Evaluation of a CCSM3 Simulation with a Finite Volume Dynamical Core for the Atmosphere at 1° Latitude × 1.25° Longitude Resolution

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ABSTRACT

A simulation of the present-day climate by the Community Climate System Model version 3 (CCSM3) that uses a Finite Volume (FV) numerical method for solving the equations governing the atmospheric dynamics is presented. The simulation is compared to observations and to the well-documented simulation by the standard CCSM3, which uses the Eulerian spectral method for the atmospheric dynamics. The atmospheric component in the simulation herein uses a 1° latitude × 1.25° longitude grid, which is a slightly finer resolution than the T85-grid used in the spectral transform. As in the T85 simulation, the ocean and ice models use a nominal 1-degree grid. Although the physical parameterizations are the same and the resolution is comparable to the standard model, substantial testing and slight retuning were required to obtain an acceptable control simulation. There are significant improvements in the simulation of the surface wind stress and sea surface temperature. Improvements are also seen in the simulations of the total variance in the tropical Pacific, the spatial pattern of ice thickness distribution in the Arctic, and the vertically integrated ocean circulation in the Antarctic Circumpolar Current. The results herein demonstrate that the FV version of the CCSM coupled model is a state-of-the-art climate model whose simulation capabilities are in the class of those used for Intergovernmental Panel on Climate Change (IPCC) assessments. The simulated climate is very similar to that of the T85 version in terms of its biases, and more like the T85 model than the other IPCC models.

1. Introduction

The Community Climate System Model (CCSM) is a global coupled ocean–atmosphere modeling framework designed to simulate the climate of the Earth. It is a comprehensive General Circulation Model that consists of complex submodels for the atmosphere, ocean, ice and land. The fidelity of the simulations using the most recent version CCSM3 was described in detail in the June 2006 special issue of the Journal of Climate. The global-scale overview of the present-day climate simulation with the standard configuration is given by (Collins et al. 2006c) in the same special issue.

The atmospheric component of the coupled model is now called the Community Atmosphere Model (CAM). In the earlier versions, a variety of spectral and finite difference methods were available to solve the transport of water vapor, temperature and momentum variables (Williamson 1983; Williamson and Rasch 1994). The “spectral” dynamical core uses a combination of spectral finite difference discretizations to provide solutions for momentum, and temperature, and a semi-Lagrangian technique for trace constituents (in this case water substances). An alternate “semi-Lagrangian” formulation was developed in which the semi-Lagrangian method was used for the transport of both tracer constituents and dynamical variables (Williamson and Olson 1994). In this case, the dynamical component of the atmospheric model is referred to as the “semi-Lagrangian dycore.” This improved the consistency of the transport formulation. This core is not inherently conservative so a posteriori “mass fixer” is employed to enforce conservation.

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A third dynamical core uses a terrain-following Lagrangian control volume (finite volume) formulation (Lin 1997, 2004; Lin and Rood 1996, 1997) and treats the dynamical and tracer constituents uniformly and more consistently. This dynamical core is incorporated as an option in CAM, and is referred to as the “FV dycore.” In this dycore, a two-dimensional conservative semi-Lagrangian advection scheme is used for transport within the Lagrangian control volume, which helps to circumvent the polar singularity problem. The FV method is also free of high wavenumber distortions (Gibbs oscillations) near mountainous regions when compared with the spectral methods (Lin and Rood 1997).

Both the spectral and FV dynamical cores are well established to represent the important long-scale waves of atmospheric motion accurately. However, the approach to the numerical solutions can be viewed as representative of two extremes. In the spectral dycore, global basis functions represent the spatial structure of the field variables; hence, information from both upstream and downstream influences a particular point in space. The finite volume scheme is designed to keep the calculation of spatial derivatives local and to use, as much as possible, only information from upstream. These design features are intended to assure that the physical relationship between different parameters is well represented. The spectral method, from the point of view of numerical approximation, is optimal. As discussed in Lin (2004), not only do the two approaches contrast local versus global approximation, but also contrast physical and mathematical approaches to solving the equations of fluid dynamics.

While the two schemes represent well-resolved spatial scales accurately, their treatment of the small scales is very different. The spectral scheme is contaminated with Gibbs phenomena that manifest themselves as ripples near small-scale features such as topography and frontal zones. Diffusive filters are used to remedy this rippling effect. The finite volume scheme uses diffusive flux limiting to reduce the rippling, and for the most part, assures that the model parameters never exceed the extreme maxima and minima that are physically realizable. As originally designed (Lin and Rood 1996, 1997), it was anticipated that giving priority to physical consistency and the local representation of dynamical systems could have important consequences for climate and weather modeling.

The FV dycore has been installed in several modeling environments. Within the CCSM environment, the tracer transport properties of the spectral, semi-Lagrangian, and finite volume numerical methods in CAM3 were compared in a recent study (Rasch et al. 2006). They found that the FV core is, unlike spectral and semi-Lagrangian, conservative and less diffusive, and more accurately maintains the nonlinear relationships among variables required by thermodynamic and mass conservation constraints. Their findings favor the use of the FV core for tracer transport and atmospheric model dynamics.

The latest version of the Geophysical Fluid Dynamics Laboratory (GFDL) coupled climate model, CM2.1, uses the FV dycore at $2^\circ$ latitude ($\text{lat}) \times 2.5^\circ$ longitude ($\text{lon}$) resolution (Delworth et al. 2006). When compared to the earlier version, CM2.0, that uses a finite difference method, the FV core leads to an improved simulation of the midlatitude westerly winds. There is also dramatic improvement in the simulation of the Southern Ocean in CM2.1 relative to CM2.0 (Gnanadesikan et al. 2006).

At the National Aeronautics and Space Administration (NASA), the FV dycore is coupled to the physics module of NCAR CCM3 and the Community Land Model (CLM) to produce the “fvGCM” model. This model has been used as a global weather forecasting model at higher resolutions (Atlas et al. 2005; Shen et al. 2006a,b) and in many stratospheric (e.g., Douglass et al. 2003) and tropospheric (Li et al. 2002) chemistry-transport studies. The NASA fvGCM is also used as a climate model to study regional climate change at a resolution of $1^\circ$ lat $\times 1.25^\circ$ lon (Coppola and Giorgi 2005; Coppola et al. 2005).

This paper will investigate the sensitivity of long climate simulations to two dynamical cores, the spectral and the FV dycores. On the one hand, the differences in many of the time and spatially averaged variables that are used in the evaluation of climate models are expected to be only minimally impacted by the atmospheric dynamical core. On the other hand, there are more subtle measures of the global circulation, for example age of air, which are sensitive to both the generation of small-scale structure in the model as well as to the dissipation of both large-scale and small-scale waves. Further, there are many mechanisms and attributes of climate models that are local in their determination. A clear example is the representation of precipitation in the mountainous terrain of the western United States, where Rossby-scale waves transporting moisture from the Pacific Ocean cause precipitation on the up slope of the mountains. There are definitive rain shadows on the down slope. Ultimately these local processes are critical to robust climate predictions and the confident use of information from climate predictions.

With the experience of the numerical experiments described above and the anticipated benefits of the finite volume dynamical core, the atmospheric component of the next version of the CCSM is likely to use the
FV dycore. It is of interest, therefore, to examine the sensitivity of the simulation to this change in dycores. Here, we focus on establishing the fidelity of the climate simulations using standard temporally and spatially averaged parameters. This is a necessary part of the evaluation, and we do not expect (or find) any zero-order changes in standard climate diagnostics.

Our investigation of the local processes revealed that the Arctic sea ice is sensitive to changes in the local atmospheric dynamics. We also found positive impact of the atmospheric dynamical core on local processes, especially, the pattern of continental precipitation. These improvements in the simulation of local climate will be discussed in detail in future papers.

The horizontal resolution in the configuration of the spectral dynamical core reported in Collins et al. (2006c) is T85. As implemented, this corresponds to a computational grid of approximately 1.4°. The CCSM3 simulation documented here uses the FV dycore at a 1° resolution, which is slightly finer than T85 and is the first CCSM3 simulation using the FV dycore at this resolution.

The resolution sensitivity of CAM3 and CCSM3 for the spectral dycore version of the model was investigated by (Hack et al. 2006). They found that CAM3 shows robust systematic improvements with higher horizontal resolution for a variety of features, most notably associated with the large-scale dynamical circulation. We assert that since the resolution in our simulation and the standard spectral T85 configuration is comparable (D. Williamson 2007, personal communication), any robust differences in our CCSM3 simulation are likely due to the change in the dycore.

In summary, the main goal of this paper is to document the global-scale simulation characteristics of the FV version of the CCSM3 model. A brief description of the model and our simulation is given in the next section. The FV simulation is compared to the standard CCSM3 simulation (Collins et al. 2006c) and observations in section 3. In the last section, some concluding remarks are given.

2. Model and the simulations

A special issue of Journal of Climate (June 2006) was dedicated to the documentation of climate simulations from the standard spectral configuration of CCSM3. CCSM3 is the third-generation coupled model. Its predecessors are Climate System Model version 1 (CSM1; Boville and Gent 1998) and Community Climate System Model version 2 (CCSM2; Kiehl and Gent 2004). The new model CCSM3 has updated model components: the Community Atmosphere Model version 3 (CAM3; Collins et al. 2006a,b), the Community Land Surface Model version 3 (CLM3; Dickinson et al. 2006), the Community Sea Ice Model version 5 (CSIM5; Briegleb et al. 2004), and Parallel Ocean Program version 1.4.3 (POP; Smith and Gent 2004). CCSM3 has been designed to produce simulations with reasonable fidelity over a wide range of resolutions and with a wide variety of atmospheric dynamical frameworks (Hack et al. 2006; Rasch et al. 2006; Yeager et al. 2006).

a. The FVx1 configuration

For the simulations presented here, we use the Finite Volume (FV) dynamical core for the atmospheric component. The horizontal resolution is 1° lat × 1.25° lon. Hereafter, this configuration will be referred to as FVx1. The horizontal resolution of the spectral simulation T85 (Collins et al. 2006c) in physical space is about 1.4°. There are 60% more grid points in the FVx1 configuration than in the T85 simulation. This FVx1 configuration necessitated slight retuning of the model. This means that either a new error was introduced by the FV dynamical core or an error cancellation in the spectral model was removed, or perhaps the slight difference in resolution is playing a role; it is difficult to identify the cause in contemporary climate modeling.

The tuning was performed on the uncoupled atmospheric model CAM3 using prescribed observed climatological sea surface temperatures and sea ice. In this iterative tuning procedure, we brought the meridional distribution of the zonal-mean shortwave and longwave cloud radiative forcing into reasonable agreement with Earth Radiation Budget Experiment (ERBE; Barkstrom and Smith 1986) observations. The global- and annual-mean net energy flux at the top of the atmosphere is also reduced to within a few tenths of 1 W m⁻². This is a conventional approach to climate model tuning, and the changes to parameters in the cloud and convection parameterizations are given in Table 1. The change in these parameters is comparable to those in tuning other configurations at different resolutions of the spectral dynamical core.

In implementing the FV dycore, the specification of the terrain filter is changed. The topography used in the spectral dycore was fit with the spectral functions used in the dynamical core and filtered to reduce the obvious presence of Gibbs phenomena. The spectral filtering is performed in the spectral space, and is dispersive in the physical space. This smoothing yields a topography that is significantly smoother than the topography that would be consistent with the numerical grid of the finite volume model (S.-J. Lin 2006, personal communication). Therefore, we adopt an “optimal” two-grid filter in the physical space. It filters only the grid-scale (twice the grid length) waves. It uses a third order conservative and monotonic remapping in the physical space. A
Table 1. Dycore-dependent parameters for CCSM3 T85 and FVx1 simulations (Collins et al. 2006b).

<table>
<thead>
<tr>
<th>Parameter description</th>
<th>T85</th>
<th>FVx1</th>
</tr>
</thead>
<tbody>
<tr>
<td>Threshold for autoconversion of warm ice (kg kg(^{-1}))</td>
<td>(4.0 \times 10^{-4})</td>
<td>(2.0 \times 10^{-4})</td>
</tr>
<tr>
<td>Threshold for autoconversion of cold ice (kg kg(^{-1}))</td>
<td>(1.6 \times 10^{-6})</td>
<td>(1.8 \times 10^{-6})</td>
</tr>
<tr>
<td>Stratiform precipitation evaporation efficiency parameter [((\text{kg m}^{-2} \text{s}^{-1})^{1/2})]</td>
<td>(5.0 \times 10^{-6})</td>
<td>(5.0 \times 10^{-6})</td>
</tr>
<tr>
<td>Convective precipitation evaporation efficiency parameter [((\text{kg m}^{-2} \text{s}^{-1})^{1/2})]</td>
<td>(1.0 \times 10^{-6})</td>
<td>(1.0 \times 10^{-6})</td>
</tr>
<tr>
<td>Minimum relative humidity threshold for low stable clouds</td>
<td>0.91</td>
<td>0.91</td>
</tr>
<tr>
<td>Minimum relative humidity threshold for high stable clouds</td>
<td>0.70</td>
<td>0.77</td>
</tr>
<tr>
<td>Parameter for shallow convection cloud fraction</td>
<td>0.07</td>
<td>0.04</td>
</tr>
<tr>
<td>Parameter for deep convection cloud fraction</td>
<td>0.014</td>
<td>0.10</td>
</tr>
<tr>
<td>Top of area defined to be midlevel cloud (Pa)</td>
<td>(2.5 \times 10^{-2})</td>
<td>(75.0 \times 10^{-2})</td>
</tr>
<tr>
<td>Shallow convection precipitation production efficiency parameter (m(^{-1}))</td>
<td>(1.0 \times 10^{-4})</td>
<td>(1.0 \times 10^{-4})</td>
</tr>
<tr>
<td>Deep convection precipitation production efficiency parameter (m(^{-1}))</td>
<td>(4.0 \times 10^{-3})</td>
<td>(3.5 \times 10^{-3})</td>
</tr>
</tbody>
</table>

A series of experiments was performed with the finite volume model to determine the sensitivity to the two types of terrain-filtering specifications. From a global perspective, changes in surface pressure and upper-air geopotential were small. Changes in the pattern of precipitation and the moisture flux, for instance, in the North American monsoon were readily discernible. A consistent specification of topography is an integral part of the dynamical core.

In the first tests with the coupled model there was excessive marginal sea ice in the Arctic and in the North Atlantic south of Greenland during the winter. The investigation of the sea ice bias contributed to a series of independent experiments at NCAR using the 2° lat × 2.5° lon FV dycore. These experiments led to turning off the nonlinear Smagorinsky viscosity (Smagorinsky 1993) option in the anisotropic formulation of horizontal diffusion for momentum components in the ocean model (Smith and Gent 2004), and turning on a linear option wherein the viscous tensor is linearly related to the velocity gradients (P. Gent 2006, personal communication). The Smagorinsky viscosity tends to damp the ocean currents strongly near curved coastlines, and thus it weakens the West Greenland Current into the Davis Strait significantly. With the linear option for the ocean viscosity, the West Greenland Current became stronger and the sea ice was improved in the 2° lat × 2.5° lon FV configuration of the model and the improvements carried over to FVx1. A detailed analysis of the sensitivity of the simulated ocean circulation to the choice of the ocean viscosity formulation is beyond the scope of this paper.

All other aspects of the model are as in the T85 configuration discussed by Collins et al. (2006c). We performed the FVx1 simulation for 400 yr, starting from the observed ocean (Levitus et al. 1998) and sea ice data. Our simulation is a present-day control simulation using greenhouse concentrations for CO\(_2\), CH\(_4\), N\(_2\)O, CFC-11, and CFC-12 prescribed at year 1990 levels of 355 ppmv, 1714 ppbv, 311 ppbv, 280 pptv, and 503 pptv, respectively. In this paper, this simulation is compared to a corresponding 700-yr T85 present-day control simulation performed at NCAR. The climate in the two numerical simulations is stable (section 3.2; Collins et al. 2006c) and we are confident that our conclusions would be the same with longer simulations of FVx1.

The difference in the ocean diffusivity complicates the direct comparison of the current results with those of Collins et al. (2006c). To establish that the improvements seen in the FVx1 can be attributed to the dynamical core, we analyzed a 200-yr FVx1 experiment with the Smagorinsky nonlinear diffusion in the ocean. This simulation had the same specification of diffusion in the ocean as the T85 control run. With the exception of the Arctic sea ice concentrations, the changes in the time-mean, spatial-mean quantities were small. This simulation and the sensitivity of the Arctic ice concentration to the formulation of the ocean viscosity will be discussed in detail in a forthcoming paper.

b. Computational aspects

The FVx1 simulation presented here was run on the LLNL Thunder machine, which consists of 1024 nodes containing 4 Itanium-2 processors and having 8 GB of memory, connected by a Quadrics switch. The simulation utilized 472 processors, with 336 assigned to the atmosphere, and achieved a throughput of roughly nine simulated years per computing day. On the same platform, a T85 simulation achieved roughly 5 yr day\(^{-1}\) on 260 processors, with 128 assigned to the atmosphere. However, the T85 simulation discussed in this paper was performed at NCAR.

The numerical implementation of the FV dycore is described in Mirin and Sawyer (2005). As with the Eulerian spectral dycore, it utilizes multiple domain decompositions that intercommunicate through transposes. However the FV dycore is able to invoke two-dimensional decompositions for the dynamics and is able to...
utilize roughly 4 times as many computational processors for a given resolution (Mirin and Sawyer 2005). This comparative scaling depends only weakly on resolution.

We are not aware of any comprehensive study comparing the computational efficiencies of the different dynamical cores. Scaling curves for both dynamical cores on leading parallel processors have been examined (Mirin and Worley 2007), but those for the FV dycore are at a finer resolution than that used in this study. Recent improvements in scalability are brought about by the ability to implement different sized decompositions for different parts of the solution. Memory utilization is more difficult to measure. Both dycores have been demonstrated to run at very high resolution (e.g., FV at 0.25°) on machines having 2GB of memory per processor.

3. Results

In this section, we provide a global-scale overview of the 400-yr present-day control simulation. First, we study the trend or drift in the simulation in section 3.1 and find that the climate drift in the FVx1 simulation is similar to the drift in the T85 simulation as evaluated by Collins et al. (2006c). In section 3.2, we use Taylor diagrams (Taylor 2001) to assess if the global-scale statistics from the FVx1 and T85 model simulations are similar to each other, and are consistent with the other state-of-the-art coupled models. The mean climate from the FVx1 simulation is compared to that from the T85 simulation in section 3c.

a. Climate drift

The approach of a climate model simulation to equilibrium is usually assessed by investigating the trends or drifts in the globally averaged surface air temperature, area of sea ice and energy budgets at the top of the model and surface. We assess the drift for the period 100–399 yr, and consider the first 100 yr as a spinup period. The superior conservative properties of the FV dycore could not be assessed here because global conservations of mass and energy were enforced in the formulation of CCSM. While the spectral discretization does not conserve dry air or tracer mass and total energy, its implementation in CCSM uses mass and energy adjustments to ensure conservation (Collins et al. 2006b). The FV dycore in CCSM conserves the dry air and tracer mass exactly. While the vertical discretization conserves total energy, the horizontal transport scheme in the FV dycore is associated with a loss of kinetic energy, and a global energy adjustment is added to the thermodynamic equation so that loss of kinetic energy is balanced by a gain in internal energy, thereby conserving total energy globally.

The global- and annual-mean net radiative flux at the top of the model (TOM) is \(-0.26 \pm 0.33 \text{ W m}^{-2}\) over the period 100–399 yr (Fig. 1a). This net energy flux is computed as the difference between the net solar flux and longwave flux at the TOM. On average, the FVx1, like the T85 simulation, loses energy. This loss rate is quite comparable to \(-0.21 \pm 0.28 \text{ W m}^{-2}\) reported for the T85 simulation (Collins et al. 2006c). The corresponding loss of energy flux at the surface is \(-0.26 \pm 0.25 \text{ W m}^{-2}\) over the same period. The surface net flux is computed by subtracting the sum of the net all-sky longwave flux, the latent heat flux and the sensible heat flux from the all-sky surface solar absorption. The net energy absorbed by the atmosphere is just the difference between the TOM and surface energy balances, and its mean with standard deviation is \(-0.007 \pm 0.15 \text{ W m}^{-2}\). This is negligibly small because the atmospheric model includes a uniform vertical adjustment to the temperature such that the change in atmospheric energy equals the globally integrated fluxes exchanged at the surface and the top of the model.

Figure 1b shows the time series of the global- and annual-mean surface air temperature (2 m). By this measure, FVx1 produces a remarkably stable climate after 15 yr. The mean surface air temperature is 287.4 ± 0.09 K over the period 100–399 yr. The linear trend over this period is \(-0.026 \text{ K century}^{-1}\). Much of this trend arises from the Southern Hemisphere between 30°S and 90°S, which cools at a rate of \(-0.06 \text{ K century}^{-1}\). The temperatures in the tropics between 30°S and 30°N and the Northern Hemisphere between 30° and 90°N have trends at much lower rates of \(-0.026\) and 9.56 \(\times 10^{-3}\) K century\(^{-1}\), respectively.

The area of Arctic sea ice is stable after 50 yr (Fig. 1c). It oscillates about its mean value (9.88 ± 0.24 \(\times 10^6\) km\(^2\) over the years 100–399 is in good agreement with observations (Fig. 1c). The Antarctic mean sea ice area (11.81 ± 0.44 \(\times 10^6\) km\(^2\) over the same period is about 20% higher than observations. After this initial 100 yr of adjustment, the trends in the Arctic and Antarctic sea ice are 0.01 and 0.18 \(\times 10^6\) km\(^2\) century\(^{-1}\), respectively. These changes correspond to changes in the sea ice concentration of 0.001% and 0.015% century\(^{-1}\). The trends are similar to the trends in the T85 simulation (Collins et al. 2006c). The larger rate of expansion of the sea ice area in the Antarctic is consistent with the decrease in the Southern Hemisphere temperatures.

The trend in the global volume-mean ocean temperature is \(-0.056 \text{ K century}^{-1}\) and the trend in the global volume-mean ocean salinity is \(-8.5 \times 10^{-4}\) psu century\(^{-1}\) (Figs. 1d and 1e). Compared to the global mean value of 34.72 psu, the relative trend is \(-2.4 \times 10^{-3}\)%.
century$^{-1}$. This reduction in salinity is caused by the gradual release of excess soil moisture to the oceans by river runoff in CCSM (Kiehl and Gent 2004). As in the T85 simulation (Collins et al. 2006c), the slow changes of ocean temperature occur well below the mixed layer, and the salinity adjustments are largely confined to the upper 1 km of the ocean (figure not shown). In general, the trends in our simulation are about the same or less than obtained in control simulations of other coupled climate models (Covey et al. 2006).

b. Comparison to other Intergovernmental Panel on Climate Change (IPCC) AR4 models

In this section, we compare the CCSM3 FVx1 and T85 simulations to present-day control simulations available from the CMIP3 (Coupled Model Intercom-
parison Project 3) database (http://www-pcmdi.llnl.gov/ipcc/about_ipcc.php). The main goal is to demonstrate that the magnitude of errors in the FVx1 and T85 model simulations are similar to each other and the errors in FVx1 simulation are consistent with the other state-of-the-art coupled models. The CMIP3 simulations are from the following models: PCM, MIUB-ECHO-G, MRI-CGCM2.3.2a, and UKMO-HadCM3, each with documentation available at the CMIP3 Web site. For each model, 20-yr climatologies were computed for selected variables, and global-scale statistics were computed comparing the simulations with available reference datasets.

We display these global-scale statistics on Taylor diagrams (Taylor 2001), which assess the performance of climate models in a compact fashion by portraying the standard deviation, correlation and the centered RMS on a single diagram. Multiple variables from multiple models can be displayed on the single diagram. In Fig. 2, colors identify variables and small circles indicate the models other than the T85 and FVx1 simulations in the CMIP3 database. The radial distance on the Taylor diagram represents the standard deviation (normalized by the standard deviation of the corresponding reference dataset). The cosine of the angle is the correlation between the simulations and the reference datasets, and the distance between a simulation and the reference point located along the x axis at unit distance is the normalized rms. See Taylor (2001) for additional details.

Space–time statistics over the annual cycle and global domain are shown in Fig. 2a. Correlations of the simulated 500-hPa geopotential height and 2-m temperature with observation are greater than 0.98 for all simulations. Correlations for zonal winds (850 hPa) and ocean surface stress are typically 0.95, but for the remaining fields shown the correlations are lower and the scatter between simulations is considerably larger. For each variable, Fig. 2a shows that the range of errors among all models is much greater than the difference in error between FVx1 and T85. For both the T85 and FVx1 runs the variance is too large in the zonal wind stress, 850-hPa winds and sea level pressure, and 200-hPa air-temperature, but similar problems can be identified in all models. The FVx1 RMS error is reduced in zonal wind stress and 200-hPa air temperature, relative to T85.

Annual cycle space–time statistics for the tropics (20°S–20°N) are shown in Fig. 2b. The annual cycle is generally weaker in the tropics, and the correlation and RMS are therefore more heavily influenced by the spatial characteristics of the fields. The surface air temperature and geopotential are quite smooth and are not strongly constrained by the insolation pattern, so their correlations are consequently lower than in Fig. 2a. To a lesser degree, this reduced correlation (in the tropics) is apparent in most other fields. One notable exception is the temperature at 200 hPa, which has an improved correlation in low latitudes owing to a well-known and common model cold bias in the high latitudes of the lower stratosphere.

Figure 2c is analogous to Figs. 2a,b, with the domain limited to the Northern Hemisphere extratropics (20°N–90°N) and with the zonal mean for each calendar month removed before the statistics are computed. This highlights characteristics associated with stationary wave patterns induced by land–sea contrasts or orography (e.g., the Rockies). While the correlations are generally lower than in Fig. 2a, for most fields the scatter is smaller among models, presumably because of the additional geographical constraints. The overall impression conveyed by Fig. 2 is that the magnitude of errors in the FVx1 and T85 model simulations are similar to each other and consistent with other state-of-the-art models.

c. Mean climate of FVx1

The mean climate produced by FVx1 is represented by the average of years 300–399. In this section, we compare it to the T85 mean climate represented by the average of years 600–699. The drift in the mean quantities is sufficiently small that they do not significantly impact the comparison.

1) Global-mean climate

Global and annual means along with root-mean-square errors of selected climate variables (Table 2) show that the global-mean FVx1 climate is nearly identical to the T85 climate. This is because of the retuning of some cloud parameters (Table 1). The global- and annual-mean shortwave and longwave cloud forcings are well tuned in both T85 and FVx1 to match the ERBE estimates (Barkstrom and Smith 1986) to within 0.1 W m⁻². The model differences are less than 1 W m⁻². The root-mean-square errors in circulation-related quantities such as sea level pressure, 200-mb zonal wind, and 500-mb geopotential are slightly reduced in FVx1, but improvement is not uniform across all fields considered.

2) Surface wind stress

The impact of changing the atmospheric dynamical cores on the simulated climate should manifest in the
simulated surface wind stress field prominently if we assume that the changes to the simulated climate are small and linear. These changes in wind stress in FVx1 would impact the sea surface temperature, ocean heat transport, sea ice distribution, etc., in the coupled simulation. By many measures, the simulation of surface wind stress in FVx1 shows significant improvement over T85 (Fig. 3) when compared to European Remote

**Fig. 2.** Taylor diagrams (Taylor 2001) comparing the CCSM3 FVx1 and T85 simulations to a few CMIP3 models. Shown are (a) global annual cycle space–time statistics, (b) annual cycle space–time statistics for the tropics, and (c) annual cycle space–time statistics of zonal-mean anomaly for the Northern Hemisphere extratropics (20°–90°N). The following reference datasets are used: CMAP (Xie and Arkin 1997) for precipitation, ISCCP (Rosen and Schiffer 1999) for total cloud cover, ERBE (Barkstrom and Smith 1986) for outgoing longwave radiation and reflected shortwave (both at the top of the atmosphere), ERS (Bentamy et al. 1999; Queffeulou et al. 1999) for zonal wind stress (ocean only), an updated version of the Jones 2-m temperature dataset (Jones 1988) and ERA-40 (Uppala et al. 2004) for winds, temperature, geopotential, and sea level pressure.
Sensing (ERS; Queffeulou et al. 1999) satellite observations. The mean RMS error decreases from 0.037 to 0.032 N m⁻² in FVx1. The overestimation of the wind stress in the North Pacific and the North Atlantic in T85 are reduced in FVx1. The stronger trades in the subtropical Pacific and the weaker westerly bias in the equatorial central Pacific are also reduced in FVx1. However, the North Atlantic still shows some of the largest biases in the simulation. There is also some reduction in the wind stress bias in the Antarctic Circumpolar Current region. The Southern Hemisphere stormtrack region is shifted to the south in FVx1, in better agreement with observations; similar improvement was noticed when the spectral model resolution was increased from T42 to T85 (Hack et al. 2006). The correlation between the two error patterns shown in Fig. 3 for the magnitude of wind stress is 0.89, indicating strong similarity of the two model configurations. The differences in the magnitude of the wind stress between the FVx1 and T85 simulations are statistically significant at the 1% level over 76% of the global ocean area (hatched area in the bottom panel of Fig. 3).

3) SEA SURFACE TEMPERATURE

A common model diagnostic to assess the performance of a coupled model is the sea surface temperature (SST) because SST represents, to a large extent, the fidelity of a coupled simulation due to its role in the air–sea exchange processes. The error patterns produced by subtracting the modeled SST from observations (HadISST dataset; Rayner et al. 2003) are shown in Fig. 4. The warm biases in the narrow coastal regions near Peru, Chile, and Baja, California are reduced. The cold SST biases in the North Atlantic are also significantly reduced in FVx1 in association with improved wind stress (Fig. 3). However, off the coast of Brazil, the cold biases are increased. Both the T85 and FVx1 simulations produce equatorial surface water in the eastern Pacific that is colder than observed, and extends too far west into the warm pool (Fig. 4). The North Pacific is warmer in FVx1 so that the cold bias in the western part of the basin is replaced by a smaller warm bias, and the warm bias in the eastern part of the basin is enhanced. Statistical analysis using the Student’s t test shows that these SST differences between the FVx1 and T85 simulations are statistically significant at the 1% level over 76% of the global ocean area (hatched area in the bottom panel of Fig. 4). The differences are noticeably not significant at the 1% level in the equatorial eastern Pacific where the variability is larger in association with ENSO (Fig. 5). The RMS error over the global domain is reduced from 1.54 K in T85 to 1.42 K in FVx1. The correlation between the two SST error patterns shown in Fig. 4 is 0.83, indicating strong pattern similarity in the SST simulated by the two versions of the model.

4) PRECIPITATION

A major model bias, the double ITCZ (Collins et al. 2006c), persists in the FVx1 simulation (figure not shown). Pattern correlations between Global Precipitation Climatology Project (GPCP) observations (Adler et al. 2003) and the T85 and FVx1 simulations are 0.81 and 0.82, respectively. The RMS error in FVx1 decreases to 1.30 from 1.37 mm day⁻¹ in T85. Both T85

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Table 2. Global and annual means and rmse of selected climate variables.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Observation</th>
<th>Mean T85</th>
<th>Mean FVx1</th>
<th>Rmse T85</th>
<th>Rmse FVx1</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface air temperature (K)</td>
<td>287.6⁵^a</td>
<td>287.3</td>
<td>287.4</td>
<td>3.2</td>
<td>3.2</td>
</tr>
<tr>
<td>Precipitation (mm day⁻¹)</td>
<td>2.6⁶ᵇ</td>
<td>2.8</td>
<td>2.8</td>
<td>1.4</td>
<td>1.3</td>
</tr>
<tr>
<td>Precipitable water (mm)</td>
<td>24.6⁷ᶜ</td>
<td>23.9</td>
<td>23.9</td>
<td>3.6</td>
<td>3.3</td>
</tr>
<tr>
<td>Total cloudiness</td>
<td>66.7⁸ᵈ</td>
<td>55.8</td>
<td>55.0</td>
<td>15.2</td>
<td>16.4</td>
</tr>
<tr>
<td>All-sky outgoing longwave radiation (W m⁻²)</td>
<td>233.9⁹ᵉ</td>
<td>235.5</td>
<td>235.8</td>
<td>10.9</td>
<td>10.7</td>
</tr>
<tr>
<td>Net all-sky shortwave flux at the top of the atmosphere (W m⁻²)</td>
<td>234.0⁹ᵉ</td>
<td>237.2</td>
<td>237.5</td>
<td>15.3</td>
<td>18.7</td>
</tr>
<tr>
<td>Longwave cloud forcing (W m⁻²)</td>
<td>30.4⁹ᵃ</td>
<td>30.3</td>
<td>30.5</td>
<td>9.6</td>
<td>10.1</td>
</tr>
<tr>
<td>Shortwave cloud forcing (W m⁻²)</td>
<td>–54.2⁹ᵃ</td>
<td>–54.1</td>
<td>–54.1</td>
<td>17.1</td>
<td>19.4</td>
</tr>
<tr>
<td>Net all-sky surface shortwave absorption (W m⁻²)</td>
<td>165.9⁹ᵇ</td>
<td>159.5</td>
<td>160.1</td>
<td>17.7</td>
<td>17.7</td>
</tr>
<tr>
<td>Net all-sky surface longwave flux (W m⁻²)</td>
<td>49.4⁹ᵇ</td>
<td>58.4</td>
<td>59.3</td>
<td>15.5</td>
<td>16.1</td>
</tr>
<tr>
<td>Sea level pressure (mb)</td>
<td>1011.6⁹ᵇ</td>
<td>1010.6</td>
<td>1011.1</td>
<td>5.0</td>
<td>4.7</td>
</tr>
<tr>
<td>200-mb zonal wind (m s⁻¹)</td>
<td>15.3⁹ᵇ</td>
<td>18.4</td>
<td>17.4</td>
<td>4.5</td>
<td>4.0</td>
</tr>
<tr>
<td>500-mb geopotential height (m)</td>
<td>5659⁹ᵇ</td>
<td>5674</td>
<td>5671</td>
<td>36</td>
<td>30</td>
</tr>
</tbody>
</table>

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³Legates (Legates and Willmott 1990); ⁴GCP (Adler et al. 2003); ⁵NVAP (Randel et al. 1996); ⁶ISCCP (Rossow and Schiffer 1999); ⁷ERBE (Barkstrom and Smith 1986; Harrison et al. 1990; Kiehl and Trenberth 1997); ⁸ISCCP FD (Zhang et al. 2004); ⁹ECMWF (Uppala et al. 2004); ¹⁰NCEP (Kalnay et al. 1996).
FIG. 3. Differences in annual-mean surface wind stress (N m$^{-2}$) between (top) CCSM3 T85 and (middle) FVx1 simulations, and the ERS dataset (Bentamy et al. 1999). The vectors show the direction and magnitude of wind stress differences. (bottom) Differences between the FVx1 and T85 simulations are shown; the differences in magnitude are significant at the 1% level in the hatched area.
FIG. 4. Same as Fig. 3, but for the annual-mean surface temperature (°C) and the HadISST dataset (Rayner et al. 2003).
and FVx1 produce excessive rainfall in the tropical Pacific and Indian oceans. The wet biases in the central Pacific and in the South Pacific convergence zone are reduced somewhat in FVx1. The regional land precipitation deficits over the southeast United States, Amazonia, and Southeast Asia, and the excess over central Africa, northern India, Australia, and the western United States (Collins et al. 2006c) persist in the FVx1 simulation.

5) CLOUD FORCING

The meridional distribution of zonal-mean shortwave and longwave cloud radiative forcings (SWCF and LWCF) are also similar in the T85 and FVx1 simulations (figure not shown). The models are close to each other because cloud parameters (Table 1) in each model were tuned in similar manner to obtain agreement with ERBE observations. The largest zonal-mean difference in SWCF between observations (ERBE; Barkstrom and Smith 1986) and both models occurs in the Southern Hemisphere storm-track regions. The model simulations underestimate LWCF in midlatitudes.

6) TROPICAL VARIABILITY

We computed the total variance of the annual-mean sea surface temperatures during the last 100 yr of the model simulations and 40 yr (1960–1999) of the HadISST dataset (Fig. 5). The location of the peak total variance is shifted to the west in the tropical Pacific in the model simulations relative to the observations (Deser et al. 2006). There is more variance near the coast of Peru in FVx1 than in T85, suggestive of improvement in FVx1. The total variance in the Niño-3 region of the eastern Pacific (5°S–5°N, 90°–150°W) for the HadISST dataset is 0.52 K², and 0.33 and 0.57 K² for the T85 and FVx1 simulations, respectively. Clearly, the amplitude and location of the total variance of the sea surface temperature in the eastern tropical Pacific has improved in the FVx1 simulation. However, the variance is extensive in the zonal direction leading to stronger variability in the western tropical Pacific in the FVx1 simulation than in the observations.

The power spectrum of the sea surface temperature (figure not shown) in the Niño-3 region shows that the amplitude and period of the interannual variability are similar in the T85 and FVx1 simulations. The observations show a broader peak in the spectrum with the period of 3–7 yr; the model produces a sharper peak with period around 2 yr in both T85 and FVx1 simulations. There is an overestimate of the amplitude and underestimate of the peak period of the major mode of the tropical variability. This is a persistent bias in the CCSM (Deser et al. 2006). It is speculated that the narrow meridional scale of variance in the simulations (Fig. 5) may contribute to the excessively high frequency (Deser et al. 2006).

To investigate if the large differences in the variance between the two simulations lead to different teleconnection patterns, we compute the correlation between the Niño-3 index (mean SST anomaly in the Niño-3 region) and SST anomalies (top panels of Fig. 6). Annual-mean sea surface temperatures during the last 100 yr of the model simulations and 40 yr (1960–99) of the HadISST dataset (Rayner et al. 2003) are used for this calculation. The correlation pattern is similar in both the simulations (Fig. 6); global SST anomalies have weak teleconnection to the tropical Pacific region when compared to the observations. For instance, there is a strong positive correlation in the Indian Ocean, South Pacific, and along the entire west coasts of North and South America in the observations, which the models are not able to reproduce the strength of. The strong negative correlation in the western north and south tropical Pacific is also not simulated by the T85 and FVx1 simulations. The positive correlation in the equatorial Pacific extends too far to the west in the simulations, which is consistent with the zonally extended variance (Fig. 5) in this region.

The atmospheric midlatitude teleconnection is shown...
in the bottom panels of Fig. 6, which shows the slope of the linear regression relationship between the Niño-3 index and the 300-mb geopotential. Annual-mean sea surface temperatures during the last 100 yr of the model simulations and 40 yr (1960–99) of the HadISST SST dataset and ERA-40 geopotential (Uppala et al. 2004) are used to compute the regression. While the models reproduce the observed relationship in the North Pacific and over North America reasonably well, they differ in other regions. In the Arctic and over Asia, the T85 simulation is in better agreement with observations while the FVx1 simulation shows better agreement in the South Pacific.

7) OCEAN CIRCULATION

The mean vertically integrated ocean circulation is represented by the barotropic streamfunction (Fig. 7). The annual-mean climatologies of the ocean from the last 10 yr of the simulations are used to compare the barotropic streamfunctions. In the Southern Ocean, the barotropic streamfunction is reduced by 20–30 Sv (1 Sv = 10^6 m^3 s^-1) in the FVx1 simulation. This is reflected in the mean volume transport through the Drake Passage in the FVx1 simulation. The Drake Passage volume transport of 169 Sv in the FVx1 simulation is substantially closer to the observed value of 134 ± 10 Sv (Whitworth and Peterson 1985) than that simulated in T85 (192 Sv).

The maximum transports simulated by the T85 configuration are in good agreement with observations in the North Atlantic subtropical and subpolar gyres, and the North Pacific subtropical gyre (Large and Danabasoglu 2006). The barotropic transport in T85 is underestimated off the Cape of Good Hope (about 85 versus 100 Sv), and overestimated in the South Atlantic subtropical gyre (about 60 versus 30 Sv). These biases persist in the FVx1 simulation. The mean volume transports in other key passages are comparable in the T85 and FVx1 simulations: Florida and Cuba (27.2 versus 28.1 Sv), Indonesian throughflow (13.7 versus 13.8 Sv), and the Bering Strait (0.9 versus 1.5 Sv).

We find that the difference in global and North Atlantic total heat transport between the two model simulations is of the order of 10% or less (figure not shown). In the Southern Hemisphere, the total poleward heat transport is greater in the FVx1 simulation, while in the Northern Hemisphere, it is weaker near the equator but stronger between about 20° and 50°N. In the western boundary current separation regions of the Northern Hemisphere, we find that the FVx1 total heat transports are higher. This is likely due to the differing representation of the Gulf Stream–North Atlantic Current

FIG. 6. (top) Correlation coefficient between the Niño-3 index and global SST anomalies and (bottom) linear regression between Niño-3 index and the 300-mb geopotential field. The correlation and regression for observations, and CCSM3 T85 and FVx1 simulations are shown on the lhs, center, and rhs, respectively.
FIG. 7. Annual-mean vertically integrated mass transport (Sv; barotropic streamfunction) from CCSM3 (top) T85 and (middle) FVx1 simulations. (bottom) The difference in transport between the FVx1 and T85 simulations.
in the two models (see the barotropic streamfunction in Fig. 7).

8) Sea ice distribution

Figure 8 compares the mean DJF (mean over December, January, and February) and JJA (mean over June, July, and August) ice concentration in the Arctic from the model simulations with the Special Sensor Microwave Imager (SSM/I) satellite data (Cavalieri et al. 1997; Comiso 1999). General agreement is seen in the position of the ice edges in T85 and FVx1. However, the models simulate excessive winter ice along the eastern Greenland coast, in the Hudson Strait, the Sea of Okhotsk, and the Sea of Japan (Holland et al. 2006). Both configurations show no ice coverage in the Barents Sea as observed, with the FVx1 simulating the ice-free regions better. The winter ice distribution in the FVx1 configuration has improved in the Davis Strait, Labrador Sea and along the western Greenland coast and Newfoundland coast; the coastal sea ice has decreased. This improvement in the FVx1 simulation is due to the viscosity change in the ocean model (B. Large 2006, personal communication). More discussion on the sensitivity of the Arctic winter ice distribution to the ocean viscosity will be provided in a forthcoming paper. In the summer, the mean Arctic ice distribution in the T85 and FVx1 simulations agrees well with the SSM/I dataset. The models simulate less ice area in the Hudson Bay. Large differences from observations can be seen near Canada.

The seasonal cycle of sea ice (figure not shown) shows that both T85 and FVx1 simulate 10%–20% more winter ice than observed. With the summer ice area closer to observations, this yields a higher than observed seasonal cycle. The seasonal amplitude is consistently reduced in the FVx1 simulation and is closer to the observed seasonal cycle, and this reduction is due to the viscosity change in the ocean model.

The mean sea ice thickness in the central Arctic is about 2–2.5 m in both of the simulations (Fig. 9), in reasonable agreement with submarine measurements of sea ice thickness (Rothrock et al. 1999). The regional distribution has improved in the FVx1 simulation relative to the T85 simulation, notably in the Canadian Archipelago and East Siberian Sea. The T85 simulation underestimates the sea ice in the Canadian Archipelago by 1 m and overestimates by 2 m in the East Siberian Sea (Collins et al. 2006c). This bias has been reduced by 0.6 m or more in the FVx1 simulation. The sea ice thickness is about 4–5 m in the Canadian Archipelago, in good agreement with observations (Bourke and Garrett 1987). The tendency for thinner ice along the Siberian coast and thicker ice along the Canadian coast, relative to the T85 simulation, was also seen in the FVx1 simulation with the Smagorinsky viscosity (figure not shown).
When the resolution of CCSM3 is increased from T42 to T85, the simulation of the sea ice thickness distribution improves substantially (DeWeaver and Bitz 2006). At T42 resolution, Arctic ice is thick off the Siberian coast and too thin in the Canadian Archipelago. Both of these biases are reduced at T85 resolution. The improvement along the Canadian coast was primarily attributable to the reduction in the strength of the erroneous North Polar summer anticyclone. DeWeaver and Bitz (2006) noted that the simulated Canadian ice thickness difference is related to the North Polar summer cyclone, and the modeled ice build-up difference on the Siberian coast is associated with the anticyclonic circulation in the Beaufort Sea in the fall and winter.

In Fig. 10, we show the DJF- and JJA-mean surface winds over the Arctic. In the winter, anticyclonic circulation in the model simulations in the Beaufort Sea is much weaker than the National Centers for Environmental Prediction (NCEP) reanalysis (Kalnay et al. 1996) winds. The reanalysis winds have a magnitude of 2–3 m s$^{-1}$ while the magnitude in the models is less than 1 m s$^{-1}$. The reanalysis shows winds crossing the Arctic Ocean from Siberia to Canada. In the models, however, winds originating from Siberia near 90°E return to the eastern Siberian coast. Although the circulation in the Arctic are very similar in both T85 and FVx1 simulations, the difference wind field between the FVx1 and T85 simulations showed (figure not shown) winds crossing the Arctic Ocean from the Siberian coast to the Canadian coast, indicating more ice build up in the Canadian coast in the FVx1 simulation.

There are pronounced differences between the T85 and FVx1 simulations in the summer. The NCEP reanalysis and the FVx1 simulation show a cyclonic circulation around the pole (Fig. 10). The T85 simulation has no closed circulation at the pole, and the winds cross the Arctic Ocean from the Canadian Archipelago to the Siberian coast. Though cyclonic winds are weaker in the FVx1 simulation than observed, they circulate in the right direction with the flow from East Siberian Sea toward the Canadian Archipelago. Thus, the more realistic simulation of the North Polar summer cyclone in FVx1 results in significant improvement in the ice thickness distribution in the Canadian Archipelago. This cyclonic circulation in the summer was also present in the FVx1 simulation with the Smagorinsky viscosity in the ocean, indicating the robustness of the

**Fig. 9.** Mean sea ice thickness (m) in the Arctic in the NH (top) winter and (bottom) summer in the CCSM3 (left) T85 and (middle) FVx1 simulations, and (right) the difference between the FV and spectral T85 simulations.
improvements in the Arctic atmospheric circulation in the FVx1 simulations.

4. Discussion

We present an overview of the fidelity of the CCSM3 simulation using the Finite Volume dycore for the atmospheric dynamics. We compare our simulation to the standard configuration of the model that uses the spectral dycore, and to observations. Substantial retuning of the FV atmospheric model CAM3 is carried out before coupling CAM3 to CCSM3. The tuning brings the shortwave and longwave fluxes in close agreement to ERBE measurements. It also brings the top of the atmosphere budget to within a few tenths of 1 W m$^{-2}$.

In implementing the FV dycore, the specification of the terrain filter is changed. We adopt an “optimal” two-grid filter in the physical space. It filters only the grid-scale waves. This change in the terrain filter indicates that the formulation of the topography should be
part of the dynamical formulation. To obtain a realistic distribution of sea ice concentration in the Arctic, the horizontal viscosity formulation in the momentum equations of the ocean model is modified in the FVx1 simulation; a linear option replaces the nonlinear Smagorinsky viscosity option in the anisotropic formulation of horizontal diffusion for the momentum components in the ocean model.

Quantitative comparison to other coupled models and observations using Taylor diagrams demonstrate that the FVx1 version of the CCSM coupled model is a state of the art climate model whose simulation capabilities are in the class of those used for IPCC assessments. The simulated climate is very similar to that of the T85 version in terms of its biases, and more like the T85 model than the other IPCC models. Our results suggest that the FVx1 version of CCSM3 is a suitable coupled model for future climate change assessments.

The climatological means from the last 100 yr of the FVx1 simulation are compared to the T85 simulation. We find marginal improvements in the simulation of sea surface temperature, surface wind stress, and other atmospheric circulation fields, such as sea level pressure, 200-mb zonal wind, and 500-mb geopotential. There is significant improvement in the ocean transport in the Antarctic Circumpolar Current through the Drake Passage. The Arctic sea ice thickness distribution has also improved, with thicker ice along the Canadian coast and thinner ice in the East Siberian Sea. The improvements in the polar region are associated with an improved surface wind stress simulated by the atmospheric model in the FVx1 version.

To assess if the improvements in the FVx1 simulation are due to the FV formulation or the ocean viscosity formulation, we analyzed a 200-yr FVx1 simulation that used the nonlinear Smagorinsky viscosity formulation. Examination of this simulation confirmed the following improvements compared to T85: wind stress in the Antarctic Circumpolar Current (Fig. 3), ocean transport in the Antarctic Circumpolar Current (Fig. 7), summer cyclonic circulation in the Arctic (Fig. 10), thicker ice distribution in the Canadian Archipelago (Fig. 9), and the variance in the tropical Pacific (Fig. 5). This analysis confirmed that the improvements in the atmospheric circulation and resultant Arctic sea ice thickness distribution in the FVx1 simulation are independent of the ocean viscosity change.

Some troublesome model biases persist in the FVx1 simulation: the double ITCZ, the 2-yr periodicity of the ENSO cycle, the colder tropopause in the high latitudes, warmer winter time land temperature in the high latitudes, precipitation deficits in the southeast United States, Amazonia, and the Southeast Asia, the semian-urnal SST cycle in the eastern Pacific, and the underestimation of downwelling solar radiation in the Arctic.

In this paper, we focused on establishing the fidelity of the climate simulations using standard temporally and spatially averaged quantities. This is a necessary part of the model evaluation procedure. However, our preliminary investigations of the local processes do reveal that the Arctic sea ice is sensitive to changes in the local atmospheric dynamics. The preliminary investigation also showed positive impact of the FV dycore on local processes, especially, the pattern of continental precipitation and wind stress near steep topography. These improvements in the simulation of local climate will be discussed in detail in forthcoming papers. A detailed analysis of ocean circulation in the FVx1 simulation will be also the subject of a future paper.

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