Relative Importance of the Annual Cycles of Sea Surface Temperature and Solar Irradiance for Tropical Circulation and Precipitation: A Climate Model Simulation Study

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ABSTRACT: A recent version of the Goddard Earth Observing System GCM which contains several upgrades to the model’s prognostic cloud physics and microphysics as well as snow and ice hydrology, was used to isolate the influences of the annual cycles of solar irradiation and sea surface temperatures (SSTs) on the annual cycle of circulation and precipitation. Four 50-month-long integrations were produced with the GCM. The first integration, called the control simulation, C, was forced with daily interpolated SSTs from...

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a 30-yr climatology of monthly SST data. In this simulation both SSTs and incoming solar irradiance had their normally prescribed annual cycles. The next two companion simulations were called S1, which used annual mean prescribed incoming solar irradiation, and S2, which used annual mean prescribed SST; everything else was kept similar to C in these two simulations. In the fourth simulation, called S3, both SSTs and incoming solar irradiation at the top of the atmosphere were prescribed to always maintain their annual mean values. This constraint virtually eliminated all the annual cycle forcings in the simulation. Nevertheless, all simulations had the diurnal cycle of solar irradiation. An intercomparison of these simulations revealed the following.

1. The poleward excursions of the zonal mean ITCZ and precipitation are strongly modulated by the annual cycles of SSTs and solar forcings. For the majority of the regions, particularly in the subtropical monsoonal regions, for example, India, Southeast Asia, and Australia, the influence of the annual cycle of solar heating was found to be stronger than that of the annual cycle of SSTs.

2. The precipitation and circulation patterns over the Kuroshio Current region off the east coast of Asia were most affected by the SSTs and were strongly linked to the annual cycle(s) of local SSTs.

3. The annual mode of precipitation over Amazonia had two regimes: an equatorial regime with a maximum in the month of March in S1 and a corresponding maximum in the month of January in S2. Control C, which had both annual cycles, produced both the “month of January” and “month of March” modes of precipitation. This shows how solar and SST annual cycles jointly influence the simulated annual modes of precipitation over South America. Surprisingly, the annual modes of precipitation in C were roughly equal to the sum of the annual modes of precipitation of S1 and S2.

4. Precipitation over Sahelian Africa is significantly reduced in simulations lacking the annual cycle of solar irradiation. The opposite kind of influence of the annual cycle of solar radiation was noted in almost all other monsoonal regions: India, Southeast Asia, as well as Australia. The only exception is the continental United States, where the solar annual cycle showed only a relatively minor influence on the annual mode of precipitation.

5. The simulated tropical intraseasonal oscillations (TIOs) were reasonably robust in each of the four simulations. This suggests that TIOs are an outcome of the internal dynamics of the atmosphere that may in turn be forced by the interactions among the physical and dynamical processes of the atmosphere. This conclusion is consistent with the robustness of the observed TIO modes throughout the annual cycle and the significant dependence of TIOs on physical parameterization(s).

KEYWORDS: General Circulation; Radiative processes; Ocean/atmosphere interactions; Theoretical modeling

1. Introduction

It is common knowledge that the Sun is the ultimate cause of the dynamics of the Earth–atmosphere system; naturally therefore solar irradiance drives all bio-
The Earth's geophysical and thermal processes involve the Earth (including both land and ocean) and its atmosphere. However, the solar irradiance at the top of the atmosphere is essentially constant (varies by less than 0.1% over the decadal to centennial timescales) whereas the Sun's declination greatly changes during the course of an annual cycle and is responsible for large variations of incoming solar energy at different latitudes. This, in turn, leads to a multitude of interrelated annual cycles of the Earth's environment. These annual cycles are also found in each of the three major heat and moisture/water reservoirs, namely, the atmosphere, oceans, and land. These reservoirs absorb selectively the incoming solar fluxes and convert them into thermal and/or biogeophysical energy. The thermal energy, with diurnal and/or annual time lag, can appear either at the same location (as in the case of land), or emerge at a variety of distant locations due to hydrodynamic transports within the reservoir (as is the case for oceans and the atmosphere). The latter exhibits a much more complex time lag structure, which could vary substantially—anywhere from a few hours to an entire season or longer—even decades and centuries. These, in turn, produce a variety of complexities in the annual cycles of Earth's climate. Therefore, the annual cycle of solar irradiance is the primary source of energy for the entire internal dynamics of the Earth–atmosphere system and imparts a variety of timescales to the thermal forcing of different regions of the system, which, in turn, immensely complicates the outcome because of nonlinear feedback interactions among them.

There is a vast body of literature on the influence of the sea surface temperatures (SSTs) and the solar heating of Eurasian landmasses (e.g., the Tibetan Plateau) on the ensuing tropical circulation and Asian–African and Australian monsoons (e.g., Krishnamurti and Kishtawal, 2000). Studies by Gadgil (Gadgil, 2000), Kitoh et al. (Kitoh et al., 1999), Torrence and Webster (Torrence and Webster, 1999), Lau and Wu (Lau and Wu, 1999), and Chandrasekar and Kitoh (Chandrasekar and Kitoh, 1998) discuss several aspects of these interactions vis-a-vis Asian monsoons. Historically, some have viewed monsoons to be a gigantic sea breeze arising from the thermal contrast among Eurasian landmasses and tropical SSTs (Webster et al., 1998). Recently, Chao and Chen (Chao and Chen, 2001) argued that monsoons are essentially associated with the location of the intertropical convergence zone (ITCZ) and the related westerly outflow into the Rossby waves. On the other hand, some view the tropical intraseasonal oscillations (TIOs) with the associated northward propagation as the primary modulator of monsoon variations (Lau and Chan, 1986), while others claim to simulate monsoons irrespective of the realism of the TIOs in their model. A version of the Goddard Laboratory for Atmospheres (GLA) general circulation model (GCM) with a reasonable representation of TIO modes also captured the Indian monsoon variability quite well in the first Atmospheric Model Intercomparison Project (AMIP-I) simulations (Yang and Lau, 1998; Gadgil and Sajani, 1998). For these reasons, there is considerable uncertainty about how physical processes and/or surface forcings influence the tropical circulation, particularly the northward propagation of monsoons into higher latitudes.

The most dominant component of the external physical forcing of the Asian monsoons is the atmospheric annual cycle and the associated dynamics. Indeed, the annual cycle is a primary driver and determinant of all biogeophysical pro-
cesses that affect the terrestrial biosphere, SSTs, as well as the weather and climate of the Earth. Hence, understanding the influence of the annual cycle of solar irradiance (either as local land heating or as thermodynamically generated SSTs) on the Earth’s climate is the first step in unraveling the complex interactions among different physical forcings and responses. Even though the interannual variations of the dynamical climate, such as those forced by SSTs, for example, through the well-known El Niño/La Niña episodes, are equally important, the goal of this study is to discern the relative influences of solar and SST forcing on atmospheric circulation and dynamics through a short- (long-) response time-scale of land (ocean). The understanding of the relative response of the climate system to the solar and SST annual cycles (to be differentially prescribed in our simulation studies) would potentially decipher the complexities of the stationary and transient variations of the annual cycle of climate. Such considerations have provided the primary motivation for the investigation reported in this paper.

Atmospheric GCMs have been used extensively for simulating the influence of external boundary forcings on the development of weather and climate (seasonal means), hydrologic processes, and precipitation and its interannual variability (see, e.g., Houghton et al., 1996). We examine the influence of the annual cycle related changes in surface fluxes at the Earth’s surface on the annual mode of climate using a state-of-the-art GCM. We will examine these influences by suppressing the solar and/or SST annual cycles in otherwise similar sets of environments. The key requirement as well as the limitation of such an investigation is its reliance on a particular GCM; consequently, the chosen GCM must be sufficiently credible to realistically respond to the prescribed changes in external forcings. The climate version of the Goddard Earth Observing System, (GEOS), GCM fulfills these requirements. In view of the useful research and several ongoing upgrades (discussed in section 2) to the physical parameterizations of the GCM, the current version (4° × 5° horizontal resolution × 20 sigma layers for vertical resolution) is particularly suitable for the proposed investigation. The only other consideration is to arrive at a reasonable way to suppress the observed annual cycles of SSTs or solar forcing at the top of the atmosphere, which is described in section 3.

For many problems, particularly those that critically depend upon changes of a few watts per square meter in the systematic forcing of the atmosphere, the intrinsic limitations of a GCM and the random variability of the climate system can interfere with the statistical analysis of the simulated data and its effect on the confidence level of the findings. In other cases, such as Amazonian deforestation studies (e.g., Hahmann and Dickinson, 1997), the disagreement among GCMs in the simulated climate scenarios has raised some outstanding concerns, causing some key questions to remain unresolved. Therefore, one must ascertain that a GCM used for a specific study be suited for it. Nevertheless, one would like to eventually reaffirm the GCM-inferred findings with other state-of-the-art GCMs. The only earlier study on this subject is that of Shukla and Fennessy (Shukla and Fennessy, 1994), who used the Center for Ocean–Land–Atmosphere Studies (COLA) GCM to perform a similar, though not identical, simulation study. In that study the solar radiation and/or the SSTs were held constant at the equinox values for the subsequent boreal spring and summer simulations. The
documented version of their results (M. Fennessey, 2000, personal communication) primarily focused on the Indian monsoon and will be used for comparing our findings in the Indian region. Our paper endeavors to discern the influence of the annual cycle of SSTs and solar forcing on the annual cycles for all tropical circulation and precipitation. We will compare Indian monsoon–related findings with Shukla and Fennessy (Shukla and Fennessy, 1994); however, our diagnostics are broad based because we also examine several other consequences of the solar and SST annual cycles. To the best of our knowledge, these influences of solar irradiance and SST annual cycles have not been examined in this fashion in any previous modeling study. The rest of the paper is divided into the following four sections. The GEOS GCM description is given in section 2, the design of the simulation experiments is summarized in section 3, key results are described in section 4, and discussion and conclusions are provided in section 5.

2. General circulation model

The general circulation model utilized in this study is the GEOS GCM. It is a gridpoint global model and employs a staggered Arakawa C grid with a version of the Aries-GEOS Dynamical Core for its finite-difference algorithm. This algorithm invokes a fourth-order energy- and enstrophy-conserving scheme for hydrodynamic transports, which has been adapted for a nonuniform grid on the spherical Earth (Suarez and Takacs, 1995). The GCM has 20 sigma levels with \(4^\circ \times 5^\circ\) horizontal resolution. Any tracer can be advected in the model following the procedure used for advecting humidity. In the physics package, the model uses a relaxed Arakawa–Schubert (Arakawa and Schubert, 1974) cumulus convection scheme (Moorthi and Suarez, 1992, discussed in section 2.1), a parameterized planetary boundary layer (Helfand and Labraga, 1988) using Mellor and Yamada’s (Mellor and Yamada, 1974; Mellor and Yamada, 1982) level-2.5 turbulent closure, Zhou et al.’s (Zhou et al., 1996) gravity wave drag, and a few new upgrades to land surface and cloud dynamical and microphysical processes. This version of the GCM has two recent additions that are significant. One is the prognostic cloud-radiative forcing (CRF) invoked through the use of the microphysics of clouds with relaxed Arakawa–Schubert scheme (McRAS) (Sud and Walker, 1999a; Sud and Walker, 1999b; discussed in section 2.1). This upgrade is particularly relevant for assessing the influence of interactive clouds. The second upgrade is in the surface hydrology, specifically the influence of the snow/ice process on land (Mocko and Sud, 2001). The other features of the GEOS GCMs are (i) the prognostic fractional cloudiness and cloud optical properties and associated cloud microphysics and (ii) the ability to perform coordinate translation and rotation with a provision for relocating the mathematical poles to any arbitrary location (irrelevant for this investigation).

2.1 Cloud Physics in McRAS

Relaxed Arakawa–Schubert Scheme (RAS) based on the work of Moorthi and Suarez (Moorthi and Suarez, 1992) is the moist convection scheme of McRAS (Sud and Walker, 1999a). McRAS explicitly provides for three types of clouds:
convective, stratiform, and boundary layer. It uses a 10-min adjustment time step with an assumed relaxation timescale of 1 h. The location of the convective cloud base is diagnosed to be the top of the nearest layer from the surface of the Earth in which the relative humidity (RH) exceeds 90% of the critical relative humidity, RHcrit, of the large-scale clouds. This search is limited to the four near-surface levels, however. The buoyancy to carry the convective mass flux (with associated precipitation loading and momentum dissipation) to its detraining level is provided by the thermal energy of moist convection through the critical cloud work function (CCWF) constraint (Sud et al., 1991). These clouds coexist and interact with each other and have simultaneous life cycles invoking condensation, cloud generation and dissipation, and precipitation production. We briefly summarize different elementary modules of McRAS. Moist convection causes supersaturation and in-cloud condensation, which generates cloud water and fractional cloud cover using the cloud microphysics of McRAS.

Stratiform clouds can form if the grid-average RH exceeds the observationally estimated critical relative humidity, RHcrit (Slingo, 1987). There are three plausible scenarios: supersaturated RH, and RH more than, or less than, the RH needed to maintain the existing clouds. Each is handled differently. The first is straightforward; that is, excess RH over 100% turns into condensate and becomes cloud water. Naturally, in-cloud RH must always be maintained at 100%. If in-cloud RH drops below 100%, sufficient cloud water must be evaporated adiabatically to raise RH back to 100%. If the available cloud water is insufficient to maintain 100% in-cloud RH, all the cloud water evaporates and the entire cloud vanishes. On the other hand, if RH rises, both fractional cloudiness and in-cloud water increase following Sud and Walker’s (Sud and Walker, 1999a) formulation.

The boundary layer (BL) clouds are produced when the BL convection, which generally commences as dry convection, turns cloudy in its ascent toward the detrainment level at neutral buoyancy. This enables BL eddies to become supersaturated at or before detrainment. These eddies naturally provide BL clouds an ability to deposit any water vapor content (often more than its surroundings) into the detraining environment—a typical configuration of countergradient fluxes.

Conversion of condensate into precipitation follows the work of Sundqvist (Sundqvist, 1988). There is no special treatment for the ice phase beyond the implicit adjustment of the timescales and the empirical cloud microphysical constants, and saturation vapor pressure. Full cloud microphysics remains active at all times and it affects all cloud condensate(s) including the cumulus towers and anvils. Clouds in McRAS convect, diffuse, and advect both horizontally and vertically. The cumulus tower debris can produce grid-scale cloudiness and humidification regardless of the RH of the host/detraining layer.

The cloud destruction mechanisms are the same for all clouds. They include (i) diffusion of dry air into the clouds at subgrid scale, called cloud munching; (ii) evaporation of in-cloud water through convective-scale subsidence and associated adiabatic warming; (iii) cloud-top entrainment instability (CTEI) among adjacent cloudy and clear layers (Del Genio et al., 1996); and (iv) cloud-scale mergers including entrainment of ambient clouds into convective towers and downdrafts.
2.2 Convective downdrafts

The convective downdrafts and rain evaporation follow the work of Sud and Walker (Sud and Walker, 1993). It has been widely shown that convective downdrafts can have a significant influence on the simulation (e.g., Mandke et al., 1999). The precipitation falling inside cumulus towers (assumed to be saturated) does not evaporate. However, all the anvil precipitation, and some of the tower precipitation that emerges into the unsaturated environment, as a consequence of tilting of the convective tower, evaporates and is able to produce downdrafts if the excess negative buoyancy production criteria discussed in Sud and Walker (Sud and Walker, 1993) were satisfied. Downdrafts entrain ambient air and cloud water, which evaporates instantaneously because of the small size of the cloud water droplets and the relative humidity deficit produced by adiabatic warming within the subsiding downdrafts. In addition, the evaporation of tower precipitation, emerging beneath the cloud base, often satisfies the excess negative buoyancy criteria and leads to intense downdrafts. The statistical distribution of hydrometeors in idealized cloud geometry follows the work of Del Genio et al. (Del Genio et al., 1996) for water clouds and Ou and Liou (Ou and Liou, 1995) for ice clouds. For more details, the reader may refer to Sud and Walker (Sud and Walker, 1999a).

2.3 Land model

Our basic land model is the Simplified Simple Biosphere (SSiB) model, developed by Xue et al. (Xue et al., 1991). This model has been tested with several datasets, but land models of the present day need to be evaluated and improved region by region. A brief discussion of several improvements to the land model can be found in Mocko and Sud (Mocko and Sud, 2001). One also notes that SSiB parameterizes interlayer hydraulic conduction among soil layers invoking Richard’s equation with several assumptions; these help to simulate reasonable vertical fluxes. Indeed, the lack of subgrid-scale heterogeneity is an outstanding limitation of the current SSiB, but its amelioration is a separate issue and must be postponed at this time. SSiB was also evaluated with International Satellite Land-Surface Climatology Project (ISLSCP) Initiative I data under the Global Soil Wetness Project (GSWP) (Dirmeyer et al., 1999) as well as against other simple land schemes (Mocko and Sud, 1998). Only 10 different biomes are currently allowed in SSiB. Moreover, another restriction is that any single grid cell is allowed to maintain only one soil type and one biome. In other words, tiling is not feasible in the current design. However, some influence of gridscale variability can be introduced by varying the fractional vegetation covers and leaf area indices.

2.4 Radiative transfer

McRAS is designed to perform cloud-radiative forcing in a fully interactive and dynamical framework. It estimates cloud mass fraction, cloud droplet and/or cloud ice-crystal pathlengths, and effective radii of cloud particles. The in-cloud water and ice mass fractions are diagnosed from cloud temperature. One modification is in the calculation of equivalent plane-parallel optical thickness following Ca-
halan’s (Cahalan, 1994) correction for cloud water inhomogeneity. The second modification is in the diagnosed number density and effective radius of ice clouds as a function of temperature following Lohmann et al. (Lohmann et al., 1999). The structure of this distribution for an arbitrary in-cloud water substance is shown in Sud and Walker (Sud and Walker, 1999a).

There are four bands in the shortwave radiation (Chou et al., 1998) and nine bands in the longwave radiation (Chou et al., 1999). For each band, we require the optical thickness, single scattering albedo, and asymmetry factor for the clouds (see Table 1 for optical parameters). The linearized approximations to the precise radiative transfer equations are given in Chou and Suarez (Chou and Suarez, 1994). The optical thickness, single scattering albedo, and asymmetry factor for in-cloud water and ice mixtures employ a weighted summation.

### Table 1. Parameters for cloud-radiation interaction.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Unit</th>
<th>Land value</th>
<th>Ocean value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Effective radius</td>
<td>$r_e$</td>
<td>$10^{-6}$ m</td>
<td>7.0</td>
<td>10.0</td>
</tr>
<tr>
<td>(water)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Number density</td>
<td>$N$</td>
<td>cm$^{-3}$</td>
<td>170.0</td>
<td>60.0</td>
</tr>
<tr>
<td>(water)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Effective radius</td>
<td>$r_e$</td>
<td>$10^{-6}$ m</td>
<td>25.0</td>
<td>25.0</td>
</tr>
<tr>
<td>(ice)*</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Number density</td>
<td>$N$</td>
<td>cm$^{-3}$</td>
<td>0.06</td>
<td>0.06</td>
</tr>
<tr>
<td>(ice)*</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* Also from Ou and Liou (Ou and Liou, 1995) and Moss et al. (Moss et al., 1996). See Fig. 2b for details.

The clouds in any atmospheric column are divided into three height groups: (i) surface to 700 hPa, (ii) 700 to 400 hPa, and (iii) 400 hPa to the top of the model’s atmosphere. Within each group, the clouds are assumed to be maximally overlapped, while among different groups, they are assumed to be randomly overlapped. For optical thickness of different clouds within the group, smaller clouds are smeared to the size of the largest cloud within the group; this entails adjustment of the in-cloud optical thickness for each cloud except the largest. The specific calculation differs somewhat for shortwave and longwave radiation. For longwave radiation, the effect of backscattering is folded into the emission of an atmospheric layer, and absorption between the levels is calculated by scaling the cloud optical thickness appropriately. With these approximations, the longwave radiative transfer equations for a cloudy atmosphere are identical to that for a clear atmosphere. More details can be found in Chou et al. (Chou et al., 1998; Chou et al., 1999).

### 3. Design of simulation experiments

We performed four separate 50-month-long simulations (Table 2) with the GEOS GCM. In the control simulation, C, the solar radiation at the top of the atmosphere was calculated hourly while the SSTs were interpolated from 12-monthly means of SST analyses averaged for 30 yr; thus, all the usual annual cycles and solar radiation forced diurnal cycles are simulated over land. The particular choice of 30-yr mean SSTs for the annual cycle makes a comparison of a model simulation
with observations impossible, but for that, a large number of model validation studies (e.g., Sud and Walker, 1999b) can be cited. On the positive side, however, climatological SSTs eliminate all extraneous interannual interactions except for the annual cycle of SSTs.

**Table 2. The simulation experiments.**

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Name</th>
<th>Period</th>
<th>Starting day</th>
<th>Initial soil moisture</th>
<th>Surface albedo and vegetation variations</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control; annual cycle solar radiation and SST</td>
<td>C</td>
<td>50 months</td>
<td>1 Jan</td>
<td>GSWP analysis, 1987</td>
<td>Annual cycle albedo; annual cycle vegetation</td>
</tr>
<tr>
<td>Annual mean solar radiation with annual cycle of SST</td>
<td>S1</td>
<td>50 months</td>
<td>1 Jan</td>
<td>GSWP analysis, 1987</td>
<td>Annual mean albedo; annual mean vegetation</td>
</tr>
<tr>
<td>Annual mean SST with annual cycle of solar radiation</td>
<td>S2</td>
<td>50 months</td>
<td>1 Jan</td>
<td>GSWP analysis, 1987</td>
<td>Annual cycle albedo; annual cycle vegetation</td>
</tr>
<tr>
<td>Annual mean solar radiation and SST</td>
<td>S3</td>
<td>50 months</td>
<td>1 Jan</td>
<td>GSWP analysis, 1987</td>
<td>Annual mean albedo; annual mean vegetation</td>
</tr>
</tbody>
</table>

Here, C was complemented by three companion integrations. In the first integration, called S1, the annual cycle of the solar irradiation was removed, while the SST annual cycle was maintained identically to that in the control. The monthly zonal means of solar irradiation at the top of the atmosphere are shown in the left panel of Figure 1. Moreover, it shows three lines representing time mean incoming solar irradiance in the right panel of Figure 1: boreal summer (June–July–August, JJA; red line), boreal winter (December–January–February, DJF; yellow line), and the annual mean (blue line). The annual mean incoming solar irradiance was used in S1. The annual cycle of SSTs was suppressed in simulation S2. This was done by generating a 12-month annual mean of SSTs from the 30-yr monthly mean climatology of SSTs that was used in C. This SST climatology provided a single annual mean value of SST for each grid point (Figure 2) that was prescribed in this simulation while its solar irradiance for both annual and diurnal variations was maintained identically to that of C. The boreal summer (JJA) and the boreal winter (DJF) SST anomalies are shown in Figure 2b and Figure 2c.

With the SST annual cycle removed, daily interpolation of SST became unnecessary. However, if changes in SST patterns generate accompanying anomalous cloudiness patterns, they would clearly interact with the solar and longwave transmissions. In the last integration, simulation S3, the annual mean solar irradiance of S1 and the annual mean SSTs of S2 were prescribed simultaneously. Consequently, there is no annual cycle forcing in this integration and we do not expect to see any annual cycle after the usual adjustment period of a couple of
months. One might ask the following question: how does annual mean solar irradiance compare with the Sun held at equinox (e.g., as is done in Shukla and Fennessy, 1994)? We argue that a persistent imbalance in the annual cycle of solar heating over a 50-month-long period of integration, such as used here, is likely to cause some violence to the surface temperature, soil moistures anomalies, and the hydrological cycle. In other words, it may produce some regions with excessive drying, while others may stay too wet and together they might sacrifice the annual character of the hydrologic processes. However, as we will show, the influence of solar and SST forcings on the Indian monsoon turned out to be similar to that found by Shukla and Fennessy (Shukla and Fennessy, 1994). Hence, the simulated influences of the annual cycles of SSTs and solar irradiance on the Indian monsoons appear to be fairly robust.

All of our simulations started from an analyzed initial condition for 1 January 1987. The control plus the three additional companion simulations (Table 2) were sufficient to isolate the effects of the annual cycles of solar heating and SSTs. We have examined several diagnostics and simulation statistics to isolate the influence of the annual forcings on the annual and higher-frequency modes of simulated circulation and precipitation. We did not perform ensemble runs because the prescribed changes in atmospheric forcings were so large that the intrinsic variability of the model is not expected to influence the outcome of such a fundamental study.
4. Results

The control run, C, plus the three companion simulations, S1 (without solar annual cycle), S2 (without SST annual cycle), and S3 (without both the solar and SST annual cycles), have been described in section 3. Clearly, large changes were introduced in solar radiation and/or SST forcings of S1, S2, and S3; therefore,
we ignored the first 2 months of each model integration as an initial adjustment period for significantly perturbed forcings. Consequently, most of our analyses are based on the remaining 48 months, or a subsequent period of 4 yr. The internal dynamics is the only cause of interannual variability of the simulations since both major external forcings were prescribed. Because the soil moisture is interactive in these simulations, it can introduce a systematic drift in the hydrologic cycle by excessively drying or moistening different regions. This must be kept in mind when interpreting the simulation results. In the following sections, JJA and DJF fields are analyzed to isolate the influence of the annual cycles of SST/solar forcings on the annual cycle of circulation and precipitation.

4.1 Time series of zonal mean precipitation

Figure 3 shows the time–latitude fields of zonal averaged monthly precipitation for each of the four simulations. The north–south excursions of the ITCZ precipitation for C, S1, S2, and S3 can be seen in the four panels of Figure 3 from top to bottom, respectively. The zonal excursions in S1 are relatively weaker as compared to C; however, as compared to S2, they are much stronger in the Tropics and weaker at the midlatitudes, particularly in the Northern Hemisphere, which has more landmass (Figure 3c). Clearly, annual cycles of SST (solar irradiance) exert a significant influence on the ITCZ as well as surface fluxes; both of which affect the zonally averaged precipitation. Since both SST and land respond to the annual cycle of the solar radiation, while land cover is only 30% of the Earth’s area, one might expect its influence to be about 30% of the total, which turned out to be roughly true in these comparisons. Thus the SST annual cycle has a much stronger influence on the meridional excursion of the tropical ITCZ and associated precipitation. In the higher latitudes, particularly in the Northern Hemisphere characterized by its larger landmasses, the solar annual cycle corresponds significantly with the annual cycle of precipitation. Here a flat and broad rainy region in S2 replaces the ITCZ excursions. Simulation S3, with no annual cycle for the SST or solar irradiance, was expected to produce a steady-state response and indeed it does (Figure 3d); the small vacillatory variability merely reflects the nonlinear internal dynamical influences, such as the influence of TIOs or other atmospheric wave activity.

4.2 Global precipitation

The 4-yr JJA (DJF) mean precipitation fields are shown in Figure 4 (Figure 5). In accord with Figure 3a, the boreal winter and summer ITCZs along with the associated precipitation fields are well simulated in C; see Figure 4a (Figure 5a). In JJA, S1 shows an ITCZ similar to that of C particularly in its structure across the tropical Pacific Ocean but with large differences in the coastal regions, for example, tropical South America. However, the simulated precipitation over India and China increases (decreases) in response to the annual cycle with JJA (annual mean) solar irradiance. In addition, the simulated precipitation is less over the warm pool region of the tropical western Pacific (Figure 4b versus Figure 4a). In contrast, the simulated JJA precipitation over tropical Africa increases (decreases) in S1 (S2) as compared to C; this will be discussed at some length later. Summer
Figure 3 Monthly zonal monthly precipitation (mm day$^{-1}$) for four 50-month integrations: (a) control, C; (b) simulation S1 with solar irradiance held at the annual mean; (c) simulation S2 with SST held at the annual mean; and (d) simulation S3 with both incoming solar and SST held constant at the annual mean value.

solar irradiance, on the other hand, produces intense heating of the Eurasian landmasses, which can be expected to draw trade wind moisture transports into India. This could be viewed as thermally driven and frictionally controlled moisture flux transport. Lacking the strong summer solar insolation in S1, the moisture flux transport converges partly over the ocean and is partly diverted to Africa whereby
Figure 4. Summer (JJA) precipitation simulated over the 4 yr for (a) C, (b) S1, and (c) S2. Contours are drawn for 1, 2, 4, 6, 8, 12, and 16 mm day$^{-1}$.

the precipitation over tropical Africa increases (Figure 4b). On the other hand, S2, with the full annual cycle of the solar forcing, but without the SST annual cycle, generates relatively more precipitation over India and China and less over Africa as compared to S1. However, the monsoons-on-land (over India and China) features of S2 are midway between S1 and C. This suggests that SST and solar
insolation annual cycles must exert a positive feedback influence on each other. In the third integration, called S3, without the annual cycle of SST and incoming solar radiation, African precipitation was strong and robust though somewhat to the south, while the Indian monsoon did not develop at all (not shown). For DJF, S1 had too much (little) solar insolation in the Northern (Southern) Hemisphere. Accordingly, it produces less convergence and precipitation over land for

Figure 5. Same as in Figure 4 except for winter (DJF).
the Southern Hemisphere; see, for example, Australia and extratropical South Africa and South America. Without the SST annual cycle in S2, both the JJA and DJF rainfall patterns on land (solar insolation forced) are closer to those of C; but there are large differences over the ocean, where the SST annual cycle would be important. Comparisons of midlatitude land precipitation in DJF of C and S1 also show large differences (Figure 5a and Figure 5b). The corresponding differences over the oceans are larger due to the lack of the SST annual cycle (Figure 5b and Figure 5c in comparison to Figure 5a). If the GCM can be assumed to simulate realistic rainfall scenarios, the summer solar heating of land implicit in the annual solar cycle leads to copious (deficient) precipitation over India (Africa). Indeed, without such a solar heating of large landmasses of the Northern Hemisphere, all major precipitation distribution centers tend to position themselves southward, particularly over India. In this way, the annual solar and SST cycles appear to work in concert for producing the simulated Indian monsoon rainfall in JJA.

For boreal winter (DJF), we see that the Australian monsoon (Figure 5a) also reveals some interesting characteristics of the combined influence of solar and SST forcings. When invoked in isolation, neither the SST annual cycle of S1 (Figure 5b) nor the solar insolation annual cycle of S2 (Figure 5c) was able to produce the observed Australian monsoon. Even when the two were added (not shown), they did not sum up to the Australian monsoon precipitation and accompanying circulation simulated by C (Figure 5a). We infer that the SST–solar insolation feedback interactions are also important for the Australian monsoons. The tropical African precipitation is strongest without the solar annual cycle. It is somewhat weaker, but nevertheless stronger than that of C, for S2 without the SST annual cycle. Over Amazonia, the precipitation gets a typical three-center structure in response to both the SST and solar annual cycles. The 4–6 mm day$^{-1}$ precipitation over the North Pacific is substantially reduced (maximum of only 4 mm day$^{-1}$) in S1. In other words, the results show that the spatial distributions of precipitation are strongly linked to the solar annual cycle. We shall examine this further in section 4.6. There are some other differences in the ITCZ, the South Pacific convergence zone (SPCZ), as well as in the precipitation over the Himalayas. For example, the Himalayan precipitation, which is attributed to wintertime disturbances, shows a substantial decrease in S1 but without much change in the structure, and it develops a somewhat different structure in S2. We will not discuss these and other global-scale consequences at this time.

### 4.3 Hadley cells

Each of the four Hadley cells for boreal summer (JJA) shows a consistent response to the annual cycle of solar and/or SST forcings. The simulated Hadley cell is most realistic as well as the strongest in C (Figure 6a). It gets significantly weaker in simulations without the solar annual cycle, S1 (Figure 6b). In S2, without the SST annual cycle, the Hadley cell (Figure 6c) is much stronger than that in S1, but it is only somewhat weaker than that in C. This again suggests that boreal summer season solar heating is imperative to the development of a strong Hadley cell. However, if both the SST and solar annual cycles are eliminated, the simulated Hadley cell is the weakest among all the simulations (S3;
Figure 6. Summer (JJA) Hadley circulation (109 kg s⁻¹) for the 4-yr mean model integration for (a) C, (b) S1, (c) S2, and (d) S3.

Figure 6d). This is to be expected. For this reason, we reaffirm that both the solar and SST annual cycles work in unison to produce the observed strength and structure of the Hadley cell of JJA. The second equally interesting feature is the Ferrel cell. Let us start with the Northern Hemisphere. In the absence of summer solar heating, particularly over land, the air mass at about 30°N (with large landmass) cools and sinks to induce compensatory adiabatic warming that must make up for the deficit in solar and condensation heating that maintains the observed horizontal temperature gradients and vertical lapse rates. This subsidence must be larger in simulations lacking JJA solar insolation. It strengthens the indirect Ferrell cell significantly (Figure 6b and Figure 6d) for S1 and S3, both of which lack the boreal summer Sun. On the other hand, Figure 6a and Figure 6c show that the solar annual cycle with strong heating of landmasses in the Northern Hemisphere is responsible for the observed structure of the Hadley cell. Figure 6c and Figure 6d show a relatively weaker Ferrell cell in the Southern Hemisphere; both of these simulations (S2 and S3) use an annual mean SST. Since there is large ocean cover over the southern midlatitudes, such an influence of SST on the meridional circulation is an expected consequence.

The corresponding zonal mean diabatic heating, which is the sum of radiative and condensation heating fields for JJA (Figure 7), shows how the diabatic heating relates to the mean meridional circulation fields of Figure 6. First, Figure 7a and Figure 7c are much more similar to each other in the tropical regions, suggesting the importance of the JJA insolation for tropical diabatic heating. The annual
mean solar forcing implies more (less) sunshine than in DJF (JJA). Increased land cooling (Figure 7b and Figure 7d) produces springlike solar forcing, particularly over land, as compared to C and S2 (Figure 7a and Figure 7c). The similarities between Figure 7a and Figure 7c, and also between Figure 7b and Figure 7c, provide a qualitative assessment of the influences of the annual cycles of solar insolation and SST, respectively. The least diabatic heating or cooling occurs in S3 (Figure 7d), with no annual cycles of SST or solar irradiance; moreover, it also corresponds with the weakest Hadley circulation (Figure 6d). There are several other features, such as large differential heating or cooling at the polar latitudes in response to changes in the incoming solar radiation between the annual mean versus JJA insolation, but such differences are self-evident. It may be pointed out, however, that the changes in the mean meridional circulation and diabatic heating structures are internally consistent and reasonably explainable. Similar features with 6-month phase, that is, Northern Hemisphere features replacing Southern Hemisphere features and vice versa, were noted for DJF periods and are not discussed here.

4.4 Indian monsoon

4.4.1. Seasonal circulation

A “wet summers and dry winters” type of climatology generally characterizes Indian monsoons. The classical view has been that solar heating of land and the
lower troposphere produces a thermal low (during boreal summer season) that helps to establish the Indian monsoon. The water vapor transport and rising motion generated by the solar heating of the subcontinent supports moist convection, which releases the latent heat of condensation to exert a positive feedback on the maintenance of the monsoon circulation and precipitation. The condensation heating in turn positively feeds back on the trade winds, which converge into India from the west replacing the local winter/spring season easterlies. The influence of the annual cycle of solar forcing and SSTs on the Indian monsoon is clearly demonstrated in the JJA precipitation and winds (Figure 8a, Figure 8b, and Figure 8c). Without heating by the boreal summer Sun, the 4-yr mean precipitation of C (Figure 8a) is markedly reduced over the Indian subcontinent (S1; Figure 8b) as well as in S3 (not shown). The typical monsoon westerlies that usually emanate from the Indian Ocean into the Indian subcontinent do not develop in S1. In fact the entire monsoon circulation, except for westerly flow from 5°S to 5°N, is in disarray (relative to typical observational patterns). Naturally, there is very little precipitation over coastal western India (including the Ghats) and there is no discernible summer monsoon. Nevertheless, more precipitation accompanies the monsoonlike circulation over the equatorial Indian Ocean. In S2, with the annual cycle of solar forcing, that is, with JJA solar forcing, the GCM is able to simulate discernible signatures of the Indian monsoon even without the help of JJA SST (Figure 8c). From these results, we infer that solar heating of the landmasses in the summer season is the primary driver of the Indian monsoon circulation into northern India, which has also been the classical concept. If we look at the differences between the precipitation fields of Figure 8a and Figure 8c, we note the relatively secondary importance of the SST annual cycle (not shown). In these experiments, the influence of the SST annual cycle on the temperature and circulation of the landmass of the Northern Hemisphere is relatively smaller than the corresponding solar annual cycle. Although annual cycles, SST, and solar insolation play a positive feedback role on the strength of the simulated Indian monsoon, their comparative importance is abundantly clear from these simulation studies. Without the intrinsic support of solar heating of land, the annual cycle of SST is not able to invigorate the Indian monsoon (S1), while the JJA solar insolation helps significantly to simulate the structure of the monsoon circulation (S2). We infer that it is the land heating by the Sun during the local summer season that draws primarily the trade wind moisture into the Indian subcontinent causing winds to change from easterly to westerly, which causes the onset of the Indian monsoon. On the other hand, the simulated tropical African precipitation is larger in S1 as compared to S2. Indeed, if solar insolation is the main driver, how can one explain the SST-modulated rainfall changes? The answer is that the monsoon circulation and associated rainfall is also there, but the SST anomalies, through oceanic circulation, are able to displace the positions of the centers of tropical rainfall by several hundred kilometers, and that is enough to produce major droughts in large tropical regions such as India or China. Lacking JJA solar forcing, the simulated tropical African precipitation increases while the Indian monsoon gets weaker—a simulation result that could be evaluated in observational data by aggregating years in which Indian and African JJA precipitation could be compared to determine the existence of the above relationship. These
Figure 8. Summer (JJA) precipitation (mm day\(^{-1}\)) and 850-hPa winds (m s\(^{-1}\)) as a mean of the 4-yr simulation for (a) C, (b) S1, and (c) S2. Precipitation isolines are drawn for 1, 2, 4, 6, 8, 12, and 16 mm day\(^{-1}\).

Results of the relative response of the Indian monsoon to the boreal summer SSTs and solar insolation are in general agreement with those of Shukla and Fennessy (Shukla and Fennessy, 1994).
4.4.2 Pentad of precipitation for the Indian region

The pentads of the Indian monsoon season (from the months of May through September) were examined over India and the nearby Indian Ocean on a weekly timescale. Selected sample plots of the simulated precipitation for five months (May, June, July, August, September) for each of the three simulations (C, S1, S2) are shown in Figure 9a, Figure 9b, and Figure 9c. These were also compared against the Data Assimilation Office’s (DAO) Data Analysis System (DAS) reanalysis. In C, the Indian monsoon, which is at about 10°N in the beginning of May, moves to engulf the entire subcontinent by mid-June, a scenario that is well documented in the observations and is also evident in the DAO DAS data products. Nevertheless, the plots depict averaged precipitation and circulation for the 4-yr period starting from 1 January of the first year. The monsoon of simulation C withdraws around the third week of September (not shown), whereas in the DAO DAS the precipitation over the Indian subcontinent persists until the first week of October. Both scenarios are plausible; however, we must remember that our control simulation with climatological SSTs does not represent any observed period and therefore it cannot be truly compared with the DAO DAS data products that pertain to a specific time period. The main purpose of comparing and analyzing these precipitation patterns is to determine if the model simulates a
respectable onset and withdrawal of the monsoon, which was evident in Figure 9a. In contrast, there is virtually no monsoon onset or development in S1 (Figure 9b). On the other hand, S2 (Figure 9c) has a sudden monsoonlike onset and withdrawal, even though the onset is somewhat delayed and much weaker. The DAO DAS had a much earlier onset of monsoon, which was also much stronger and longer lasting (not shown). Clearly, Indian monsoon data in DAO DAS are at some variance with the current GEOS GCM simulation, as well as with documented observations; however, the key point to emphasize is that the climate version of GEOS GCM simulates a reasonably realistic time mean character of the Indian monsoon. This potentially gives us confidence in the reliability of our simulation studies and the key inferences based on the model simulations. In summary, the pentad pictures show that monsoons can go much farther into northern India including into the Himalayas in S2 with the JJA insolation but with annual mean SST, as compared to S1, which employs annual mean insolation with JJA SST. This shows that the annual cycle of the Sun is even more important than the annual cycle of the SSTs, particularly for India. This finding is in agreement with Shukla and Fennessy (Shukla and Fennessy, 1994). We explore further the latitudinal extent on the Indian monsoon in section 4.4.3.

4.4.3 Time–latitude precipitation over the Indian longitudes

The precipitation over 65°–90°E, the longitude of the Indian subcontinent plus the surrounding oceans of the region, is analyzed further by examining the zonal (65°–90°E) average precipitation time series. The simulated annual cycles of precipitation for the three simulations (C, S1, and S2) show that in the summer the northward extension of the monsoon precipitation that engulfs the Indian subcontinent is clearly forced by solar heating of land (see Figure 10a and Figure 10c). Figure 10a shows a sudden onset of the monsoon in mid-May, its maintenance until the end of August, and then a slow subsequent withdrawal. It is similar to what has been documented in the analysis of the precipitation observations. For the annual mean insolation simulation, S1, the annual cycle of precipitation does not have any development in the summer season (Figure 10b). In simulations C (Figure 10a) and S2 (Figure 10c), both invoking the annual solar cycle, there is a distinct region of heavy precipitation between 5° and 20°N during JJA around the latitude of the Indian landmass. The simulation without the annual cycle of SST (Figure 10c) shows, in addition, a distinct deficit in precipitation in the equatorial region, which is not found in either C or S1 (Figure 10a and Figure 10c, respectively). We can safely infer that the lack of the SST annual cycle is the cause of this precipitation deficit. We compared the simulated precipitation vis-à-vis the analyzed precipitation (not shown) and found that the model potentially does a reasonable job of simulating precipitation over the Indian subcontinent during the summer monsoon season. This too gives us reasonable confidence in the above inferences. Moreover, the precipitation at about 30°N, which is evidenced in Figure 10a and Figure 10c, is virtually missing in Figure 10b. Thus boreal summer insolation not only is responsible for the Indian monsoon, but it also affects precipitation over the land region as far as 30°N and beyond.
4.5 Annual cycle of precipitation

The structure of the annual cycle of precipitation can be discerned by Fourier decomposition of the monthly precipitation at each grid point independently. Clearly, the first harmonic is the annual mode. The global distribution of this...
mode for the control simulation, C (Figure 11a), shows a JJA phase over India, China, and large parts of Southeast Asia. It represents a well-simulated placement of boreal summer monsoons. We also see a boreal summer (winter) mode over northern (southern) Africa. There are some equally strong annual modes of precipitation over the Indian, Pacific, and Atlantic Oceans. Figure 11a also reveals
their relative phases and strengths. Warm water of the Gulf regions along the east coast of North America produces a boreal winter mode, while North and South Pacific regions have some very strong modes that reverse phase over a short distance. In the ITCZ region over the equatorial Pacific (100°–175°W), the phases of precipitation to the north and south of the ITCZ merely reflect the 6-month out-of-phase relationship of the ITCZ in the boreal summer and winter. The eastern Pacific annual mode shows a gradual advance in the phase of the summer mode (extending to September) as we go northward from the equator; the case for the southeastern Pacific is quite the opposite, but if the phase of the annual cycle is removed, there are some significant similarities among them. These phase changes represent the movement of the land ITCZ of the tropical Pacific into northern (southern) latitudes in boreal summer (winter) seasons.

The annual modes of precipitation simulated by S1 (Figure 12a) and S2 (Figure 12b) show some very interesting characteristics. First, the annual precipitation mode over Amazonia has a “month of April” phase in the tropical regions and a “month of January” phase southward of the equator (Figure 11a). Simulation S1, with the annual cycle of SST, shows only the month-of-April mode in the tropical region (Figure 12a), whereas simulation S2 with the annual cycle of solar forcing shows only the month-of-January/February mode mostly in regions southward of the equator (Figure 12b). The sum of these two phases (Figure 11b) produces patterns and strengths similar to that of C (Figure 11a). Thus C represents a scenario that exhibits the combined effects of both modes. These modes explain up to 70% of the annual precipitation variability, particularly in the Tropics. The annual precipitation modes (Figure 11a) in the northern Pacific regions are also reproduced by the sum of the S1 and S2 modes where S1 has a month-of-September phase and S2 has a month-of-July phase. This shows that solar forcing in the summer season contributes to a strong phase of the annual cycle of precipitation in the northeastern Pacific. Both over India and Australia, the monsoon simulated in C is much stronger than the monsoon simulated in either S1 or S2 or even the sum of S1 and S2. This again suggests that strong monsoon circulation and precipitation are caused by the combined influence of the solar and SST forcings in the respective monsoon season. In other words, boreal summer SST and solar insolation feedback interaction strengthen the monsoon even more than they each can do individually. However, as was pointed out earlier, weaker monsoon precipitation over India is accompanied by a stronger monsoon over tropical Africa. Regardless, the annual cycle of solar forcing has a stronger influence as compared to the annual cycle of the SST.

4.6 Tropical intraseasonal oscillation

One of the crucial features affecting tropical circulation and precipitation, even the onset of Asian monsoons, is the 30–60-day mode. It is often called the tropical intraseasonal oscillation (TIO). We had simulated them well in the earlier version of our GCM, called the GLA GCM (Slingo et al., 1996). Indeed, the current GEOS GCM with McRAS also simulates them reasonably well although they are somewhat weaker, presumably as a result of relaxation procedures of RAS that reduce precipitation intensities. A common feature of all the panels of Figure 13
is that the TIOs are equally well simulated in all the cases, that is, with or without the annual cycle of the solar insolation or SST. This result suggests that TIOs are neither SST nor solar irradiance annual cycle driven. In fact even huge changes in global-scale circulation in these simulations (C, S1, S2, and S3) do not affect them much. Consequently, they must be internal modes of the atmospheric dynamics and must be related to the interaction timescales associated with physical
processes. This would also explain large variations in TIOs from one model to the other. Even those models that have TIOs have been known to easily lose them as a result of new upgrades in the parameterizations, presumably because the adjustment timescales are disposable parameters often chosen by the scientist arbitrarily (Lee et al., 2001). However, this does not rule out the interaction of the surface fluxes with the dynamics of the boundary layer circulation and other surface–atmosphere interactions. Nevertheless, TIOs have a natural vacillation and are known to be stronger in winter—a scenario identifiable in C and easily discernible in the simulation S1 with only the annual cycle of SST, but TIO differences are unremarkable in S2 and S3 in which SST annual cycle was eliminated. This suggests that stronger TIOs in winter may be forced by seasonal changes in SST as opposed to the influence of solar insolation changes that affect the land regions of our model. A more detailed investigation of these issues is beyond the purview of the current study, but could be the subject of a future investigation.

4.7 Diabatic heating and moisture transports

The column mean moisture transports and convergences, which are inferred from precipitation minus evaporation over land are much less alike for C and S1 (cf.
Figure 14. Simulated JJA atmospheric moisture convergence averaged over the 4 yr for (a) C, (b) S1, and (c) S2. Contours are drawn for 0 and ± 2, 4, and 8 mm day$^{-1}$. Colors distinguish the patterns.

Figure 14a and Figure 14b) as opposed to C and S2 (cf. Figure 14a and Figure 14c). We note that the Indian monsoon, tropical African circulation, and South American wind fields markedly resemble each other in Figure 14a and Figure 14c. Although the SPCZ and southern high latitudes are very similar in all the
simulations, the climatologies of the northern Atlantic and Pacific are better simulated in the annual mean SST simulation, S2, but with the solar annual cycle. These simulations again show that the solar heating, with its strong influence in the land regions during the summer seasons, is primarily responsible for the peculiar structure of the global patterns of precipitation and diabatic heating. As pointed out earlier, both SSTs and local summer solar heating have a role in sustaining the Indian, all Asian, as well as Australian monsoons. Nevertheless, a comparison of the three panels of Figure 11 clearly shows that the solar heating of landmasses and its dependent structures of circulation are the dominant driver of the Indian and Australian monsoons.

5. Discussion and conclusions

A climate version of the GEOS GCM that contains important new upgrades to the GCM’s prognostic clouds, cloud microphysics, and snow and ice hydrology was used to isolate the relative influences of the annual cycles of solar irradiation and SSTs on the simulated atmospheric circulation and precipitation. The scientific motivation for comparing three companion simulations, one with the annual mean SST and one with annual mean solar flux at the top of the atmosphere, plus one with both the annual mean SST and solar irradiation, with the control simulation and DAO DAS data was to isolate how these forcings affect the tropical precipitation and circulation. Four 4-yr-long integrations, C, S1, S2, and S3, were generated. These simulations have shown how the land and ocean interact with each other to produce the observed circulation and rainfall climatology. Our second objective was to understand the influence of each annual cycle, the SSTs, and solar irradiation on circulation and rainfall separately. These objectives have been adequately accomplished within the intrinsic limitations of the GCM used for the study. However, because the GEOS GCM is a remarkably credible state-of-the-art climate model, we believe that model’s limitations do not substantially impair the current findings. The 4-yr control run, C, and its comparison with the corresponding 4-yr companion simulations, S1, S2, and S3, reveal the following.

The northward excursion of the monsoon into the Indian subcontinent in the boreal summer is modulated by the SST annual cycle, but it is influenced more by the annual cycle of solar insolation, particularly for India and South America. Our simulations show that the Indian summer monsoon is relatively stronger (weaker) in S2 (S1) invoking only the annual cycle of solar irradiance (SSTs) while using the annual mean SST (incoming solar radiation). Without the solar heating in the local summer season, neither the Indian nor the Australian monsoons develop. On the other hand, with only the solar insolation annual cycle and without the annual cycle of the SSTs, both monsoons emerge better than in S1, but both remain much weaker as compared to C.

The annual mode of precipitation over Amazonia shows (i) a month-of-April phase in the tropical latitudes in the SST annual cycle simulation and (ii) a month-of-January phase in the midlatitude (poleward of the Tropics) for the solar annual cycle simulation. It is interesting to note that C, which has both annual cycles, exhibits a combination of both annual modes of precipitation. We infer these
phases of the annual precipitation over Amazonia are jointly forced by the annual cycles of SST and solar radiation working in concert.

The boreal summer precipitation in Sahelian Africa is significantly reduced lacking increased solar heating of the Northern Hemispheric landmasses. This holds equally true for all monsoonal regions: India, Southeast Asia, as well as Australia. The only exception is the continental United States, where the two annual cycles, SST and/or solar forcing, are found to produce relatively marginal effects. There can be many reasons for this outcome. For example, over the continental United States, the global interactions can feedback in a complex way to produce an insignificant net effect on the circulation and precipitation. This issue has not been researched adequately at this time; therefore, we refrain from providing any explanation beyond describing the result.

The simulated circulation patterns over the Kuroshio Current region of the North Pacific off the east coast of Asia are affected far more in S2 as compared to S1 or S3. Since the Gulf Stream and its annual cycle are directly involved, we infer that these circulation and precipitation patterns are strongly linked to the SST; naturally for these regions, the annual cycle of the SSTs is important.

The 30–60-day oscillations (also called tropical intraseasonal oscillations or TIOs) are simulated equally well in each of the four integrations. These results suggest that TIOs are independent of the annual cycles of solar insolation and/or SSTs. Hence, they must be produced largely by the internal dynamics of the atmosphere. However, they are weakly linked to the annual cycles of heating or cooling as manifested in the annual cycles of land temperatures or SSTs. This conclusion is also borne out by the behavior and persistence of TIOs, which remain robust throughout the year despite being somewhat stronger during boreal winters.

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References


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