Simulation of the Global Hydrological Cycle in the CCSM Community Atmosphere Model (CAM3): Mean Features

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Abstract

The seasonal and annual climatological behavior of selected components of the hydrological cycle are presented from coupled and uncoupled configurations of the atmospheric component of the Community Climate System Model (CCSM), the Community Atmosphere Model Version 3 (CAM3). The formulations of processes that play a role in the hydrological cycle are significantly more complex when compared with earlier versions of the atmospheric model. Major features of the simulated hydrological cycle will be compared against available observational data, and the strengths and weaknesses will be discussed in the context of the fully-coupled model simulation.

1. Introduction

The accurate treatment of Earth's hydrological cycle, the circulation of water in the climate system, is central to scientific explorations of climate dynamics and climate change. Hydrological processes are an integral part of the general circulation in that they play a major role in the energy budget of the planet, and do so in a very complex way on a wide range of space and time scales. Water can be found in all three phases in the climate system, and is a strongly radiatively-active atmospheric constituent in all forms. The liquid and frozen phases are most familiar in the form of clouds. Clouds play a dominant role in regulating the energy budget of the planet, and their behavior remains a major source of uncertainty in our ability to project the effects of climate change ((e.g., Stephens and Webster, 1981; Cess et al., 1990; Stephens and??, IPCC TAR). They cool Earth by reflecting solar radiation back to space, while at the same time warming the planet by absorbing thermal radiation emitted from the surface and

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lower regions of the atmosphere. The processes responsible for phase transitions of water also contribute to the diabatic forcing of Earth's dynamical circulations, and are key to the overall energy budget. This is particularly true for the thermally driven circulations in the tropics and subtropics (Chahine, 1992).

The hydrological cycle begins with the evaporation or sublimation of water from Earth's surface where it is transported by the ambient motion field. When the air is lifted, it cools and allows water in the vapor phase to condense in clouds. where it can exist in both the liquid or frozen phase. Microphysical processes determine whether the cloud condensate is re-evaporated, changes phase, or grows particles large enough in size to precipitate back to the surface. Once the precipitation reaches the surface it can be re-evaporated, can produce runoff that finds its way into lakes, rivers, and oceans, or can infiltrate the surface and be stored in the water table. All of these processes operate on an all-inclusive range of time and space scales, and are very difficult to quantify observationally. The most reliable observational guidance on the characteristics of the hydrological cycle is on relatively long time scales and relatively large space scales. As such, current observational data provide relatively weak contraints on the formulation of hydrological processes in global models. A major challenge to the design of global climate models is to realistically incorporate physical processes that operate on scales of motion distinctly separate from those of the larger-scale resolvable circulation, but strongly influence the behavior of the atmosphere on all time and space scales.

Version 3 of the Community Atmosphere Model (CAM3) is the latest in a succession of general circulation models that have been made widely available to the scientific community, originating with the NCAR Community Climate Model (CCM). This model is the atmospheric component of the Community Climate System Model (CCSM), a fully-coupled modeling framework that can be used for a broad class of scientific problems. The CCSM3 represents the latest generation of this modeling framework, and is discussed in more detail by (Collins et al. 2004a). CAM3 incorporates a significant number of changes to the dynamical formulation, the treatment of cloud and precipitation processes, radiation processes, and atmospheric aerosols, and is discussed more fully in (Collins et al. 2004c). The representation of cloud and precipitation processes has been significantly revised, including separate treatments of liquid and ice condensate, large-scale advection, detrainment, and sedimentation of cloud condensate; and separate treatments of frozen and liquid precipitation (Boville et al. 2004). The parameterization of radiative transfer has been updated to include a generalized treatment of cloud overlap (Collins et al. 2001) and new treatments of longwave and shortwave interactions with water vapor (Collins et al. 2002a, 2004b). Finally, a prescribed climatological distribution of sulfate, soil dust, carbonaceous species, and sea salt based upon a three-dimensional assimilation (Collins 2001; Rasch et al. 2001) is used to calculate the direct effects of tropospheric aerosols on the heating rates (Collins et al. 2002b). This latter change is noteworthy in the context of what follows given that the radiative effects of atmospheric aerosol has been shown to strongly influence the behavior of hydrological processes (Hansen, ??; Ramanathan and ??).

The CAM3 has been designed to provide simulations with comparable large-scale fidelity over a range of horizontal resolutions for several different dynamical approximations. This principally requires modifications to adjustable parameters in the parameterized physics package associated with clouds and precipitation. Consequently, the detailed hydrological process behavior has some dependence on horizontal resolution and the formulation of the dynamical core. These issues, along with the sensitivity of the simulated climate to model resolution is discussed in more detail in Hack (2004), Yeager et al. (2004), and DeWeaver and Bitz (2004). The standard configuration of the CAM3 is based on an Eulerian spectral dynamical core, where the vertical discretization makes use of 26 levels (L26) treated using second-order finite differences (Williamson 1988). The vertical domain is essentially the same as in earlier models, but employs 8 additional levels which were included to better refine the upper troposphere and lower stratosphere. The discussion that follows will focus on the standard CAM3 configuration that uses a 26-level 85-wave triangular spectral truncation (T85L26). This truncation translates to a 1.4°transform grid ($\sim$ 150 km near the equator) on which non-linear and parameterized physics terms are evaluated. This reflects a four-fold increase in the number of horizontal degrees of freedom when compared to earlier models (Hack et al. (1994), Kiehl et al. (1998), Kiehl and Gent, 2003). A 20-year 5-member ensemble, using observed sea surface temperatures and observed sea ice, is used to characterize the mean features of the simulated hydrological cycle in the uncoupled configuration. These simulation characteristics are then contrasted with simulation properties obtained from the fully-coupled CCSM3 (Collins et al. 2004a).

There are a large number of observational datasets related to Earth's hydrological cycle. The datasets we will use include the 40-year reanalysis dataset (ERA40) from the European Center for Medium Range Weather Forecasting (ECMWF). The archive consists of monthly mean data on 23 pressure levels in the vertical at 2.5 degree resolution. The data are regridded to T42 spectral resolution for our analysis. The Nimbus-7 Cloud Matrix total cloud data are derived from the Temperature Humidity Infrared Radiometer (THIR) and Total Ozone Mapping Spectrometer (TOMS) measurements for the period April 1979-Mar 1985. Retrieval algorithms are described in detail in Stowe et al. (1988, 1989). The Global Precipitation Climatology Project (GPCP) Version-2 Monthly Preciptation Analysis is a global precipitation dataset on a 2.5 degree grid extending from 1979-2003. It is a merged dataset consisting of satellite microwave and infrared data and surface rain guage data. The CPC Merged Analysis of Precipitation (CMAP) is a global monthly precipitation dataset on a 2.5 degree grid, covering the period 1979-1998. The National Aeronautics and Space Administration Water Vapor Project global water vapor dataset (NVAP) is a water vapor and liquid water path archive at 1 degree resolution that extends from 1988-1999. The blended analysis includes satellite retrievals of water vapor from the Television and Infrared Operational Satellite (TIROS) Operational Vertical Sounder (TOVS), the Special Sensor Microwave/Imager (SSMI), as well as radiosonde measurements. THIS NEEDS MUCH MORE WORK AND THE ADDITION OF ISCCP, SSMI RETRIEVALS OF WENTZ, AND THE NEW MODIS DATASET(S).

2. Mean-State Simulation Properties: Uncoupled Configuration

a. *Global Mean Properties:*

We begin our discussion of the simulated hydrological cycle by examining the global annual budget of water in the CAM3 using a a 22-year 5-member ensemble that was driven by observed sea surface temperatures and observed sea ice for the period 1979 through 2000. The land surface in these simulations is fully interactive. The simulated global annual precipitation rate of 2.86 mm/day is approximately 7% larger than the CMAP satellite estimate (ref). We also note that the magnitude of the hydrological cycle, as defined by the global annual precipitation rate, is a bit more than 8% weaker than in the previously documented atmospheric model in this series (Hack et al. 1998). The reduction in the hydrological cycle is largely attributable to significant changes in the surface energy budget associated with the introduction of specified atmospheric aerosols, and will be discussed elsewhere in this special issue. Fig. 1 shows the breakdown of the precipitation and evaporation exchanges of water in absolute terms between the atmosphere and land, ocean, and sea ice surfaces, along with runoff from the land and ice to the oceans. The relative distribution of surface exchange is remarkably similar to renormalized observational estimates (e.g., Peixoto and Kettani, 1973). The observational estimates of precipitation, evaporation, and runoff, are normalized by the CAM3 simulated value of global precipitation, and show that all terms in the surface exchanges generally agree to within a few percent of the relative partitioning contained in the observational estimate.

Figure 1 also shows the time averaged storage of water in the atmosphere in all three phases as simulated by the CAM3, along with estimates of water vapor (i.e., precipitable water) and cloud liquid water retrieved from MODIS. We have used MODIS estimates since this retrieval provides the most comprehensive global coverage of cloud liquid water, but not necessarily the most definitive retrievals of precipitable and cloud liquid water. The total reservoir of water in its frozen, liquid, and vapor state is broken down by its distribution over land, ocean and ice. From this perspective, the CAM3 does a reasonable job of simulating the distribution of water with regard to surface type. Since these numbers are strongly weighted by the fractional areas of ocean, land and sea ice, it's no surprise that the largest fraction for each phase resides over the oceans, The table in Fig. 1 suggests that the CAM3 generally overestimates precipitable water in the atmosphere, overestimates cloud liquid water over the oceans, and underestimates cloud liquid water over land and sea ice. Another way to look at the distribution of water, and its exchange with the surface, is to quantify the average properties for each of the three underlying surfaces, as shown in Tables 1 and 2.

Table 1 shows precipitation and evaporation data for the CAM3, evaporation data for the ERA40 reanalysis, and precipitation data from the CMAP climatology. We have not included ERA40 precipitation estimates because of known spin-up problems in the analysis of precipitation (Sakari Uppala, personal communication). The evaporation side of the water budget, however, does not suffer as much from hydrological spin-up problems, and compares well with the da Silva et al. (1994) climatology (Anton Beljaars, personal communication). CMAP provides only one side of the water budget,

Table 1: Annual Average Precipitation and Evaporation Rates by Surface Type (mm/day)

	Ocean		Land		Sea Ice		Global	
	P	E	P	E	P	E	P	E
CAM3	3.22	3.61	2.26	1.48	1.37	0.43	2.87	2.87
CMAP	3.11	—-	1.93	—-	1.10	—-	2.68	—-
ERA40	—-	3.58	—-	1.49	—-	0.41	—-	2.84

Table 2: Annual Average Storage of Vapor, Cloud Water, and Cloud Ice by Surface Type (mm)

	Ocean			Land			Sea Ice		
	Vapor	Liq.	Ice	Vapor	Liq.	Ice	Vapor	Liq.	Ice
CAM3	27.92	.1270	.0191	18.29	.1055	.0199	6.40	.1530	.0334
MODIS	26.59	.0982	——	15.64	.1311	——	4.45	.2017	——
NVAP	25.65	.1127	——	20.15	——	——	5.96	——	——
ERA40	28.27	.1170	.0358	19.77	.0718	.0339	5.84	.0274	.0470

but provides a useful quasi-independent measure of the magnitude of the hydrological cycle, and the distribution of precipitation across surface types. To some extent, the disagreements in these estimates help illustrate continuing uncertainty in quantifying the magnitude of Earth's hydrological cycle, even on long time scales. We see that the magnitude of the CAM3 hydrological cycle is approximately 7% larger than the CMAP estimate using global annual precipitation as our measure, but bears a much closer relationship to the reanalysis if we use global annual evaporation as the measure. With regard to precipitation, in a relative sense precipitation rates are greater over land and sea ice in the CAM3 when compared to oceanic rates using CMAP estimates as the reference observation.

The components of water storage in the atmosphere are generally difficult quantities to observe on global scales. Much of this data comes from satellite retrievals, often blended with in-situ and/or analysis data, and is most often limited to vertical integrals of precipitable water and cloud water. Table 2 shows the simulated precipitable water, cloud liquid water, and cloud ice water by surface type for the CAM3, along with comparable available observational estimates from NVAP, MODIS, and ERA40. One can see that there are considerable differences between the various observational estimates, and are of comparable magnitude to differences with the CAM3. Generally speaking, the CAM3 appears to agree reasonably well with NVAP and ERA40 estimates of precipitable water, which are slightly higher than the MODIS retrievals. Similarly, simulated cloud liquid water over the oceans is greater than MODIS, but is significantly lower than MODIS over land and sea ice covered surfaces. WE NEED TO SAY SOMETHING MORE ABOUT THE ANALYSES NOW THAT WE HAVE THE DATA.

b. *Zonal Mean Properties:*

We next look at the zonally time-averaged characteristics of water exchange between the atmosphere and the Earth's surface. The zonally averaged seasonal and annual distribution of precipitation for the CAM3 is shown in Fig. 2 in comparison with precipitation estimates from CMAP. In most respects the CAM3 exhibits similar biases to those seen in the CCM3 simulation. The amplitude of the tropical precipitation is generally well captured, although there is a slightly more exaggerated double-ITCZ than in CCM3, most notably during DJF. The simulated location of the DJF ITCZ maximum, more than 10°north of the analyzed CMAP maximum illustrates a major problem with the representation of tropical precipitation, which is the persistence of ITCZ-like precipitation in the Northern Hemisphere year round. This contrasts with all observational estimates which show a clear seasonal migration of ITCZ precipitation across the equator. Subtropical minima are generally displaced too far poleward seasonally as well as annually, as are the secondary precipitation maxima in the extratropical storm tracks. The poleward shift of the CAM3 Southern Hemisphere storm track results in a modest positive precipitation bias when compared to satellite retrievals.

The zonally averaged seasonal and annual evaporation rate is shown in Fig. 3 for the CAM3 and CCM3. CCM3 is used in this comparison because of problems with identifying a comparable global observational dataset, and to illustrate the reduction in the magnitude of the hydrological cycle between CCM3 and CAM3. These figures clearly show a significant and systematic reduction in the CAM3 evaporation rates when compared to the CCM3. This reduction in the magnitude of the hydrological cycle is primarily associated with the introduction of a climatologically-specified distribution of atmospheric aerosol that produce a significant reduction of net solar radiation at the surface. As was the case for the CCM3 atmospheric model, the most vigorous transfer of water to the atmosphere occurs in the subtropics with evaporation rates reaching maximum values near 15°N and 15°S. Consistent with observational analyses, the Southern Hemisphere oceans are the principal source of water powering the atmospheric hydrologic cycle in the CAM3. The suppression of evaporation in the vicinity of ITCZ convection is a realistic feature of the CAM3 simulation, which is in good agreement with corresponding oceanic estimates (e.g., see Oberhuber, 1988; Doney and Large, 1997; Kiehl, 1997; Large and Yeager, 2004). We also note a substantial reduction in the surface flux of water to the atmosphere poleward of 80° N which introduces a large change in the water budget over the Arctic when compared to the CCM3.

The zonally averaged seasonal and annual net surface exchange of water, $E - P$, is shown for CAM3 and CCM3 in Fig. 4, and is quantified in units of energy (where 1 mm day^{-1} $\sim$ 29.055 Wm^{-2}). The CAM3 simulates a strong seasonal meridional oscillation in the source regions, but a relatively weak seasonal movement of the equatorial water sink. The weak meridional excursion of the net water sink in the deep tropics is largely attributable to the unrealistically weak seasonal migration of ITCZ precipitation. The regions 10°N – 40°N and 10°S – 40°S are well defined source regions of total water, where the deep tropics and high-latitude extratropics represent the principal sinks. In most respects, the CAM3 net water budget bears remarkable similarity to the CCM3, especially considering the relatively large local changes in the individual components of the water exchange. The largest changes in $E - P$ occur along the equator, and poleward of 60 $^{\circ}$N. The equatorial differences are largely a consequence of reduced precipitation rates that is manifested in the form of a stronger double ITCZ, particularly apparent in the Indian Ocean. Over the Arctic the evaporation deficit is nearly twice as large as in the CCM3, largely due to large systematic reductions in the surface evaporation rate.

The zonally averaged precipitable water, or vertical integral of the specific humidity, is shown in Fig. 5, along with the NVAP climatology. The CAM3 is systematically moister than the CCM3, and in better agreement with zonally-averaged observational estimates such as NVAP. The largest bias is present year round near 30°N, exceeding 4 kg m^{-2} (or 4 mm) over most of the region between 10°and 40° N. As we will show, the agreement in the zonal mean distribution of precipitable water is the consequence of a fortuitous cancellation of errors in the longitudinal distribution.

As mentioned earlier, clouds provide important forcings on the climate system through their modulation of the radiative heating field. The climatological distribution of cloud and cloud condensate is therefore worthy of some discussion. The radiative effects of the simulated cloud field are discussed in Collins (2004), where we confine this discussion to the extent of cloud cover and the simulated path length.

The annually and zonally averaged meridional distributions of total cloud amount are shown in Fig. 6 for CAM3, ISCCP, and Nimbus 7. The CAM3 cloud field is markedly different from the CCM3. Total cloud cover in the tropics and poleward of the extratropical storm tracks is significantly reduced in the CAM3. This is dominated by a sharp reduction in high cloud over most of the globe, and reductions in mid and low-level cloud at high latitudes. The reductions in tropical high cloud are compensated by increases in middle-level cloud, where low level cloud has systematically increased equatorward of 60°N and 60°S. The reduction in high cloud is more consistent with ISCCP estimates, while the increase in low level cloud amount is more consistent with Warren (19XX). Despite some improvement in the distribution of simulated cloud amount, important biases in the meridional distribution remain in CAM3. One of the more obvious longstanding deficiencies is the location of the minima in subtropical cloud cover, near 20°latitude in the observational record, but closer to 30°in the CAM3 simulation. This difference has important consequences for the radiative budget of the subtropics where, for example, the tropical shortave cloud forcing is too broad and therefore too strong for much of the subtropics. The radiative issues related to the simulation of cloud and cloud optical properties is not the focus of this manuscript, but will be discussed elsewhere using the ISCCP simulator developed in Klein and Jakob (1999) and Webb et al.(2001).

As noted in Table 2, condensed water in the atmosphere is several orders of magnitude smaller than storage in the vapor state, and yet is of comparable climate importance in terms of modulating the global radiation balance (e.g., Wielicki et al., 1995; Kiehl, 1994b). Zonally and annually averaged distributions of liquid water path are shown in Fig. 7 for the CAM3, and for several satellite-derived products. The CAM3 exhibits sharply defined maxima for cloud liquid water in the deep tropics and at 60°N and 60°S. Simulated cloud water in the tropics and subtropics agrees most closely with the SSMI retrievals of Wentz et al (19XX), and represents a $\sim$ 30% overestimate of cloud water in the ITCZ for both the MODIS and NVAP retrievals. Cloud water in the extratropical storm tracks is approximately twice as large as in the ITCZ, and approximately twice as large as diagnosed by any of the available retrievals. The one exception is a new MODIS retrieval under development by members of the CERES Science Team, which shows high latitude cloud liquid water paths of comparable magnitude to the CAM3 simulation. Unlike the CCM3, the seasonal behavior of the simulated zonal average of cloud liquid water does not show a strong seasonal oscillation at high latitudes. The JJA simulation shows the strongest departure from the annual mean distribution, with slightly enhanced liquid water paths in the ITCZ and Northern Hemisphere high latitudes. Similar enhancements are seen in the various observational estimates, although the relative biases discussed earlier remain.

A quantity for which little in the way of globally observed data are available is ice water path. Zonally, annually, and seasonally averaged distributions of ice water path as simulated by the CAM3 are shown in Fig. 8. As is the case for cloud liquid water, cloud ice water exhibits large differences between the tropics and extratropics. Southern hemisphere extratropical ice water paths are nearly three times as large as in the ITCZ, exceeding 40 gm m^{-2}. Unlike the liquid water distribution, ice water has a very strong seasonal cycle at high latitudes, with maximum extratropical values occuring in the respective winter hemispheres. There is also a much stronger seasonal shift in cloud ice at low latitudes with much stronger tropical ice water loading during Boreal summer.

c. *Vertical Structure:*

Temperature and water vapor are the two state variables that jointly define the static stability of the atmosphere. The ability to properly simulate the thermal structure also plays an important role in the ability to properly simulate the water distribution. Fig. 9 shows the CAM3 annual zonal average differences of temperature and specific humidity differences from the ERA40 reanalysis climatology. Overall, the CAM3 does a relatively good job of reproducing the analyzed thermal structure. Simulated temperatures are within 1°K to 2°K of the analyzed field for most of the domain bounded by 50°N and 50°S. Overall, the CAM3 simulation represents a modest improvement over the CCM3. Tropical tropopause errors are nearly halved when compared to CCM3, and high-latitude lower tropospheric temperatures have been significantly warmed. A sizable part of the warming changes is associated with the increased horizontal resolution, most notably in high-latitude mid-to-upper tropospheric temperatures (see Hack et al., this issue). Despite these improvements, the difficulty in properly simulating polar tropopause temperatures remains, a longstanding documented problem for atmospheric general circulation models (Boer et al., 1992; more recent?).

Global observational estimates of the vertical distribution of water in the atmosphere are notoriously difficult to find, where analysis products provide the best available estimates. Even analysis products continue to contain large uncertainties in the moisture field (e.g., Trenberth and Guillemot, 1995; more recent?), since the water vapor distribution continues to depend upon the parameterized treatment of processes involved in the hydrological cycle. Nevertheless, comparison of the reanalysis product against locally available radiosonde observations suggest that the vertical structure of the biases shown in Fig. 8 are robust. These zonally averaged biases generally show a wetter than analyzed atmosphere throughout most of the domain. The major exception is the meridionally-broad low-level dry bias between 600 and 900 mb in the tropics, with biases exceeding 1 gm kg^{-1} in the zonal annual mean. The overall structure of the water vapor bias is similar to CCM3, but slightly exaggerated in magnitude. Preliminary analyses of this error suggest that the vertical structure in the tropics, such as the positive water vapor anomaly at 500 mb, is strongly determined by the form of parameterized moist convection. Further evidence that these large-scale biases are real is shown in Fig. 10, which illustrates vertical profiles of θ_e and specific humidity over Yap Island in the tropical west Pacific during the month of July. These figures show how well the ERA40 reanalysis compares to radiosonde data, and that the lower tropospheric dry bias and mid-to-upper tropospheric moist bias is a robust feature of the simulation. The dry bias maximizes near 850mb reaching 3 gm kg^{-1}. Water vapor biases of this magnitude and structure have a significant impact of the moist static stability of the tropical atmosphere, as is seen in the θ_e profiles, and are likely to play an important role in the low latitude dynamical response to diabatic heating.

The zonal average of the simulated vertical distribution of condensate is shown in Fig 11. The left panels show liquid water concentration, and the right panels ice water concentration. Most of the liquid water shown in Fig. 7 resides below 900 mb with concentrations ranging from 0.05 – 0.15 gm m^{-3} in the zonal annual mean. The mid- and high-latitude extratropics exhibit very strong vertical gradients, while the vertical distribution in the deep tropics is considerably more diffuse. The cloud ice water distribution generally reaches it's maximum concentration several km above the freezing level, between 500 mb and 600 mb in the extratropical storm tracks and near 300 mb in the deep tropics. Maximum ice water concentrations reach 0.007 gm m^{-3} and 0.003 gm m^{-3} in the respective zonal annual extratropical and deep tropical means. The high latitude seasonal swing in ice water path is primarily determined by changes in ice water concentration in the lowest kilometer of the atmosphere in the Northern Hemisphere. The Southern Hemisphere seasonal cycle is largely determined by ice water loading changes in the free atmosphere. Also, much of the high latitude cloud condensate consists of mixed phase clouds, whereas ice and liquid water regimes are more clearly separated in the tropics.

Finally, we illustrate the vertical structure of the meridional transport of water vapor in Fig 12. Mean meridional water transport is shown in the left panels, and transient meridional water transport is shown in the right set of panels. As shown in Fig 4, the deep tropics are a sink of moisture, the subtropics are a source of moisture, and regions poleward of 40°are sinks of moisture. As might be expected, the transport of water from the subtropics into the deep tropics is generally confined to the lower 1500 m of the atmosphere and largely handled by the mean meridional circulation. This transport exhibits the strong seasonal asymmetries associated with the Hadley Circulation (Trenberth et al., 2000). The majority of water vapor transport to higher latitudes occurs over a slightly deeper portion of the atmosphere, occurring in the form of transient eddy transport. Although much weaker than the equatorward transport by the Hadley Circulation, a third of the total poleward transport occurs in the indirect or Ferrel Circulation. At higher latitudes eddy transports toward the pole dominate transport by the mean circulation (Polar Cell), which acts to move water vapor from the polar regions to lower latitudes.

d. *Horizontal Structure:*

In this section, we will examine the horizontal distribution of vertically integrated measures of water and surface water exchange in the CAM3. We begin with the annually averaged precipitation field shown in Fig 13. Although the CAM3 simulation captures many of the observed features in the global precipitation distribution, it continues to share many of the same biases exhibited by the CCM3. Most of the available retrieval data agree that the most serious simulation errors occur in excessive precipitation over the western Indian Ocean, the central subtropical Pacific, and in the vicinity of Central America. The CAM3 also continues to underestimate the strength of the Atlantic ITCZ. The simulation also continues to exhibit a tendency for the simulated tropical precipitation maxima to remain in the Northern Hemisphere year round, and a slightly greater tendency for reduced precipitation along the equator, patricularly in the Indian Ocean. This is in sharp contrast with most satellite estimates, which show a clear seasonal migration of ITCZ precipitation across the equator, including the Indian Ocean.

The precipitation anomalies in the western Indian Ocean are related in part to deficiencies in the Zhang-McFarlane closure assumptions, a topic which has been explored by several investigators (Xie, 1999; Zhang, 200?). The most serious manifestation of these problems appears in the form of excessive precipitation rates over the Arabian Peninsula during the Northern Hemisphere summer months. Other factors contribute to this bias throughout the year, including the tendency to anomalously shift precipitation to the northern Indian Ocean during the boreal winter, coupled with an overactive and displaced precipitation regime to the west and northwest of Madagascar extending from the Mozambique Channel into the Indian Ocean east of Tanzania. Boreal summer also exhibits an extremely overactive precipitation simulation just north of the equator and 1000 km to the southwest of the Indian subcontinent. The excessive precipitation in the central Pacific subtropics is associated with two simulation challenges. The CAM3 continues to have difficulty in properly positioning the South Pacific Convergence Zone, which is too strong in magnitude and too zonal in structure, not extending far enough into the southern extratropics. This precipitation feature also has a tendency to extend too far east, another symptom of the tendency for the model to produce a double ITCZ. The northern Pacific bias is associated with a poor simulation of the very well defined precipitation pattern that extends from the South China Sea through the Philippine Sea and into the tropical equatorial Pacific during the months of July through August. This precipitation pattern is represented as a relatively diffuse extension of the southeast Asian Monsoon well into the central Pacific subtropics, and is a longstanding precipitation bias in the CCM and CAM models. Other notable features of the precipitation distribution is the inability to capture the seasonal cycle of precipitation off the west coast of Central America, and weaker than analyzed precipitation rates along the extratropical western boundary currents. Precipitation over the land surfaces generally tends to be excessive, especially over the Congo. Exceptions include large areas over the Amazon Basin, and much of United States east of the continental divide. Finally, simulated precipitation rates in the high-latitude extratropical storm track regions continues to be slightly higher than in current satellite retrievals suggest.

The simulated evaporation field (not shown) illustrates the important role played by the ocean surface as a source of water vapor to the atmospheric general circulation. As suggested by the zonal means, both the northern and southern oceans contribute to the evaporation of water vapor year round, but with important seasonal longitudinal migrations of evaporation centers. The simulation exhibits a clear evaporation minimum in the ITCZ year round, with extensive regions of high evaporation in the respective winter hemispheres. Boreal winter includes evaporation maxima along the western boundary currents (the Kuroshio and Gulf Stream), the Red Sea, the eastern Bay of Bengal and eastern equatorial Pacific, all of which exceed 10 mm day^{-1} ($\sim$ 290 W m^{-2}) in the 3-month seasonal mean. Other features include maxima in the subtropical western Pacific, the western equatorial Atlantic, and southern Indian ocean with evaporation rates $>$6 mm day^{-1} ($\sim$ 175 W m^{-2}). As was the case with CCM3, broad regions of evaporation are also seen over much of South America and Southern Africa exceeding 4 mm day^{-1} ($\sim$ 120 W m^{-2}) in the seasonal average. During the Boreal summer, the well defined evaporation centers transition to an extensive region of high evaporation across the southern oceans, with maxima in the Southern Indian Ocean and tropical west Pacific. Evaporation maxima in the northern Pacific ocean migrate eastward to the vicinity of the Hawaiian Islands with maximum evaporation rates of 6 mm day^{-1}. The principal evaporation regime in the Atlantic migrates into the southern hemisphere, and continental evaporation moves into the northern hemisphere, most notably eastern North America, India, large portions of east and southeast Asia, and sub-Saharan Africa.

Together, the evaporation and precipitation fields define the properties of fresh water exchange between the atmosphere and Earth's surface. The annually averaged horizontal distribution of $E - P$ is shown in Figure 14 for the CAM3. We note that a comparable global observational dataset does not exist. The principal tropical precipitation features are clearly visible. Local water deficits in the Inter-Tropical Convergence Zone generally exceed 4 mm day^{-1}) in the annual mean. The eastern Pacific subtropics, central Atlantic subtropics, and southern Indian subtropics, are the principal sources of water for the atmosphere. The CAM3 simulates a large seasonal cycle in $E - P$ over much of South America, central and southern Africa, India, and southeast Asia, mostly a reflection of the seasonal migration of deep convection in response to solar insolation. Similar seasonal variability is seen over most of Europe extending into central Asia, and over much of North America. Most of Europe and a large portion of North America can be clearly characterized as water source regions during JJA, and water sink regions during DJF.

The horizontal distribution of the annually-averaged precipitable water, and its difference from the NVAP climatology, is shown in Fig. 15. To a large extent, the CAM3 does a very good job of capturing the structure and correct magnitude of precipitable water in the atmosphere. There are, however, important large-scale systematic biases, despite exceptionally good agreement in the zonal mean structure. The longitudinally compensating arrangement of these biases is responsible for the good agreement in the zonal mean, where some of these regional biases are well correlated with biases in the precipitation distribution. Precipitable water is generally overestimated throughout most of the Pacific basin, in the western Indian Ocean, Arabian Sea, and central Africa. In sharp contrast, the simulation exhibits a large spatially-coherent dry region stretching from the Americas, across the equatorial Atlantic, Northern Africa, and into Southern and Southeast Asia. In general terms, the simulation is systematically dry over continental regions, most notably during the warm season. These water vapor biases are locally significant, particularly over Saharan Africa where they can exceed 10 mm, or often half of of the observed precipitable water.

Fig. 16 shows the annual global distribution of cloud liquid water and cloud ice for the CAM3 simulation. The extratropical storm tracks, features of the low latitude tropical circulation, such as the subtropical subsidence regimes, and continental deserts, are all clearly visible in the cloud liquid water field. The liquid water path frequently exceeds 200 gm m^{-2} in the storm tracks, in contrast with many of the satellite-retrieved cloud liquid water climatologies. Liquid water loading at low latitudes is generally in better agreement with satellite-derived values, although path lengths in the subtropical subsidence regimes are considerably smaller, particularly in the Southern Hemisphere. This is a surprising feature of the simulation, given the greater than observed cloud radiative forcing of these regions in the simulation. Cloud ice shares many of the same regional characteristics as cloud water. The tropical distribution is highly correlated with areas of deep convective activity, such as over the Congo, western Indian Ocean, Tropical west Pacific, and Amazon. As suggested by the zonal means in Fig. 8, significantly greater ice water loading is found at high latitudes in the storm track regions, where ice water paths are frequently well in excess of two times the maximum ice water paths seen in the tropics.

3. Low Frequency Forced Variability: Uncoupled Configuration

The seasonal cycle and the equatorial SST anomalies associated with El Niño-Southern Oscillation (ENSO) are two examples of major modes of low frequency variability in the climate system. These are essentially forced modes of variability in uncoupled integrations of the CAM3, and provide a useful basis for evaluating the simulated local and far-field responses as compared to what is observed.

The CMAP and GPCP analyses of global precipitation provide an observational opportunity to quantitatively examine the CAM3 simulated precipitation response to ENSO. Fig. 17 is a Hovmöller plot of precipitation anomalies as estimated by CMAP averaged over the deep tropics (averaged between 10°N and 10°S) and the CAM3 simulation of precipitation for the period January 1979 through December 2000. The CMAP product shows the evolution of strong positive and negative precipitation anomalies in response to the warm and cold phases of the observed ENSO cycle. Generally, the CAM3 does extremely well at capturing both the structure and amplitude of the anomaly pattern in the central and eastern Pacific. The eastward extension of the warm phase anomalies are well reproduced. The most serious weakness is in the simulation of the anomaly pattern in the western Pacific and Indian Ocean, which is much more weakly represented than observed.

A second way of examining the response of the hydrological cycle to ENSO is to explore the spatial pattern of the anomaly response associated with the time averaged precipitation difference between a specific warm and cold event. This approach also has the advantage of amplifying the response to the ENSO cycle. Fig 18 shows the monthly averaged precipitation difference between July 1994 (warm phase) and June 1999 (cold phase) as analyzed by GPCP and as simulated by CAM3. Both panels show a very large positive precipitation anomaly stretching across the central equatorial Pacific flanked by negative anomalies to the north, west, and south (in the SPCZ). The CAM3 simulation does a very good job of representing the structure and amplitude of the positive anomaly. The structure and magnitude of the negative anomaly response is not as well represented, particularly in the western equatorial Pacific and eastern equatorial Pacific. The western Pacific anomaly is too strong immediately north of the equator and too weak along the equator. This response pattern is consistent with the time-dependent response shown in the Fig. 17 Hovmöller diagram. Nevertheless, the CAM3 simulation does a remarkably good job of capturing the overall pattern and amplitude of the response, including the far-field response seen in the Atlantic and the western Indian Ocean. An important exception is the rainfall anomaly over the Amazon basin, which is very weakly represented.

Finally, we examine the ability of the CAM to simulate the seasonal migration of water vapor between the Northern and Southern hemispheres, which represents a regular major meridional redistribution of mass in the atmosphere, and has an impact on Earth's angular momentum budget (e.g., Lejenas et al., 199X). As seen in the zonal means of Fig. 5, the CAM3 correctly simulates a strong seasonal meridional migration of precipitable water. Fig 19 shows this seasonal redistribution of water vapor by subtracting the JJA distribution of precipitable water from the DJF distribution. Despite the biases discussed earlier, the seasonal redistribution of water vapor is well represented in the CAM3. The structure of the seasonal response is very similar to the observed climatology, even on relatively small scales. There are large-scale systematic biases, such as the slightly weaker seasonal cycle in the Northern Hemisphere, and slightly stronger seasonal cycle in the Southern Hemisphere, that lead to local differences in amplitude. But the overall properties of this mode of variability are well represented in the CAM3 simulation. These results demonstrate the ability of the CAM3's hydrological cycle to respond to lower-frequency externally imposed forcing.

4. Mean-State Simulation Properties: Coupled Configuration

In this section we provide an overview of the hydrological cycle as represented in CCSM3 coupled simulations (Collins et al. 2004a), which employ the CAM3 as the atmospheric component. The simulation we will discuss makes use of the CAM3 model, discussed in the earlier sections and by Collins et al. (2004c). We will examine the principal differences in the hydrological cycle as simulated by the atmosphere, along with major features of the hydrological cycle as seen from the perspective of the land, ocean, and sea ice component models. This discussion will employ a standard CCSM control simulation in which the atmosphere and land models are represented on a T85L26 transform grid, and the ocean and sea-ice models make use of a nominal 1°horizontal discretization. The T85x1 configuration of the coupled model is what has been used to document the CCSM3 simulations for international climate-change assessments (see Collins et al. (2004a)).

a. *Atmosphere:*

In an overall sense, the CCSM3 atmospheric global water and energy cycle budget remains remarkably similar to the uncoupled CAM3 simulation. The top of atmosphere energy budget remains within 0.2 Wm^{-2}, while the individual components of the surface energy balance remain well within 1 Wm^{-2} in the global annual mean. The global cycling of water in the simulated atmosphere is nearly identical to the characterization shown in Fig. 1 for the uncoupled model. The magnitude of the global hydrological cycle is reduced by approximately 1%, primarily due to a reduction in the magnitude of the water exchange over land and sea ice, but with comparable levels of runoff. Global annual storage of water vapor and condensate in the atmosphere also remains well within 1% of the uncoupled control simulation. Seasonally, these differences in measures of the global water cycle vary only slightly more than in their annual means.

Although most global annual metrics are virtually identical, the detailed regional behavior of the simulated hydrological cycle in coupled mode includes some notable differences. These anomalies can be seen in the zonal mean quantities related to the exchange and storage of water in the atmosphere (Figs. 2, 3, 4, and 5). There is a remarkable shift in the surface exchange of water from the Northern to Southern Hemisphere tropics. Northern Hemisphere tropical precipitation rates are reduced by 1 mm/day in the zonal annual mean, and are enhanced by more than twice this rate near 10°S. This meridional shift produces a significant and unrealistic change to the freshwater budget over the tropical oceans, most notably during the Boreal winter (see Section 4b). Although the precipitation anomaly appears in both the Atlantic and Pacific basins, the zonal mean anomaly is dominated by changes over the Pacific. This takes the form of an unrealistic enhancement of a southern and more vigorous branch of ITCZ convection extending across the basin from the warm pool to the Equador coast (see Fig. 20). The change to the precipition distribution is symptomatic of the so-called double-ITZC problem that plagues many coupled models (ref). Despite the overestimated precipitation rates in the southern tropical Pacific and south eastern tropical Atlantic, several other features in the precipitation distribution are significantly improved. These include precipitation over Central America and the Caribbean, along the western mid-latitude boundary currents, over the Arabian Peninsula, and over the Northern Indian Ocean. Precipitation reductions in the north central subtropical Pacific also represent modest improvements when compared to the uncoupled simulation.

Changes in the precipitation distribution are associated with a similar shifts in the storage of water in the atmosphere. Precipitable water moves from the Northern to Southern Hemisphere showing a double-peaked tropical distribution, which is dominated by anomalies that maximize in DJF. Large positive anomalies exceeding 10 mm appear in the south central tropical Pacific and south eastern tropical Atlantic. Negative anomalies of similar magnitude are located over much of the tropical and subtropical

Northern Hemisphere with maxima centered over the Arabian Peninsula and Central America. Generally speaking, changes to the precipitable water field represent additional degradations of the simulation when compared to observational estimates. The cloud condensate distribution reflects the changes to the distribution of precipitation and precipitable water. Cloud water and cloud ice follow the convective source regions, which have migrated to the southern oceans. These changes to the horizontal distribution of water strongly impact the energy budget at both the top of the atmosphere and surface. Large local anomalies are seen in both clear-sky and all-sky radiative fluxes at the surface and at the top of atmosphere, exceeding 40 Wm^{-2} for all-sky fluxes. Since precipitation and radiative processes are integrally involved in driving the tropical circulation, significant changes to the low level wind field are also seen in the central Pacific and eastern Atlantic. These changes give rise to a meridionally complex difference in surface latent and sensible heat transfers, further affecting the freshwater and heat budgets over the oceans.

An example of tropical variability of precipitation is shown in the right Hovmöller panel in Fig. 17. This shows much weaker tropical variability than in the uncoupled model. The response is related to the strength of CCSM simulated ENSO events (as opposed to specified ENSO events in the uncoupled simulation), coupled with the altered dynamical structure of the deep tropics and a tendency for most of the precipitation to occur away the equator.

b. *Ocean:*

Ocean transport of freshwater prevents the development of significant local trends in ocean salinity where mean net freshwater flux is strongly positive or negative. Net freshwater flux is mainly a function of precipitation and evaporation which have a geographic distribution determined by large-scale atmospheric circulations. The freshwater forcing of the ocean is thus predominately a function of latitude. The ocean's principle role in the hydrological cycle of the climate system is to transport the net positive freshwater flux resulting from precipitation in the tropics and high latitudes towards the midlatitude evaporation zones. The ocean also moves freshwater poleward from ice melting regions to ice forming regions where mean net freshwater flux is negative. Lastly, the ocean must redistribute the large, highly-localized influx of freshwater arising from river runoff.

Net freshwater flux into the ocean as a function of latitude is shown in Figure 21, which compares the fully-coupled CCSM3 control (T85x1) to an ocean-alone hindcast (DEFINE) solution (x1ocn) as well as to an estimate of actual freshwater flux (obs) derived from observed atmospheric and ocean datasets as described in Large and Yeager (2004). The x1ocn and obs curves track each other closely, since the major difference between the two is that the former couples observed atmospheric state fields to a fully evolving ocean model whereas the latter couples the same fields to an observed SST dataset. River runoff fluxes from both computations are identical and are based on the climatological gauge estimate scheme outlined in Large and Yeager (2004). In addition to differences in SST, these curves deviate at high latitudes because the ocean model incorporates ice formation and ice melt flux algorithms for which there are no corresponding observational datasets. Thus, at extreme polar latitudes, x1ocn shows large negative freshwater fluxes where ice formation results in brine rejection, and more positive freshwater input than obs near ice edge latitudes (65°S, 70°N) where melting takes place. These curves provide some measure of real freshwater fluxes, with and without polar processes included, which can be contrasted with the coupled model solution.

As in the uncoupled ocean, T85x1 freshwater flux is negative at high polar latitudes, and shows a jump to positive near the ice edge to values which exceed the observed flux estimate. Compared to both observationally-based benchmarks, there is excessive coupled freshwater flux to the ocean between $\approx 45^\circ - 65^\circ$ and less flux between $\approx 20^\circ - 45^\circ$, in both hemispheres. The Southern Hemisphere excess (A) results primarily from excessive precipitation together with a more equatorward peak in ice melt flux, which relates to overly extensive ice coverage in the Atlantic and Indian ocean sectors of the Southern Ocean (Holland et al). The midlatitude freshwater flux deficit in the South (B) arises from a lack of precipitation in comparison to observational estimates, as well as a mean increase in evaporative flux over these latitudes. The Northern Hemisphere extratropics are also characterized by excessive precipitation and more equatorward ice melt, but much of the excess freshwater flux in the vicinity of 60°N (C) is due to much higher river runoff fluxes near these latitudes. As in the South, there is less coupled precipitation and more evaporation between $\approx 20^\circ N - 45^\circ$N, resulting in a freshwater flux deficit (D).

The freshwater flux to the coupled ocean is most unrealistic in the tropics, where the double ITCZ creates a spurious peak in zonal mean precipitation at $\approx 10^\circ$S. A peak in T85x1 freshwater flux at the Equator (E) is related to colder SSTs which generate less evaporation in the Pacific along with excessive precipitation in the western equatorial Pacific and Indian Ocean. The positive flux bias between 10°S and Equator in Fig. 1 is exacerbated by much higher runoff from the Congo than is observed (Oleson reference?/Section 4c). The northern ITCZ is weakly simulated in the coupled model and this lack of precipitation together with low river runoff flux compared to the observed estimates between $\approx 0^\circ - 30^\circ$N leads to the slight freshwater flux deficit in the northern tropics (F). River runoff anomalies are related to precipitation anomalies over continents, as discussed in section 4c.

The biases in T85x1 zonal mean surface freshwater flux give rise to biases in the mean meridional transport of freshwater by the ocean. In Figure 22, the northward freshwater transports of the coupled and uncoupled ocean models, computed from Eulerian mean advection, are compared. Eddy transports are missing from these curves. Both models show poleward freshwater transport at high latitudes associated with ice growth/melt processes. While x1ocn shows significant freshwater transport south across the Equator (about 1/3 of which occurs in the Atlantic basin), the zonal mean freshwater transport across the Equator in T85x1 is very near zero. In coupled CCSM3, the negative freshwater flux zones at southern midlatitudes ($\approx 10^\circ - 40^\circ$S) are freshened via ocean transport from southern high latitudes as well as from the southern tropics, where freshwater input is higher than reality. Since tropical precipitation in nature is asymmetric about the equator, the x1ocn must transport freshwater southward across the equator in each ocean basin in order to compensate the southern midlatitude evaporation zones. The coupled CCSM3 ocean transports about as much freshwater northward across the equator in the Atlantic as x1ocn transports southward across the equator, and there is much less transport of freshwater southward across the equator in the Pacific.

There would appear to be an overly robust hydrological cycle in T85x1 in which excessive midlatitude evaporation in the Southern Hemisphere is related to excessive precipitation in the southern tropics as well as in the Southern Ocean. The ocean is thus forced to transport more freshwater poleward from the tropics than in reality, much of it northward. Excessive high northern latitude river runoff (cross-check with Oleson!) results in too much freshwater transport southward to the extratropics. The causes of the biased freshwater cycle are difficult to pinpoint, but are probably related to the lack of a damping mechanism which would inhibit air-sea freshwater exchange in way comparable to heat exchange, when the ocean becomes too fresh or salty. Subtropical SSS and SST in the south are fresher and warmer than observed (ref bias paper?). This suggests that the excess tropical precipitation transported to the subtropics by the ocean renders the midlatitude upper ocean too fresh and stable, thus inhibiting the deep mixing which would lower the SST. Anomalously high SSTs results in the excess evaporation which, after atmospheric transport back to the tropics, recurs as excess precipitation. Process studies indicate that, at least in the Atlantic, correcting the eastern ocean SST bias greatly reduces excess tropical precipitation (ref bias paper).

c. *Land:*

The hydrological budget over land in CCSM represents a balance between precipitation, evapotranspiration, runoff, and the change in storage in soils or snow. As seen in Table 3, there is no appreciable change in storage during the time period analyzed here. Both annual mean precipitation and runoff compare favorably with observations, with precipitation being about 3% high and runoff about 4% low if glaciers are included in the model estimate, and about right with glaciers excluded. We note that observations of runoff do not include most of Greenland and all of Antarctica. Purely observational estimates of global evapotranspiration are not available; however, from a variety of sources Brutsaert (1984) estimates evapotranspiration as 60-65% of precipitation. CCSM evapotranspiration is 63% of precipitation.

Evaporation from the ground is the largest component of evapotranspiration (59%) followed by canopy evaporation (28%) and transpiration (13%). Other estimates of the partitioning of global evapotranspiration suggest that transpiration should be the dominant component followed by ground evaporation and canopy evaporation. In particular, Choudhury et al. (1998), using a process-based biophysical model of evaporation validated against field observations, found that the partitioning was 52% (transpiration), 28% (ground evaporation), and 20% (canopy evaporation). Furthermore, since photosynthesis is coupled to transpiration through stomatal conductance, the underestimate of transpiration has implications for carbon assimilation in the model. Global photosynthesis is about 57 Pg C, which appears to be about 50% low (Table 3). The dominant form of runoff in CCSM is surface runoff (52% of total runoff), followed by drainage from the soil column (41%), and runoff from glaciers, lakes, and wetlands (7%). This latter runoff term is calculated from the residual of the water balance for these surfaces. This term may also be non-zero for other surfaces as well because the snow pack is limited to a maximum snow water equivalent of 1000 kg m^{-2}.

Table 3: Annual averages of global land precipitation (P), evapotranspiration (E), and runoff (R) (mm day-1). The components of E are transpiration (E_T), evaporation of canopy intercepted water (E_c), and ground evaporation (E_g). The components of runoff are surface runoff (R_S), drainage from the soil column (R_D), and runoff from glaciers, wetlands, and lakes and snow-capped surfaces (R_{GWL}) (mm day-1). Photosynthesis (PS) has units of Pg C. Observations for P are from Willmott and Matsuura (2000), R from Fekete et al. (2000, 2002), and PS from Schlesinger (1997). [1]Glaciers in Greenland and Antarctica are included in the model runoff. The observations have no data over these regions. [2]Excluding Greenland and Antarctica in the model.

	P	E	E_T	E_c	E_g	R[1]	R_S	R_D	R_{GWL}	R[2]	PS
CCSM3	2.12	1.33	0.18	0.37	0.78	0.79	0.41	0.32	0.06	0.82	57.2
Observed	2.06	—-	—-	—-	—-	0.82	—-	—-	—-	0.82	120

Zonal annual average values of the hydrologic cycle are shown in Figure 23. The CCSM simulation overestimates precipitation north of 45°N, generally underestimates it in the northern tropics, and overestimates it in southern South America. The latitudinal distribution of evaporation generally follows that of precipitation with a maximum in the tropics. Generally, the runoff biases coincide with those of precipitation suggesting that improvements in the simulated precipitation may lead to improvements in the runoff. At high latitudes, the primary active hydrological component is runoff. At other latitudes, ground evaporation generally dominates. An exception to this is in the tropics (10S-10N) where canopy evaporation is equally important. Transpiration is the smallest component of evaporation at all latitudes, even in the tropics where vegetation is densest.

Total runoff from the land model is routed to the ocean using a river transport model (reference?). Thus, biases in runoff have the potential to affect sea surface salinity and regional ocean circulation. The annual discharge into the global ocean is shown in Figure 24. Total discharge is 1.33 Sv. Discharge excluding Antarctica is about 1.25 Sv, which is about 6% higher than the estimate of Dai and Trenberth (2002). The river transport scheme does not account for loss of water due to human withdrawal or impoundment of water, seepage into groundwater, or evaporation from the river channel. In particular, consumption of water for irrigation may account for some of the discrepancy. Döll and Siebert (2002) estimate net and gross global irrigation requirements as 0.035 Sv and 0.078 Sv, respectively. The loss of freshwater from Antarctica is estimated to be 0.07 Sv by Vaughn et al. (1999), which is the same as that from CCSM. However, this comparison is fortuitous because the majority of Antarctic runoff from CCSM comes from the capping of snow over glaciers. More detailed glacier models need to be incorporated into CCSM to properly describe glacial processes including iceberg calving and basal melting.

Clearly, there are notable deficiencies in the modeled discharge at certain latitudes that arise from the land runoff fields. In particular, discharge from the Amazon and the Congo is 42% low and 109% high, respectively. The deficiencies in partitioning of evapotranspiration described previously are evident in the hydrological budget of the Amazon Basin (not shown). The partitioning of annual evapotranspiration in the model is 49% canopy evaporation, 30% ground evaporation, and 21% transpiration. As discussed in Dickinson et al. (this issue), too much water is intercepted by the canopy and re-evaporated in the wet season, thus resulting in limited water availability for plant roots particularly in the dry season. Photosynthesis exhibits a significant decline in the dry season, which affects the ability of the dynamic global vegetation model to correctly simulate the composition of vegetation in this region (Levis et al., this issue). The year-round warm bias in this region that is pronounced in the dry season is further confirmation that the simulation is deficient.

While improvements in the precipitation field supplied by the atmosphere model would likely improve the land hydrologic simulation in the Amazon Basin and globally, there are clearly aspects of the land hydrology that require attention. Current research is focused on improving the sunlit/shaded treatment of photosynthesis, stomatal conductance, and transpiration, and the parameterization of canopy interception.

d. *Sea Ice:*

As ice grows from sea water, it rejects salt back to the ocean, resulting in a relatively fresh ice cover with approximately 4ppt salinity. If ice dynamics is excluded and equilibrium climate conditions are considered, the local ice growth is balanced by local ice melt and the net long-term mean sea ice freshwater flux to the ocean is zero. Even under thermodynamic only conditions however, the considerable seasonal cycle of the ice/ocean freshwater flux can modify the ocean buoyancy forcing and influence ocean mixing. When sea ice dynamics are considered, the transport of relatively fresh sea ice redistributes water in the system, influencing the global hydrological cycle. This has the potential to modify the large scale ocean circulation in both the southern (e.g. Goosse and Fichefet, 1999) and the northern (e.g. Holland et al., 2001) hemispheres.

In the southern hemisphere there is net sea ice growth along the Antarctic continent which is then transported equatorward. Figure 25 shows the annual mean meridional ice transport simulated by the CCSM3 control integration. As the sea ice has only 4ppt salinity, this ice volume transport is nearly equivalent to a freshwater transport. The transport reaches a maximum of approximately 0.25 Sv at 65S. Estimates derived from satellite ice motion observations and sparse ice thickness observations (Weatherly et al., 1997) suggest a maximum value between 0.05 and 0.1 Sv. Compared to these estimates, the CCSM3 has excessive meridional ice transport in the southern hemisphere. The CCSM3 simulated southern hemisphere ice motion compares quite well to observed estimates (not shown). However, the ice thickness is excessive, particularly in the Weddell Sea (Holland et al, this issue), resulting in the high meridional transport. This excessive ice transport and melting along the ice edge, modifies the ocean sea surface salinity conditions, resulting in a fresh bias along the Antarctic sea ice edge in the south-western Atlantic.

In CCSM3, the long-term average freshwater storage in Antarctic sea ice equals 15,630 km^3. This corresponds to an annual average area of 12 million km^2 with an average thickness of approximately 1.4 m and a salinity of 4ppt. As discussed in Holland et al (this issue), the simulated area of Antarctic sea ice is large compared to observations, which have an annual average of approximately ??.

In the Northern Hemisphere, there is net sea ice growth in the Arctic basin, resulting in a net loss of water from the Arctic ocean. The Arctic ice is transported by winds and ocean currents and enters the north Atlantic through Fram Strait. This provides an important source of freshwater to the Greenland-Iceland-Norwegian seas and has the potential to influence oceanic deep water formation in this region (e.g. Holland et al., 2001). The annual mean flux of sea ice through Fram Strait in the CCSM3 T85-gx1 control integration is 0.08 Sv. This agrees well with the observed estimate of 0.09 Sv given by Vinje (2001). The flux has a considerable annual cycle (Figure 26) reaching a maximum value of almost 0.12 Sv in late winter when the ice thickness is at a maximum and the winds are at their strongest. As the ice volume flux depends on both the thickness and velocity of the sea ice, the good agreement with observations suggests that both of these properties are reasonably simulated. This does appear to be the case, as discussed further in Holland et al (this issue) and DeWeaver and Bitz (this issue). On the long-term average, the northern hemisphere CCSM3 sea ice covers 10 million km^2 with a mean thickness of approximately 2 m. Accounting for the ice salinity, this represents a freshwater storage of 18,450 km^3.

5. Summary

We have presented selected features of the simulated hydrological cycle for the CCSM CAM3 for both coupled and uncoupled applications of the model. The CAM3 exhibits a weaker hydrological cycle when compared with predecessor models, and closer in magnitude to observational estimates. The relative distribution of surface water exchange and atmospheric water storage by surface type is in good agreement with observational estimates. The detailed distribution of water in the simulated climate system reveals real systematic errors when compared to observations, where many of these biases are longstanding simulation challenges. The longitudinal distribution of precipitable water, and its vertical distribution, remain the most significant examples of these deficiencies. The tendency to form double-ITCZ structures in the deep tropics, and to inadequately simulate the seasonal meridional migration of tropical precipitation is also a continuing problem.

THERE ARE ALOT OF GOOD THINGS THAT SHOULD BE EMPHASIZED, AND DISCUSSED. MORE IS COMING, BUT IT WILL NEED TO WAIT TILL LATER.