phi F_\phi) + \frac{\partial}{\partial z} (F_z) \right], \tag{3}$$

$F_\phi$ and $F_z$ are components of the EP flux:

$$F_\phi = \rho_o a \cos \phi \left[ -\vec{u} \cdot \vec{v}' + \frac{R}{H} \frac{\partial \vec{u}}{\partial z} \frac{\vec{v}' T'}{N^2} \right] \tag{4a}$$

$$F_z = \rho_o a \cos \phi \left[ \vec{F} \cdot \frac{R}{H} \frac{\vec{v}' T'}{N^2} - \vec{u} \cdot \vec{w}' \right]; \tag{4b}$$

$\vec{v}'$ and $\vec{w}'$ are components of the residual meridional circulation, defined as

$$\vec{v}' = \vec{v} - \frac{1}{\rho_o} \frac{\partial}{\partial z} \left( \frac{\rho_o R}{H} \frac{\vec{v}' T'}{N^2} \right),$$

$$\vec{w}' = \vec{w} + \frac{1}{\rho_o} \frac{\partial}{\partial \phi} \left( \rho_o a \cos \phi \frac{\rho_o R}{H} \frac{\vec{v}' T'}{N^2} \right).$$

The longitude-dependent structure of stationary and transient wave properties are analyzed using the radiative flux of wave activity $\bar{M}_R$ defined in Plumb (1986):

$$\bar{M}_R = \rho_o a \cos \phi \left\{ \frac{1}{2} \left( \langle \vec{b}^2 \rangle - \langle \vec{u}^2 \rangle - \frac{R^2}{H^2} \langle \vec{T}' \rangle \right) - \frac{\langle \vec{u} \vec{w} \rangle}{H \langle \vec{T}' \rangle} \right\}. \tag{5}$$

Angle brackets in Eq. (5) denote time-average values, and tilde variables deviations therefrom.

In many respects $\bar{M}_R$ can be regarded as a three-dimensional extension of the EP flux vectors discussed above. In particular, the magnitude of $\bar{M}_R$ provides a measure of eddy activity, $\bar{M}_R$ is parallel to the wave activity group velocity relative to the mean flow ($\vec{c}_g - \langle \vec{u} \rangle$), and its divergence is proportional to the eddy flux of potential vorticity. It is mainly the former two attributes which are studied here: plots of observed and modeled $\bar{M}_R$ are used to compare the amount of wave activity and its propagation. Note that the zonal mean of $\bar{M}_R$ reduces to the quasi-geostrophic approximation of the two-dimensional EP flux [Eq. (4)]. It should be pointed out here (as stressed in Plumb 1986) that the divergence of $\bar{M}_R$ is not simply related to local generation or dissipation of wave activity, because $\nabla \cdot \bar{M}_R$ does not include the (important) mean flow advection of wave activity (see the extensive discussion in Plumb 1986). Note also that the expression for $\bar{M}_R$ above has assumed that the mean potential vorticity contours make small angles with the zonal direction, an assumption verified for NH winter observations in Plumb (1986).

The calculated structure of $\bar{M}_R$ for stationary waves has the disadvantage that it is phase dependent and contains an oscillatory spatial structure on the scale of one-half the wavelength of the stationary waves (for transient eddies the phase dependency averages out us-
3. Results

a. Temperature biases and thermodynamic balances

Maps of differences between model-simulated fields and observations concisely show where (and when) strongest model biases exist; this is especially the case for zonal mean temperatures, where the meridional structures are similar to the eye even if large biases exist. Figure 1b shows zonal mean temperature differences (CCM1 minus ECMWF) for an ensemble of four (31-day) January means. January is studied here because we wish to make comparisons with CCM0 biases (shown in Fig. 1a), and CCM0 statistics are only available for perpetual January and July simulations (Williamson and Williamson 1984). The regions having differences in excess of ±2 K in Figs. 1a–b have $r_1$-statistics [Eq. (1)] of order 2–20, so that this ensemble of four averages represents true model climate biases (not sampling error). Figure 1b shows CCM1 to exhibit a cold bias throughout most of the troposphere, with values near 4 K in the tropics, and larger values (near 10–20 K) in the lower stratosphere of both polar regions. Such a cold bias throughout the troposphere is a common problem for most low-resolution general circulation models (GCMs; WMO 1988); biases near the tropical tropopause are positive in some models (such as the CCC). Figure 1a shows that CCM0 exhibited a more significant cold bias than CCM1 in the tropical upper troposphere (greater than 8 K), with smaller biases in the lower stratosphere polar regions. There was also a significant cold bias in the NH polar lower troposphere in CCM0, which was remedied in CCM1. This improvement can be traced to the elimination of negative moisture values in the polar regions (which exist in CCM0) and to the resulting clouds that form in CCM1, which emit longwave radiation back to the surface, decreasing the surface cooling of CCM0. The linearized vertical diffusion of CCM0 was converted to a nonlinear stability dependent form in CCM1. In the stable polar regions this eliminated the transfer of energy from the atmosphere to the surface that accompanied the surface radiative cooling in CCM0. Note that CCM1 is actually too warm in the lower stratosphere of both polar regions (positive values in Fig. 1b) and that the low level equator-to-pole temperature gradient is reduced by ~6 K in the model compared to observations.

One advantage of studying numerical simulations is that various equation balances can be evaluated exactly, showing which physical mechanisms are important in the model. To study the various terms contributing to the thermal balances here, terms in the time and zonal mean thermodynamic equation are analyzed, following the work of Boville (1985):

$$\frac{\partial \bar{T}}{\partial t} = \langle \Delta \bar{T}_{rad} \rangle + \langle \Delta \bar{T}_{dyn} \rangle + \langle \Delta \bar{T}_{par} \rangle \approx 0. \quad (6)$$

Here $\Delta \bar{T}_{dyn}$ is the tendency from all dynamical terms, $\Delta \bar{T}_{par}$ is the tendency from subgrid scale physical parameterization schemes included in the model (surface exchange, diffusion, release of latent heat, and dry convective adjustment) and $\Delta \bar{T}_{rad}$ is the tendency from the radiation. The residual to Eq. (6) is near zero for CCM1 (where all the terms can be explicitly evaluated), whereas the $\Delta \bar{T}_{par}$ was calculated here as a residual for CCM0, because all the terms necessary for its calculation were not readily available from the history tape.

Figure 2 shows the three separate temperature tendency terms for CCM0 and CCM1. Net radiative cool-

![Fig. 1](image-url)
Fig. 2. Meridional cross sections of January average thermodynamic forcing terms [Eq. (6)] for CCM0B (left) and CCM1 (right). Shown are $\Delta T_{\text{rad}}$ (top), $\Delta T_{\text{dy}}$ (middle) and $\Delta T_{\text{par}}$ (bottom), with contour intervals of 0.4 K day$^{-1}$.

ing rates (Figs. 2a and 2d) are similar in both models, with the differences almost entirely due to the modified longwave cooling rates and, in the polar regions, to different cloud amounts. Larger changes are seen in the dynamical and subgrid-scale terms, particularly in the tropical upper troposphere, and the changes are nearly compensating. Hoerling et al. (1990) consider some of the components of the subgrid-scale terms in more detail. We consider here only the net and the resulting balances. Although the tendency changes
nearly compensate each other, there are significant changes in upper tropospheric temperatures (Fig. 1), with CCM1 reducing the model bias by approximately a factor of two. Further tests with CCM1 (J. T. Kiehl, private communication) suggest that some form of additional subgrid-scale heating in the tropical upper troposphere (such as additional latent heating or cirrus cloud radiative effects) of order 1 K day\(^{-1}\) would largely eliminate the observed CCM1 bias of order 4 K; similar results were obtained by Albrecht et al. (1987) using a modified cumulus parameterization scheme. Note, however, that in spite of the stronger parameterized heating in the tropical upper troposphere in CCM0 compared to CCM1 (cf. Figs. 2c and 2f), CCM0 has a larger tropical cold bias.

There is a strong increase in cold bias in the lower stratosphere of both polar regions in CCM1 compared to CCM0 (Fig. 1). This is especially intriguing because there are only minor changes apparent in the individual thermodynamic terms in Fig. 2 in these regions (especially in the SH), although the individual forcing terms themselves are very small to begin with. The difference in vertical finite difference approximations between the two models is partially responsible. Williamson (1988a) has shown that the conservative vertical advection adopted in CCM1 yields a colder lower polar stratosphere than the nonconservative form when everything else is left unchanged in CCM0.

Note that the balance of the two models is very different in the lower north polar troposphere. In CCM0, the parameterizations (primarily the linearized vertical diffusion) act to decrease the temperature as energy is transferred from the atmosphere to the surface. This negative tendency by the parameterizations is compensated by heating of the dynamical tendency. This compensation is not seen in CCM1. The stability dependent vertical diffusion becomes small in the stable polar regions cutting off the downward heat flux, and no compensation by the dynamics is induced.

Unfortunately, studying the separate components of the thermodynamic balances gives little insight into the model equilibrium temperature (bias) structure. The strongest $\Delta T_{\text{par}}$ and $\Delta T_{\text{dyn}}$ coincide (with opposite signs) near 0°–10°S in the rising branch of the Hadley cell (Figs. 2b–c and 2e–f); here the balance is between subgrid scale parameterized heating and dynamical and radiative cooling. In the sinking branch of the Hadley cell (near 20°N), a quite different balance exists: parameterized and dynamical heats balance radiative cooling. In spite of these completely different physical balances, a similar temperature bias is seen over 20°N–20°S above 5 km (Figs. 1a–b); so that the bias appears insensitive to the detailed balance of the predominant terms. Overall, the lack of clear association between the spatial structure of model thermal forcings (Fig. 2) and model temperature biases (Fig. 1) shows that the model simulated temperatures are a sensitive result of nonlinear balances between large terms. This fact contributes to the elusiveness of clear causes (or solutions) for these biases.

Additional information can be gained from the horizontal distribution of model temperature biases. Figure 3a shows the CCM1–ECMWF 700 mb temperature differences, revealing pronounced maxima in temperature biases associated with midlatitude continental areas in both hemispheres. In the lowest model levels, the temperature over the oceans is highly constrained by the specified sea surface temperatures through the surface fluxes and vertical diffusion. This effect is strongest in the first two model levels. The 700 mb level is above the third model level but may still feel some of this effect. Inadequate surface exchange and surface process parameterizations are possible causes of the poor simulation in these regions. Strong tropical convective heating regions ($\Delta T_{\text{par}} > 4$ K day\(^{-1}\)) in the model are indicated as hatched areas in Fig. 3a; in these regions, parameterized convective heating is balanced almost entirely by dynamical cooling. The local temperature bias maxima in Fig. 3a are clearly separated from such regions.

Figure 3b shows the 300 mb local temperature bias structure, and here most patterns have a predominantly zonally symmetric character (in contrast to Fig. 3a). Similar behavior is seen in the much larger biases at 200 mb (not shown), suggesting that the cause of the bias is not a strong function of continentality (and that analysis of the zonal mean bias is physically meaningful in the upper troposphere).

Seasonality in zonal mean temperature biases is shown in the latitude–time sections in Fig. 4. These figures were calculated from seven years of low-pass filtered daily data; the digital filter response is similar to that for running monthly means. The 700 mb biases (Fig. 4a) show little seasonality, with maxima over latitudes 20°–40° in both hemispheres, consistent with the January values of Fig. 3. The NH polar region shows seasonal changes in the biases, with positive values in winter and negative values in summer. The 300 mb biases (Fig. 4b) show similar character year round (minima near 40°N and S), with smaller biases in both hemispheres during June–October. The lack of meridional movement of the temperature biases (in light of large seasonality in the Hadley circulation, for instance as shown below) reinforces the conclusion above that the temperature biases are not related in a simple manner to the individual forcing terms.

### b. Wind comparisons

Figure 5 shows the zonal mean zonal wind for ECMWF and CCM1 data and their difference. An ensemble average of seven annual means (both modeled and observed) is shown in Fig. 5; the corresponding annual mean temperature biases are qualitatively similar to those for January shown in Figs. 1–3, except there is less NH–SH asymmetry in the high latitude
polar regions. The CCM1 zonal wind is westerly above the surface in the tropics (Fig. 5b), whereas the ECMWF analyses (Fig. 5a) show easterlies throughout the tropical troposphere. This tropical westerly bias in CCM1 is also found in the CCC model; conversely the ECMWF model has easterlies which are too intense. The tropical bias in CCM1 (Fig. 5c) is small at the surface, increasing with height to a maximum in the upper troposphere, illustrating that it is a thermal wind bias. The approximate zonal-mean thermal wind balance equation at low latitudes is [e.g., Andrews et al. 1987, Eq. (8.2.2)]:

$$ \frac{\partial}{\partial y} \left[ f \frac{\partial u}{\partial z} \right] = R \frac{\partial^2 \bar{T}}{\partial y^2}. $$

The corresponding temperature curvature bias can be seen in low levels near the equator in Fig. 1b, i.e., there
are stronger cold biases near 20°N and 20°S than at the equator. Because of the dependence on temperature curvature at low latitudes, relatively small temperature biases can result in large wind errors; a 700 mb temperature error of only 0.5 K with a scale 10° latitude corresponds to a thermal wind bias near 10 m s⁻¹ in the upper troposphere (similar to that seen in Fig. 5c).

Large CCM1–ECMWF zonal wind differences are seen in high southern latitudes in Fig. 5; the CCM1 SH jet is much narrower than that observed (note that the annual average SH jet is much broader than that seen at any instant, because of its large seasonal movement—see the discussion of seasonality below). Note that this high latitude SH bias is large at the surface and increases further with height (Fig. 5c). The large model wind bias at the surface is associated with the poor simulation of the SH circumpolar sea level pressure trough in CCM1; a similar bias was noted for CCM0, as shown in Pitcher et al. (1983). This poor simulation of the high southern latitude sea level pressure is also observed of the ECMWF model; the CCC is closer to observed values (Xu et al. 1990). An increase with height of the high latitude zonal wind bias is observed in both hemispheres in Fig. 5c, consistent with the reduced low level poleward temperature gradient biases observed in the model (see Fig. 1b).

Figure 6 shows the seasonal evolution of 300 mb zonal mean zonal wind for observations, model simulation, and their difference. The lack of tropical easterlies in CCM1 is evident nearly year-round, as is the weakness of the simulated high latitude SH winds. In fact, even though the seasonal variation of the model zonal wind is reasonably well captured, the biases revealed in Fig. 6c are remarkably constant in time, pointing to their fundamental nature. Note that the model produces relatively good simulations of the winter zonal mean subtropical jets in both hemispheres (the NH is too weak by ≈5 m s⁻¹).

Figure 7 shows January averaged zonal-mean meridional winds, comparing seven years of model data to two years of observations (see discussion in section 2). The CCM1 mean meridional wind in the upper troposphere is approximately only half as strong as that observed; a similar finding was observed for CCM0 (Pitcher et al. 1983, their Fig. 4). Note that only values above 850 mb are plotted in Fig. 7, and there is presumably stronger return flow in the near surface layers not shown.

Seasonality in 200 mb mean meridional winds is shown in Fig. 8, and both observations and model simulation show pronounced maxima in both winter hemispheres. CCM1 values are too weak in the tropics by over 1.0 m s⁻¹, and additionally the equatorward branches of the simulated Ferrel cells (near 40°–50°N–S) are weak. Largest differences are seen in Fig. 8c in NH spring (March–May), although the observed values are subject to uncertainties as discussed in section 2.

Figure 9 shows the December–February mean horizontal wind at 300 mb for observations, simulation, and their difference, plotted as vectors and isolachs. The model captures the main features of the NH jet structure quite accurately, with distinct maxima over eastern North America and over the western Pacific stretching westward to over Africa. The latter jet maximum over the western central Pacific is approximately 10 m s⁻¹ too weak in the model. There are also large local differences over the tropics and in the SH (Fig. 9c). The model wind in the tropics is westerly in the zonal mean (see Fig. 6b), with a strong local maximum in westerlies (over 20 m s⁻¹) over the eastern Pacific.
Observations also show a westerly wind maximum in this region (Fig. 9a), although it is weaker and less extensive than that in the model. Observations show tropical easterlies over most of the eastern hemisphere, particularly over Indonesia, while in this region the model is quiescent with a slight hint of easterlies. The resultant difference map (Fig. 9c) reveals a substantial westerly tropical wind bias over nearly all longitudes, with local maxima over the Pacific and near Africa. The SH model wind bias is strongly zonally symmetric; both model and observations show primarily a zonal character, with the model westerlies not extending far enough toward the pole by approximately 10° (see also Fig. 6). Weickmann and Chervin (1988) show a diagram of differences between 250 mb wind fields from National Meteorological Center operational analyses and CCM0 (their Fig. 4): there is strong similarity between their results and the difference map shown here (Fig. 9c), due to the similarity between CCM1 and CCM0.

The divergent flow in the atmosphere is a relatively small component of the total horizontal wind but is nonetheless important for local vorticity balances and generation of rotational flows (e.g., Sardeshmukh and Hoskins 1988). Figure 10 shows the observed and modeled 200 mb divergent horizontal wind vectors, along with the divergence superimposed on each diagram. Note that only the most recent two years of ECMWF data are included in Fig. 10, due to data problems as discussed in section 2, and even these values should be viewed cautiously.

The overall spatial character of the modeled divergent flow is similar to the observations, with localized regions of strong divergence over South America, Africa, and Indonesia near 10°S. The model divergence patterns are more intense locally but smaller in spatial extent than those observed over the Indian and western Pacific oceans, and furthermore there are some spatial biases over northern South America and northern Africa. Note that in the observations the regions of strong divergence occur near the same latitude (≈10°S), whereas in the model there is some latitudinal shift between maxima; therefore, the resulting zonal mean divergence is substantially smaller in the model, as is the zonal mean meridional wind (Figs. 7–8).

c. Zonal mean wave diagnostics

The correct simulation of wave statistics constitutes a stringent test on the ability of CCM1 to model the atmosphere. Time means of eddy kinetic energy and eddy fluxes of heat, momentum, and potential vorticity provide signatures of the transient features in the model which are more fundamental than geopotential height variance maps alone, and can be related to the wind and temperature fields via wave–mean flow interaction diagnostics (see section 2d).
Meridional cross sections of wave statistics are calculated here from daily zonal covariance calculations and thus have contributions from both stationary and transient features. The following meridional diagrams present annual averages and are thus predominantly the result of transient waves. December–February average stationary and transient wave fluxes are discussed separately below.

Figure 11 shows meridional sections of the annual mean eddy kinetic energy $\frac{1}{2}(u'^2 + v'^2)$ for observations, simulation, and their difference. The model eddy kinetic energy is substantially lower than that observed in midlatitudes, with the central midlatitude maxima shifted upward and equatorward (particularly in the SH). A lack of midlatitude kinetic energy is a common
characteristic of climate models of similar resolution, observed for both the CCC and ECMWF models, for example, and a clear cause is not known. On the other hand, CCM1 has an excess of kinetic energy in the tropics compared to that observed, primarily in the upper troposphere (Fig. 11c). A similar signature is seen in the eddy temperature variance (or available potential energy) comparisons (not shown); the model is deficient in middle and high latitudes, with excessive variance in the tropics. The excessive wave variance in the tropics may be related to the lack of low latitude easterlies in the model (Figs. 5–6). Analyses of wave propagation in idealized models (e.g., Webster and Holton 1982; Karoly 1983) show that equatorial easterlies inhibit the meridional propagation of extratropically forced waves; CCM1 may thus have excessive tropical variance because it does not correctly inhibit the meridional propagation of such features.

Two fundamental quantities for the zonal mean circulation are the poleward eddy fluxes of heat \((\bar{v} T)\) and momentum \((\bar{u} T)\); these are shown in Fig. 12 for observations and simulation, along with their respective differences. The CCM1 poleward heat fluxes in Fig. 12b are in good agreement with observations (Fig. 12a) in the lower and middle troposphere. The modeled upper-tropospheric heat fluxes (above 300 mb) are somewhat weaker than those observed, and a spatially similar bias pattern is seen in both hemispheres (the model fluxes are too far equatorward at 100 mb). The poleward momentum fluxes shown in Fig. 12 reveal large biases; modeled maximum values are approximately a factor of two too small throughout the troposphere, with substantially larger absolute errors in the SH.

Seasonality in poleward eddy heat flux at 700 mb is shown for observations and model simulation in Fig. 13. There is excellent agreement throughout the year; note the stronger contribution from stationary waves in the NH winter in Fig. 13 (transient values are similar between NH and SH winters; e.g., see Trenberth and Olson 1987a). The upper tropospheric heat flux biases (whose seasonality is not shown) are present year round but are largest during NH winter and SH spring seasons. Increased biases during these periods probably result from the lack of model resolution in the stratosphere and the resulting poorly simulated upper-level planetary waves in the model (see Boville and Cheng 1988).

Latitude–time sections of observed and modeled poleward momentum flux at 300 mb are shown in Fig. 14. The deficit of momentum flux in the model is large throughout the year in the SH near 30°–40°S. Somewhat smaller differences are seen in the NH, with rel-

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**Fig. 8.** Latitude–time sections of the 200 mb zonal mean meridional wind for 1985–86 ECMWF analyses (a), CCM1 simulation (b), and CCM1 minus ECMWF difference (c). Contour interval is 0.5 m s⁻¹, and zero contours are omitted in (c).
Fig. 9. Horizontal projections of the December–February mean 300 mb horizontal wind vectors and isotachs. (a) ECMWF analyses, (b) CCM1 simulation, and (c) CCM1 minus ECMWF differences. Contour interval is 10 m s$^{-1}$ in (a) and (b) and 5 m s$^{-1}$ in (c).
ative biases comparable to those in the SH during NH spring.

Figure 15 shows meridional cross sections of observed and modeled vertical momentum flux $u''w''$. Although not of primary importance in the zonal momentum budget (contributing zonal accelerations in the troposphere of order 0.5–1.0 m s$^{-1}$/day), this diagnostic is of interest here because of the dramatically different observed and modeled signatures in extratropics. Observed (Fig. 15a) and modeled (Fig. 15b) values both show downward flux in the tropical upper troposphere, with substantially weaker observed values (although this is a poorly measured quantity in the tropics). However, the extratropical signatures are strikingly different; the observed patterns exhibit clear positive (upward) maxima in midlatitudes, whereas in the model such fluxes are absent. Although this lack of vertical momentum flux in the model is not as im-
Figure 16 shows EP flux diagrams for observations, model simulation, and their difference. As discussed in Edmon et al. (1980), the arrows indicate components of the EP flux or wave group velocity [with horizontal and vertical components given by (4a) and (4b)], while the contours are the calculated $D_F$ values [Eq. (3)]. Full primitive equation estimates of EP flux quantities are shown in Fig. 16, although ageostrophic contributions were found to be small. Both observed and modeled $D_F$ maxima of order $-5$ m s$^{-1}$/day are observed in the upper troposphere but with an important difference in their patterns; the model simulated wave driving effectively reaches deeper into the upper troposphere than that observed. The result is that the waves in the model exert a substantially stronger drag on the upper tropospheric zonal mean flow than do those observed, particularly in the SH; differences of order 1 and $3$ m s$^{-1}$/day are seen, respectively, in the NH and SH in Fig. 16c. The cause of this increased wave deceleration in the model is not increased wave activity generated baroclinically in low levels (note the accurate simulation of lower level $v^T$ in the model, Figs. 12–13) but a lack of equatorward propagation of wave activity in the upper troposphere of the model (decreased poleward momentum flux, i.e., Fig. 14). This is clearly shown in the EP flux difference diagram in Fig. 16c: the difference in upper tropospheric wave driving is associated with horizontal EP vectors (which are proportional to $u \, v$).

Comparison of the CCM1–ECMWF differenced EP flux diagram (Fig. 16c) with the ECMWF–CCM1 zonal wind difference (Fig. 5c) reveals very similar patterns of differences in the high latitude SH, with less agreement in the extratropical NH. This agreement suggests that the large high latitude zonal mean wind biases in CCM1 (particularly in the SH) are related to the excessive high latitude wave driving in the model, which is in turn related to the lack of poleward wave momentum flux. Furthermore, Fig. 16c suggests that the tropical zonal wind bias in CCM1 is not directly a result of wave driving, because both the observations and model exhibit small $D_F$ in the tropics.

Note that the above statements relating time mean $D_F$ biases to time mean $\tilde{u}$ biases is only qualitative. The time mean of Eq. (2) is, to a good approximation,

$$\langle -fv^* \rangle = \langle D_F \rangle;$$

i.e., the time tendency term is negligible in the long-term average. This is the zonal momentum equation analog to Eq. (6). Thus time mean (equilibrium) $\tilde{u}$ biases are not simply related to time mean $D_F$ biases, although the good agreement in spatial patterns clearly points to the role of wave driving in determining $\tilde{u}$.

A localized wave driving bias will also have a zonal mean temperature bias associated with it, with the spatial structure of a quadrupole in the meridional plane.
Fig. 12. Meridional cross sections of annually averaged zonal mean poleward eddy heat flux (left) and poleward eddy momentum flux (right). ECMWF analyses (top), CCM1 simulation (middle), and CCM1 minus ECMWF differences (bottom). Contour intervals are 4 K m s$^{-1}$ and 10 m$^2$ s$^{-2}$, respectively.

(high latitude warming and low latitude cooling below the forcing region, and reversed meridional maxima above—see for example Garcia 1987, his Fig. 12). This induced temperature change maintains thermal wind balance with the zonal wind change. For the midlatitude upper tropospheric wave driving bias in CCM1 (Fig. 16c), this pattern corresponds to warmer temperatures (smaller cold biases) in the polar lower troposphere and subtropical upper troposphere, along with colder temperatures (larger cold biases) in the polar
upper troposphere and subtropical lower troposphere. This is exactly the sense of the CCM1 temperature bias patterns seen in Fig. 1b (although again it is difficult to directly relate time average tendencies to equilibrium temperature values).

d. Stationary waves

Figure 17 shows EP flux diagrams for observed and modeled stationary waves, with averages calculated over December–February. Overall the model EP flux diagram is very similar to that calculated for the observed stationary waves, and a difference map is not included in Fig. 17 because of the small differences. The observed model deficiency in transient wave poleward momentum flux (transient waves are the predominant contributors in all the previous annual mean diagnostics) is not apparent in the modeled stationary waves. The model results (Fig. 17b) show less wave driving in the high latitude lower stratosphere than that observed, which is not surprising given the lack of model resolution in the stratosphere. Similar high lat-
Fig. 15. Meridional cross sections of (a) ECMWF analyses and (b) CCM1 simulated annually averaged vertical eddy momentum flux $u'w'$. Contour interval is $10 \times 10^{-3}$ m$^2$/s$^2$.

Fig. 16. Annual average EP flux diagrams for (a) ECMWF analyses, (b) CCM1 simulation, and (c) CCM1 minus ECMWF difference. The EP vectors follow standard scaling (see Dunkerson et al. 1981). Contours of $D_F$ [Eq. (3)] use contour interval of 1 m s$^{-1}$/day. The vectors in (c) are scaled twice as large as in (a) and (b).
accurately simulated, with strong equatorward propagation east of Asia and downstream and poleward \( \vec{F} \) vectors over western North America. The overall excellent agreement in Figs. 17–18 shows the high quality of the modeled stationary wave structure.

e. Horizontally local transient eddy statistics

Our remaining analyses concentrate on comparing horizontally local transient wave characteristics of the observed and modeled atmospheres. Because the most pronounced zonal asymmetries occur in the NH winter, the analyses here focus on December–February statistics. Transient wave kinetic energy \( \frac{1}{2}(\langle u^2 \rangle + \langle v^2 \rangle) \) and \( \vec{M}_R \) vector diagnostics [Eq. (5)] are shown in Fig. 19. The \( \vec{M}_R \) scaling in Fig. 19 is identical to that of \( \vec{F} \) in Fig. 18, so that the relative importance of stationary versus transient features during NH winter can be compared. The transient kinetic energy diagrams in Fig. 19 show a substantial deficit in extratropics of both hemispheres in CCM1 (as shown for zonal means in Fig. 11). There are two strong regions of vertical \( \vec{M}_R \) flux (transient wave poleward heat flux) in the NH in Fig. 19, with overall good agreement between observed and modeled patterns; however, the model values over 0°–120°W are somewhat weak—note the stationary wave is also somewhat weak here (Fig. 18). The SH exhibits an overall zonally symmetric character in both model and observations. There are substantial differences in the horizontal \( \vec{M}_R \) vector components in Fig. 19, primarily in the SH. The difference is concentrated in the meridional component, denoting a lack of poleward momentum flux in the model (cf. Fig. 14); it is furthermore seen to exhibit a zonally symmetric character.

It has been common practice to study characteristics of transient waves by filtering the circulation statistics according to fluctuation time scale. Many studies use the low-pass (greater than 10 day period) and band-pass (2.5–6 day period) filters introduced by Blackmon (1976), and similar time filters were applied to the data here (11-point filters, designed to have similar spectral responses to Blackmon’s 31-point filters, were applied to the 90-day December–February data). Figure 20 shows the observed and modeled \( \vec{M}_R \) diagrams constructed from low-pass filtered data. Note the contour levels and vector scaling are two-thirds of those in Fig. 19.

Observed regions of low-pass filtered vertical \( \vec{M}_R \) flux (contours in Fig. 20) show patchy maxima over much of the NH, with largest values over the North Pacific and North Atlantic ocean regions. The model exhibits similar NH patterns overall, with somewhat lower magnitudes. The horizontal components of \( \vec{M}_R \) suggest primarily upstream and equatorward propagation, with local maxima equatorward of the vertical component maxima. Note that the upstream propagation implied in Fig. 20 is with respect to the mean flow. Plumb (1986) has shown that the eastward mean-flow advective term dominates and that the net wave activity flux is eastward. Overall the horizontal vectors are similar between observed and modeled data in Fig. 20, although somewhat weaker for CCM1. The good qualitative agreement in \( \vec{M}_R \) shows that, at least statistically, low frequency features in the model exhibit characteristics similar to those observed. The model low-pass transient kinetic energy (not shown) is approximately 30% too low compared to observed values.

In Fig. 21 \( \vec{M}_R \) diagrams calculated for observed and modeled band-pass statistics are shown. Contours and vectors are scaled one-third of those in Fig. 19 and one-half of those in Fig. 20. The vertical components of \( \vec{M}_R \) show two distinct maxima in the NH for both observations and model simulation; these indicate regions of intense high frequency wave activity, identified as so-called storm track regions. The Atlantic maximum in the model is somewhat weaker than observed, similar to the low-pass and stationary components.
ECMWF 300 mb STAT. WAVE $Z'$

ECMWF STAT. WAVE $\vec{F}_x$

CCM1 300 mb STAT. WAVE $Z'$

CCM1 STAT. WAVE $\vec{F}_x$

Fig. 18. December-February mean stationary wave statistics from ECMWF analyses (top) and CCM1 simulation (bottom). (left) Zonally asymmetric geopotential height at 300 mb (contour interval is 50 gpm, zero contours omitted) and (right) three-dimensional $\vec{F}_z$ diagrams. The $\vec{F}_z$ diagrams show the 300 mb horizontal components as vectors (with a vector of length 10° longitude representing 80 m² s⁻²), along with the 700 mb vertical component contoured (contour interval is 0.15 m² s⁻²).
Fig. 19. December–February average transient wave statistics from ECMWF analyses (top), CCM1 simulation (middle), and CCM1 minus ECMWF difference (bottom). (left) Transient wave kinetic energy (contour interval is 50 m$^2$ s$^{-2}$) and (right) $M_z$ diagrams [Eq. (5)]. The vector and contour scaling in the $M_z$ diagrams is identical to that for the stationary waves in Fig. 18. Transient kinetic energy differences in (c) above 100 m$^2$ s$^{-2}$ are shaded.
shown above (Figs. 18 and 20). Furthermore, the SH exhibits weaker vertical $\tilde{M}_R$ flux in Fig. 21, and the region of largest band-pass difference (south of Africa) is also seen in the total (Fig. 19). There is, however, a much more fundamental difference in the horizontal $\tilde{M}_R$ components between observed and modeled band-pass statistics: the observations exhibit strong downstream and equatorward propagation (strongest in the storm track exit regions) that is virtually absent in the model. Comparison with the low-pass statistics in Fig. 20 suggests that the absence of poleward momentum flux in the model is primarily associated with higher frequency baroclinic waves; the overall deficit of momentum flux is much larger in the SH because of the relatively more important role of transient baroclinic waves in that hemisphere. The model band-pass kinetic energy (not shown) is only about 50% of that observed.

4. Summary and discussion

Detailed comparison of the climate simulated by CCM1 with 12 levels and R15 spectral resolution with that revealed by ECMWF operational analyses over 1980–86 gives the following main results:
1) There is a general cold bias throughout the troposphere in CCM1; largest biases are found over extratropical land areas in low levels, while the biases are predominantly zonally symmetric in the upper troposphere. The spatial structure of individual thermodynamic forcing terms gives little insight into the cause of the equilibrium temperature biases. The temperature biases show negligible meridional variation with season while the individual forcing terms (as associated with the Hadley circulation, for example) show strong seasonal movement.

2) CCM1 exhibits a strong westerly zonal wind bias in the tropical upper troposphere; locally the bias is largest over Indonesia as a lack of easterlies and in the eastern Pacific Ocean and over Africa as excessive westerlies. There is a pronounced lack of high latitude SH westerlies year round in the model. The seasonal evolution of zonal mean wind in the SH is quite different in the model from that observed: the summer jet is notably weak and is displaced equatorward, while the core is reasonably positioned in winter.

3) The upper tropospheric divergent flow in the
model is qualitatively similar to that observed (using only two years of analyzed data); however, modeled divergent patterns are stronger and more spatially localized than those from analyzed data. The zonal mean meridional circulation is roughly a factor of two too small in the model, primarily because the local divergence maxima do not occur at the same latitude in the model and thus are washed out in the average.

4) NH winter stationary waves in the model are in overall good agreement with observations. Transient waves in the model are too weak in extratropics and too intense in the tropics, as measured by their average kinetic energy. Extratropical poleward heat fluxes are in good agreement with observations, but poleward momentum fluxes are severely underestimated in the model, particularly in the SH. The resulting EP flux divergence (or potential vorticity flux) is substantially stronger in the model upper troposphere than that observed. These similarities and differences are clearly seen in the signatures of midlatitude baroclinic wave life cycles in CCM1 (Randel 1990). Partitioning of the various quantities reveals that the lack of transient kinetic energy and poleward momentum flux is relatively larger for higher frequency transients. The predominance of transient waves in the SH is thus consistent with the relatively larger momentum flux biases there.

There is a highly interrelated, coupled nature to the biases in different variables in CCM1. Thermal wind balance is seen between the temperature and zonal wind biases (Figs. 1 and 5). There is good qualitative agreement between the patterns of wave driving bias (Fig. 16c) and zonal wind bias (Fig. 5c) in extratropics; this suggests that the excessive wave driving in the model is related to the lack of high latitude westerlies in the SH. This comparison furthermore suggests that large-scale wave driving is not the main cause of the tropical zonal wind bias in the model. The balanced nature of the biases, along with an imprecise association between time average tendencies and equilibrium values, makes the search for causes of climate model biases an imposing task.

The lack of poleward momentum flux in the model (and resulting bias in wave driving) is an intriguing result that prompts further discussion. Note that the decreased model eddy momentum flux convergence is in balance with the decreased model Hadley circulation via the Eulerian mean momentum equation \( \langle \mathbf{F} \rangle = \langle \partial \mathbf{v} \rangle \times \langle \mathbf{u} \rangle \). The coupled nature of these fields illustrates the problem of finding the cause(s) of biases. For example, a poorly simulated Hadley circulation \( \mathbf{F} \) demands that the model momentum flux convergence be small; alternatively, a lack of momentum flux in the model (for whatever reason) mandates a reduced Hadley circulation (which can then influence the tropical temperature structure, etc.).

One possible cause of the decreased eddy momentum flux is the coarse meridional resolution used here (approximately 4.5° latitude), i.e., meridional wave propagation during the decay phase of baroclinic life cycles is not accurately simulated. A strong argument against this proposition is found in the results of Meehl and Albrecht (1988), who incorporate a modified cumulus parameterization scheme in a similar R15 global spectral model and find dramatic increases in the eddy momentum flux in the SH (in addition to modified zonal wind and temperature fields). This demonstrates that spectral resolution is not the sole cause of the reduced momentum fluxes revealed here.

We should emphasize here that in this study we have concentrated on the weaker points of the simulation. Overall, the model does a very good job in simulating many aspects of the atmosphere. Nevertheless, knowledge of the remaining deficiencies is essential to provide a base for further improvements in the model and for successful application of the model to important problems.

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