Evaluating Low-Cloud Simulation from an Upgraded Multiscale Modeling Framework Model. Part III: Tropical and Subtropical Cloud Transitions over the Northern Pacific

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ABSTRACT
An analysis of simulated cloud regime transitions along a transect from the subtropical California coast to the tropics for the northern summer season (June–August) is presented in this study. The Community Atmosphere Model, version 5 (CAM5), superparameterized CAM (SPCAM), and an upgraded SPCAM with intermediately prognostic higher-order closure (SPCAM-IPHOC) are used to perform global simulations by imposing climatological sea surface temperature and sea ice distributions. The seasonal-mean properties are compared with recent observations of clouds, radiation, and precipitation and with multimodel intercomparison results. There are qualitative agreements in the characteristics of cloud regimes along the transect among the three models. CAM5 simulates precipitation and shortwave radiative fluxes well but the stratocumulus-to-cumulus transition occurs too close to the coast of California. SPCAM-IPHOC simulates longwave radiative fluxes and precipitable water well, but with systematic biases in shortwave radiative fluxes. The broad, stronger ascending band in SPCAM is related to the large biases in the convective region but the characteristics of the stratocumulus region are still more realistic and the transition occurs slightly farther away from the coast than in CAM5. Even though SPCAM-IPHOC produces the most realistic seasonal-mean transition, it underestimates the mean gradient in low-cloud cover (LCC) across the mean transition location because of an overestimate of LCC in the transition and convective regions that shifts the transition locations farther from the coast. Analysis of two decoupling measures shows consistency in the mean location and the histogram of decoupling locations with those of LCC transition. CAM5, however, lacks such a consistency, suggesting a need for further refinement of its boundary layer cloud parameterization.

1. Introduction
Observations show that transitions between low-cloud regimes in the subtropics occur frequently and abruptly (e.g., Bretherton et al. 1995; Wood and Bretherton 2004; Wood et al. 2011). Global weather and climate models are notoriously inadequate in reproducing the temporally averaged transitions of cloud regimes, for example, along a transect in the northern Pacific (Siebesma et al. 2004; Teixeira et al. 2011, hereafter TJ11) from the Global Energy and Water Cycle Experiment (GEWEX) Cloud System Study (GCSS)/Working Group on Numerical Experimentation (WGNE) Pacific cross-section intercomparison (GPCI). This transect covers the transition from a stratocumulus-topped marine boundary layer (MBL) off the coast of California to a trade cumulus–topped, less well-defined MBL, and finally to the deep-convection regions in the intertropical convergence zone (ITCZ). TJ11 confirmed many of the typical global model problems in the representation of clouds: 1) underestimation of cloud amount and liquid water path (LWP) in the stratocumulus regime by most models with the corresponding positive biases in the shortwave (SW) radiative flux at the top of the atmosphere (TOA) and surface; 2) overestimation of clouds in the trade-wind region and in the deep tropics with the corresponding negative biases in the outgoing longwave radiation (OLR); 3) large spread between the different models in terms of cloud cover, LWP, and SW flux; and 4) significant differences between the models in terms of vertical cross sections of cloud properties, vertical velocity, and
relative humidity. Comparing the cloud cover mean statistics obtained while taking into account sharp gradients in cloud cover along the transect using the 3-hourly model output, TJ11 found that the negative cloud bias in the stratocumulus regimes is associated not only with lower values of cloud cover in these regimes but also with a stratocumulus-to-cumulus transition that occurs too early along the trade-wind trajectory. These results suggest that the understanding of cloud regime transition has not been well transferred to the development and improvement of cloud parameterizations in global models.

The Atlantic Stratocumulus Transition Experiment (ASTEX; Albrecht et al. 1995) and subsequent modeling work provided the basic understanding of the stratocumulus-to-cumulus transition (e.g., Bretherton and Pincus 1995; Bretherton et al. 1995; de Roode and Duynkerke 1997; Bretherton and Wyant 1997; Wyant et al. 1997; Bretherton et al. 1999). ASTEX depicted a transition of a stratocumulus cloud deck, predominantly driven by longwave radiative cooling at the cloud top, to thin and broken stratocumulus clouds penetrated by cumulus from below, driven mainly by convection due to an unstable near-surface stratification. De Roode and Duynkerke (1997) suggested that enhanced entrainment of dry air above inversion is a key mechanism in the stratocumulus-to-cumulus transition. Wyant et al. (1997) proposed a two-stage model for the boundary layer evolution based on detailed analyses of numerical simulations. In stage 1, decoupling is induced by increased latent heat fluxes in the deepening boundary layer and the cloudiness regime changes from a single stratocumulus layer to sporadic cumulus that detrains into stratocumulus clouds. In stage 2, a further sea surface temperature (SST) increase causes the cumuli to become deeper and more vigorous, penetrating farther into the inversion and entraining more and drier air above inversion. This mechanism is commonly shortened as “deepening–warming.” Bretherton and Wyant (1997) showed that decoupling is mainly driven by an increasing ratio of the surface latent heat flux to the net radiative cooling in the cloud, and derived a decoupling criterion based on this ratio. Xiao et al. (2011) confirmed that a stratocumulus-topped MBL is most likely to transit to a cumulus-topped one when there exists high probability of buoyancy reversal at the MBL top and the MBL is decoupled because of large surface latent heat fluxes.

Studies of other subtropical oceanic MBLs also revealed transitions similar to those observed during ASTEX, but the controlling processes were better understood (Garreaud et al. 2001; Sandu et al. 2010; Stevens et al. 2005; Wood et al. 2011). Sandu et al. (2010) used satellite observations and meteorological reanalysis to systematically examine the transition from unbroken sheets of stratocumulus to fields of scattered cumulus and the processes controlling them in four subtropical oceans. They found that both the transition and the processes driving the transition are quite similar among the subtropical oceans. The increase in SST and the associated decrease in lower tropospheric stability (LTS; Klein and Hartmann 1993) appear to play a far more important role in cloud evolution than other factors including changes in large-scale divergence and upper tropospheric humidity. Wood et al. (2011) documented a remarkably sharp spatial transition in MBL, cloud, and aerosol structures across the boundary between a well-mixed MBL containing overcast closed mesoscale cellular stratocumulus and a pocket of open cells (Stevens et al. 2005) with significantly lower cloud cover.

The complicated processes related to cloud regime transitions are the reason why global weather and climate models produce large spreads in models’ cloud, precipitation, and radiation characteristics (see TJ11). If cloud parameterizations are improved for all cloud regimes and tested well in the single-column models (Randall et al. 1996), it may not guarantee that the cloud regime transition characteristics are continuous due to the selection of distinct cloud regimes (Lock et al. 2000; TJ11). How well a model simulates the transition between different cloud regimes therefore provides a useful diagnostic of the success of representations of cloud processes. This study will present a detailed analysis of the transitions of stratocumulus, shallow cumulus, and deep precipitation, and radiation characteristics (see TJ11). If transitions are the reason why global weather and climate models produce large spreads in models’ cloud, precipitation, and radiation characteristics (see TJ11), then the latter two have been updated with the finite-volume dynamical core in the present study that is also used in CAM5. An MMF uses a two-dimensional cloud-resolving model (CRM) at each atmospheric column of the general circulation model (GCM) in place of the cloud parameterizations used in a conventional GCM (Grabowski 2001; Khairoutdinov and Randall 2001). But this approach shifts parameterizations to smaller scales as microphysics, radiation, and turbulence parameterizations are still needed. These three models offer diverse representations of cloud processes, particularly for low clouds. CAM5 has one of the most advanced cloud parameterizations among the conventional GCMs, with recent upgrades on moist turbulence and shallow cumulus parameterizations (Park and Bretherton 2009). The upgraded MMF includes an advanced third-order turbulence closure in its CRM component instead of a low-order turbulence closure in the
CRM embedded in SPCAM to better simulate the boundary layer and in-cloud turbulence.

The present study is Part III of three-part series with an overarching goal of evaluating the low-cloud simulation from the upgraded MMF called SPCAM with intermediately prognostic higher-order closure (SPCAM-IPHOC; see section 2 for details). Xu and Cheng (2013a, hereafter Part I) showed that this model can produce a global- and annual-mean low-cloud amount that is within 5.3% of observations from the merged CloudSat (Stephens et al. 2002), Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO; Winker et al. 2010), Clouds and the Earth’s Radiant Energy System (CERES; Wielicki et al. 1996), and Moderate-Resolution Imaging Spectroradiometer (MODIS; King et al. 1992) data product (C3M; Kato et al. 2011). The spatial distributions of low clouds are realistic in several ocean basins, with the relationships of low clouds with large-scale variables agreeing with those observed in these ocean basins (Part I). The global mean low-cloud amount is 3.7% higher than SPCAM with identical horizontal and vertical resolutions. Xu and Cheng (2013b, hereafter Part II) showed that the seasonal variations of these clouds are comparable to and in some instances better than those produced by the best regional high-resolution climate models (Part II; Wyant et al. 2010; Wang et al. 2011). Part II also discussed the seasonal variations in terms of the factors determining the seasonal variations of inversion strength rather than the deepening-warming mechanism.

The present study has two specific objectives: to evaluate the CAM5 and MMF simulations in the northeastern Pacific with state-of-the-art satellite observations and to understand the similarities and differences in the cloud regime transitions simulated from these models. In particular, the comparison between the two SPCAM versions can isolate the role of the advanced third-order turbulence closure in the MMF simulation of cloud regime transitions. Seasonal-mean properties and statistics of decoupling and transition characteristics are analyzed to achieve these objectives.

The rest of the paper is organized as follows. Section 2 will provide a brief description of models and experiment design. Results will be presented in section 3. Section 4 will give a summary and conclusions of this study.

### 2. Description of models and experiments

The advanced third-order turbulence closure used in the CRM component of the upgraded MMF is called intermediately prognostic higher-order closure (IPHOC) and has been documented in Cheng and Xu (2006, 2008, 2011). A detailed description of SPCAM-IPHOC with the IPHOC implemented in the Community Atmosphere Model, version 3.5 (CAM3.5; Collins et al. 2006), with a semi-Lagrangian dynamical core can be found in Cheng and Xu (2011). However, the current study uses a finite-volume dynamical core in CAM3.5 (Collins et al. 2006) for the SPCAM and SPCAM-IPHOC simulations. The finite-volume dynamical core is retained in CAM5. The host GCM used in Khairoutdinov and Randall (2001) was CAM, version 3.0 (CAM3.0), with the semi-Lagrangian dynamical core. This MMF is labeled as SPCAM v3.0 in this study. Although multiple physical parameterizations evolved in the transition from CAM3.0 to CAM3.5, most of them did not have a significant impact on the transition from SPCAM v3.0 to SPCAM v3.5 because all cloud parameterizations used in both versions of CAM were replaced by 2D CRMs. The embedded 2D CRM in SPCAM-IPHOC is the System for Atmospheric Modelling (SAM; Khairoutdinov and Randall 2003) with IPHOC, helping it to better simulate boundary layer turbulence and low clouds (Cheng et al. 2004; Cheng and Xu 2006, 2008), in place of the low-order turbulence closure in the standard SAM (and SPCAM). The physical processes such as convection and stratiform cloudiness, usually parameterized in a conventional GCM, are resolved crudely on the CRM fine grids, with 32 CRM grid columns and 4-km horizontal grid spacing. Cloud microphysics and radiation are parameterized at the CRM scale. Tendencies of heat and moisture from the CRM scale are communicated to the large scale via the GCM. On the other hand, the dynamical core computes the large-scale advective tendencies that are imposed to the CRMs. The surface turbulence fluxes over land grid points are calculated from a land surface model in CAM3.5, which are provided to the CRMs. The fluxes over ocean grid points are calculated on the GCM grids from the prescribed SSTs.

Details of CAM5 are documented in Neale et al. (2010). A brief summary of cloud parameterizations used in CAM5 is provided below. In CAM5, deep convection is parameterized using the mass flux scheme of Zhang and McFarlane (1995) with modifications (Neale et al. 2008; Richter and Rasch 2008). A moist turbulence scheme explicitly simulates cloud–radiation–turbulence interactions (Bretherton and Park 2009). Grid-scale condensation–evaporation follows a conserved-variable version of Zhang et al. (2003). Shallow convection is parameterized using a realistic plume dilution equation and closure (Park and Bretherton 2009). Stratiform microphysical processes are represented by a prognostic, two-moment formulation for cloud droplet and cloud ice mass mixing ratio and

1 The v3.5 notation after SPCAM and SPCAM-IPHOC is omitted for the rest of the paper.
number concentrations (Morrison and Gettelman 2008). Stratiform cloud fraction is determined from the assumed triangular distribution of total (including condensate) relative humidity while cumulus cloud fraction is calculated from the cumulus mass flux and cumulus vertical velocity (Xu and Krueger 1991; Neale et al. 2010).

The design of the IP-12L experiment performed with the SPCAM-IPHOC model with 12 layers below 700 hPa was detailed in Part I. That of the CAM5 and SPCAM experiments is similar. Note that the experiment names are identical to the model names except that IP-12L stands for SPCAM-IPHOC. Briefly, the CAM5, SPCAM, and SPCAM-IPHOC models were forced by specifying climatological (140-yr average) SST and sea ice with monthly mean annual cycles while coupled with the Community Land Model over land grid points (Oleson et al. 2004). All three models have a horizontal grid size of 1.9° × 2.5°. In both the SPCAM and IP-12L experiments, there are 32 layers in the vertical with 12 layers below 700 hPa to better resolve the low clouds. CAM5 has 30 layers with its standard vertical configuration, with 10 layers below 700 hPa. The embedded CRMs have the same vertical levels as the host GCM. Experiments IP-12L and CAM5 were each integrated for 10 yr and 3 months. But the results of the first 2 yr after the initial 3-month spinup are analyzed in this study in order to match with the SPCAM sensitivity experiment. All sensitivity experiments were run for only 27 months in this series of study (Part I). Results from the 10-yr-and-3-month integration of IP-12L were presented in Parts I and II for understanding the global climatology and seasonal variations of the eastern Pacific region.

The June–August (JJA) season is chosen to be the focus of discussion, as in TJ11, because it is the peak season of stratocumulus occurrence in the northeastern (NE) Pacific basin. Seasonal means of cloud, precipitation, and radiation output are analyzed to characterize the mean properties along the NE Pacific transect. The analysis of transition statistics was performed with hourly output from the last year of Experiments IP-12L and CAM5. For the other 9 yr and 3 months of these two experiments, only the daily- and monthly-mean model output was saved. The lower-frequency output also provides valuable information on the mean cloud regime transition, but not the instantaneous transition statistics.

3. Results

The results from the present analysis are organized into four parts: 1) the JJA seasonal mean properties of cloud, precipitation, and radiation; 2) the horizontal-vertical distributions of dynamic, thermodynamic, and cloud properties; 3) the cloud regime transition characteristics in JJA; and 4) the cloud regime transition characteristics in the other three seasons. In the discussion presented below, the convective region is loosely referred to as that from 2°S to 12°N, the stratocumulus region as that from 25° to 35°N and the transition region as that from 12° to 25°N. Note that the corresponding longitudes can be found from Fig. 1. As results indicate later, the boundary between stratocumulus and transition regions differs among the models.

a. Seasonal mean properties of cloud, precipitation, and radiation

Figure 1 shows the annual-mean low-cloud amounts [low-cloud cover (LCC)] in the NE Pacific region...
bounded by 10°S–40°N and 60°W–160°E. The observations are obtained from the 4-yr (2006–10) merged C3M data product (Kato et al. 2011). The state-of-the-art C3M product, a fully integrated cloud, aerosol, and radiation data product that merges CloudSat, CALIPSO, CERES, and MODIS data, provides vertical profiles of cloud fraction and cloud water content and cloud cover, LWP, and radiative fluxes, among others. The finer-spaced observations from CALIPSO (333 m), CloudSat (1.4 km), and MODIS (1 km) on the Aqua satellite are projected to the CERES/Aqua footprint with an average size of approximately 20 km × 20 km. Further details of the data processing for use in data–model comparison can be found in Part II.

The low-level clouds are referred to the clouds between the surface and 700 hPa in both models and observations. Since SPCAM-IPHOC provides cloud fraction in each CRM grid point from its IPHOC’s subgrid-scale distribution, a maximum overlap assumption is used to calculate LCC for each CRM column. The LCC for each GCM grid box is the average of the LCC of the embedded CRM columns. The same maximum overlap assumption is used to calculate the LCC from the C3M data. In SPCAM, cloud fraction on a CRM grid box is either 0 or 1, which is diagnosed from the sum of liquid and ice water path with a threshold of 10⁻³ kg m⁻². In CAM5, cloud fraction is obtained from its cloud parameterization schemes based upon the distribution of total relative humidity and other considerations (Neale et al. 2010).

As shown in Fig. 1, the spatial distributions of the annual-mean LCC from SPCAM and IP-12L resemble that of the C3M observations except for the underestimated magnitudes in the stratocumulus region of SPCAM. The LCC along the transect ranges from 60% in SPCAM and 70% in IP-12L near the coast to less than 40% near the equator. This range of LCC values is less than that of C3M observations (80%–40%). For CAM5, the low-cloud maximum near the coast is well simulated, but the cloudiness decreases significantly faster than in C3M observations as one moves farther away from the coast, suggesting that the mean stratocumulus-to-cumulus transition occurs too early along the trade-wind trajectory in CAM5.

As stated in section 2, the three experiments presented in this study are intended to simulate the long-term climatology driven by climatological SSTs. The SSTs increase southwestward almost linearly from cold water off the coast of California (289 K) to 301 K in the ITCZ region. The differences between the climatological SST and the SSTs over a particular year or several years (Fig. 2) are not negligible. For example, the JJA 1998 SSTs are colder (up to 0.7 K) over the stratocumulus region (except near the coast) and generally warmer (0.5 K) over the ITCZ. The climatological and the JJA 2006–10 (the C3M data period) SSTs look very similar in the subtropics, but different near the equator. With the SST differences shown, we could expect there to be some impacts on the magnitudes and locations of simulated low clouds and the intensity of deep convection. When compared with the simulated results presented in TJ11 later in this study, which simulated the JJA 1998 condition, the differences between the climatological SST and the JJA 1998 SST should also be kept in mind. This is also the reason why we show the standard deviations from the 10-yr mean of experiment IP-12L to measure the model’s internal variability and to distinguish robust model biases from internal noises.

Figures 3 and 4 show a few surface, TOA, and integrated physical properties along the transect, as well as the standard deviations from IP-12L. These include column-integrated water vapor (precipitable water), total cloud cover (TCC), liquid water (plus ice) path, surface precipitation rate, OLR, TOA shortwave absorbed (net) flux, and surface net longwave (LW) and net SW radiative fluxes. Note that a maximum-random overlap assumption is used to calculate the TCC in CAM5 only. But in both MMFs TCC is first calculated over a CRM grid column with the maximum overlap and then averaged to the GCM grid column. The observational datasets used in the comparison include LWP and precipitable water data from the Special Sensor Microwave Imager (SSM/I) because of the high quality and length of the data record. The monthly gridded (1° × 1°) version-7 SSM/I data were downloaded from

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**Fig. 2.** Comparison of the seasonal-mean (June–August) climatological sea surface temperature (solid), 2006–10 JJA SST (long dashes), and 1998 JJA SST (short dashes) along the Pacific transect shown in Fig. 1.
Remote Sensing System (http://www.remss.com). Data from July 1987 to December 2010 are averaged over the JJA season. The monthly precipitation data from Global Precipitation Climatology Project (GPCP) are used, which covers the period of 1979–2011 (Adler et al. 2003). The CERES Energy Balanced and Filled (EBAF) data (Loeb et al. 2009) provide the highly constrained radiative energy fluxes at TOA and surface. The diurnal cycle is accounted for by using data from Terra, Aqua, and five geostationary satellites. And the net TOA irradiance is adjusted with ocean heat storage data. The CERES data span the period between March 2000 and February 2010. The surface EBAF fluxes are calculated from a radiative transfer model but are constrained by the observed TOA CERES fluxes.

All three models simulate the variation of precipitable water along the transect rather well, with minima of ~20 kg m\(^{-2}\) off the coast of California to maxima over 50 kg m\(^{-2}\) over the ITCZ at 8°N (Fig. 3a), following the variation of SST (Fig. 2). Although precipitable water is an integral parameter, its horizontal variation illustrates well the major changes that occur in the atmospheric column. All three models overestimate precipitable water by up to 5 kg m\(^{-2}\) except near the equator of IP-12L and near the coast of CAM5, with the smallest overestimate by IP-12L (1–3 kg m\(^{-2}\)). The amounts of overestimate for CAM5 and SPCAM are comparable to the standard deviations of the multimodel ensemble shown in TJ11 but greater than the IP-12L model’s internal variability. Those of IP-12L are smaller than the standard deviation. SPCAM and CAM5 have nearly identical precipitable water values along the transect except for the steeper decrease near the coast in CAM5. This similarity may be coincidental since other properties to be discussed later are not similar.
Despite the large difference in precipitable water between CAM5 and SSM/I observation, surface precipitation rate over the convective region of CAM5 agrees best with the GPCP observations with a peak value 10 mm day\(^{-1}\) at 8°N (Fig. 3d). All three models produce little precipitation in the stratocumulus and initial transition-to-cumulus regions (north of 18°N). The peak value of SPCAM precipitation (19 mm day\(^{-1}\), which is 4–8 mm day\(^{-1}\) greater than in SPCAM v3.0 (Khairoutdinov et al. 2005) due to the impact of dynamical cores in the host GCMs on precipitation patterns (Part I), is nearly twice as large as that of GPCP. The difference between CAM5 and GPCP does not exceed the standard deviation of the multimodel ensemble in TJ11 (2–3 mm day\(^{-1}\)), but that between the two versions of the MMF and GPCP does and exceeds the IP-12L model’s internal variability. The larger differences are tied to the larger magnitudes of the MMFs’ large-scale ascent shown later.

As discussed in the introduction, typical weather and climate model problems in the representation of clouds include 1) underestimation of cloud amount and LWP in the stratocumulus regime with the corresponding positive biases in SW radiative flux at TOA and surface; 2) overestimation of clouds in the transition region and in the deep tropics with the corresponding negative biases in OLR; and 3) large spreads between the different models in terms of cloud cover, LWP, and radiative fluxes. A question to be raised is whether or not the first two problems linger in the MMF simulations. The third problem, shown later, is anticipated (Figs. 3 and 4) because of the diverse representations of clouds in the three models.

The first problem remains in CAM5 for the region between 18° and 29°N, but not north of 29°N because both TCC and LWP are adequately simulated there (Figs. 3b,c and 4b,d). SPCAM underestimates TCC in the stratocumulus region (20°–36°N), but overestimates LWP there so that the SW radiative flux biases are negative (i.e., smaller downward fluxes) instead of positive. The negative SW flux biases are slightly larger for the segment...
of 25°–36°N of IP-12L because of larger LWP overestimates (than SPCAM) and no underestimate in TCC. The second problem appears to be serious for IP-12L. There are 5%–20% overestimates in TCC over the transition and convective regions (Fig. 3b), which are (as shown later) related to overestimates in LCC, not those in the upper-level clouds. This explains the relatively small underestimate in OLR resulting from the large overestimate in TCC. The TCC is also overestimated in the transition region of SPCAM, but underestimated in CAM5. The large underestimates in OLR appear in the convective region of SPCAM and CAM5 instead of IP-12L (Fig. 4a). That is, OLRs in CAM5 and SPCAM are about 20–40 W m$^{-2}$ smaller (as compared to CERES EBAF) than in IP-12L because of excessive cirrus clouds (shown later). Additionally, SPCAM has large underestimates (i.e., exceeding the IP-12L model’s internal variability) in this region for TOA SW flux and surface SW and LW fluxes, as compared to CERES EBAF (Fig. 4). This result is related to large overestimates of LWP (Fig. 3c) because it has the strongest Hadley cell among the three models, evidence of which will be shown later.

Despite the spreads among the models, IP-12L simulates the variations of all four fluxes along the transect well except for the net surface LW flux in the transition region (Fig. 4), suggesting that cloud regime transitions are more realistic in this model when compared with the other two models. It is, however, noted that there are systematic biases of 20–40 W m$^{-2}$ in SW and 5–10 W m$^{-2}$ in LW for IP-12L; that is, the biases in the total (SW + LW) flux are still significant. These biases can be reduced with a consistent treatment of radiative effects of sub-CRM grid-scale variability of clouds (Pincus et al. 2005). In CAM5, on the other hand, misplacement of cloud regime transitions can be seen from the separations of 4°–8° in the peaks of SW and LW fluxes from those of the CERES observations. But the overall biases in these fluxes in CAM5 are smaller than those in IP-12L except for OLR.

**Fig. 5.** Vertical-horizontal (latitude) cross sections of pressure vertical velocity (Pa s$^{-1}$) from (a) CAM5, (b) SPCAM, and (c) IP-12L simulations, and from (d) ERA-40 for the GPCI transect.
b. Cross sections of dynamic, thermodynamic, and cloud properties

Differences in model physics, whether it is cloud microphysics, turbulence closure, or radiative transfer in MMFs or cloud parameterizations in CAM5, will inevitably alter the simulated large-scale dynamics and thermodynamics, in addition to cloud distributions. The differences in cloud distributions can be amplified or diminished by those in dynamics and thermodynamics. Examination of vertical cross sections of these physical measures will be helpful to understand the inherent relationships and the model biases discussed earlier.

Figure 5 shows vertical cross sections of large-scale pressure vertical velocity from CAM5, SPCAM, and IP-12L, compared with the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; Uppala et al. 2005). All models simulate features that resemble the Hadley circulation with dominant ascent over the convective region and subsidence in the free troposphere over the transition and stratocumulus regions. The convective region of ERA-40 exhibits a peak in the middle/upper troposphere (~300 hPa) and another peak in the lower troposphere (~800 hPa). The lower peak is not captured in CAM5, while it is overestimated in IP-12L by ~30%. SPCAM produces the widest and strongest ascent band throughout the entire troposphere, with overestimates being greater than 50%. This ascending band, stronger than that in CAM5, results from feedbacks with stronger convection and thus is related to model biases discussed in section 3a, such as excessive LWP, heavy precipitation, and low OLR. There is a good agreement in the intensity of subsidence in the stratocumulus region near the coast, but the subsidence in the transition region varies among the models. The subsidence in the middle/upper troposphere of SPCAM is weaker by a factor of >2, when compared with the other two models and ERA-40. This weaker subsidence cannot suppress the extensive anvil clouds, as shown below.

The relative humidity (RH) cross section is shown in Fig. 6, which is linked to both large-scale vertical motion and cloud formation. All three models have high RH (~90%) areas in the boundary layer throughout the transect, variable high values in the convective region, and extremely low RH values (<30%) in the free troposphere of the subsidence region. These features are in general agreement with ERA-40 (Fig. 6d) and the
Atmospheric Infrared Sounder (AIRS; Chahine et al. 2006) observations (Fig. 6e), particularly the spatial pattern of IP-12L. There is, however, a large disagreement in the upper troposphere, where CAM5 is nearly saturated in the convective region while both SPCAM and IP-12L exhibit large extents of moderate RHs (40%–60%). In addition, the gradual rise of the moist boundary layer from stratocumulus to deep convective regions is simulated in all models.

The large differences in RH distributions should impact the cloud distributions because high RH areas are generally associated with high cloud fraction areas (Fig. 7), a fact that has been used in many cloudiness parameterizations including that in CAM5 (e.g., Slingo 1980; Xu and Krueger 1991; Xu and Randall 1996; Neale et al. 2010). Over the boundary layer, the C3M observations show cloud fractions over 50% in the stratocumulus region, which gradually decreases to 25% in the convective region. This transition is qualitatively simulated in all three models. IP-12L produces the thickness of the cloud layer well, but cloud fractions are overestimated by ~10% (absolute amount) in the stratocumulus region and ~20% in the transition and convective regions. It is likely that the latter figure would be smaller if optically thin clouds located below optically thicker high clouds were detected by CALIPSO. Both CAM5 and SPCAM have thinner cloud layers over the boundary layer, as compared to the C3M observations. Their cloud fractions are also underestimated throughout the transect, especially in CAM5. This is a common problem in many conventional GCMs (see TJ11).

In the convective region, there is a good agreement in the location of upper tropospheric maximum cloudiness (200–250 hPa), but the cloud fraction is slightly underestimated by IP-12L (5%–10%; absolute amount) and overestimated by SPCAM (~10%) and greatly overestimated by CAM5 (~40%). The upper tropospheric clouds over the transition region of SPCAM and CAM5 are overestimated, too. These differences and those in upper tropospheric water vapor have large impacts, as...
discussed earlier, on the radiative energy budget in the atmosphere and at TOA (Fig. 4). Although CAM5 has large cloud fractions in the upper troposphere, the associated liquid (plus ice) water content (Fig. 8a) is very small. The net impact to radiative heating/cooling may not be as large as that by SPCAM because overestimated cloud fraction is accompanied by large liquid (plus ice) water content in SPCAM (Fig. 8b).

The simulated liquid water (plus ice) content shows large differences from C3M observations (Fig. 8). Both versions of the MMF, whose vertical cross sections are rather similar to each other, overestimate the liquid water in the boundary layers and liquid (plus ice) in the convective regions. The exact amount of overestimates cannot be qualified due to large uncertainties in the satellite retrievals. For example, cloud liquid water from CloudSat shown in Fig. 12 of TJ11 is higher than that shown in Fig. 8d by a factor of 2–4. An overestimate of cloud liquid water by a factor of 2 (relative to the C3M observations) in the stratocumulus regions of SPCAM-IPHOC was also noticed in Cheng and Xu (2011) and Part II.

The lack of liquid water content in the convective region of CAM5 is not surprising for two reasons. One is that convective parameterization does not output updraft–downdraft liquid water content except for convective detrainment, and the other is the inconsistency between cloud macro- and microphysical parameterizations that results in unrealistic depletion of liquid water. A solution is currently being developed to fix this problem (Caldwell et al. 2011). Explicit microphysics within a convective parameterization can also help (Song and Zhang 2011).

c. Mean transition characteristics

Seasonal mean properties are helpful to diagnose model’s mean transition characteristics. However, stratocumulus-to-cumulus transition occurs more rapidly in time and space (e.g., Wood et al. 2011). For example, the rapid reduction in low-cloud fraction off the coasts is often seen on satellite images. This abrupt transition gets smeared out when temporally averaged data are used. The sharper discontinuities can only be
captured in high-frequency cloud data. In this study, we use the hourly output available for the last year of the 10-yr-and-3-month CAM5 and IP-12L simulations.

The method proposed in Teixeira et al. (2011) is followed except using the LCC model output rather than the TCC model output. There are two reasons for this modification. One is that the presence of high-level clouds above a low-cloud layer may distort the transition location if TCC data are used and the other is that high-quality LCC data such as C3M have recently become available from observations but only TCC data were available in TJ11. The TJ11 method determines the first location along the transect where LCC drops by 30%, starting at the northernmost point in the stratocumulus region, and assumes uniform cloud cover to the northeast and southwest of the gradient’s location by taking the spatial averages of LCC for all the points to each side of the location of the sharp gradient. The result is shown as the piecewise curves in Fig. 9. Another piece of information shown in Fig. 9 is the histogram of locations of the sharp gradient. We also used a threshold of 20% and, as discussed later, found a larger sensitivity of the IP-12L results to the threshold than that of the CAM5 results.

With the 30% threshold, the mean LCC gradient location is located at 19°N for IP-12L, as compared to 29°N for CAM5. For IP-12L, the mean gradient location is about the same as that diagnosed from the International Satellite Cloud Climatology Project (ISCCP; Rossow and Schiffer 1999) data (Fig. 13 in TJ11) with the same threshold but using the TCC data (C3M data are not used because of the sparseness of data on the daily interval resulting from the narrow swaths of the satellites). It should be noted that the ISCCP TCC includes the contributions of middle and high clouds, in addition to LCC, especially the southwest side of the transect. The mean gradient strength across the gradient location is

![Fig. 9. LCC statistics for (a),(c) IP-12L and (b),(d) CAM5 simulations along the GPCI transect using a methodology based on the identification of sharp gradients of LCC (thick dashed curve; left axis) along the transect and histograms (right axis) of the locations of these strong gradients using LCC-jump thresholds of 30% in (a),(b) and 20% in (c),(d). Sharp gradients are identified nearly 100% of the time in (b)–(d), but only 91.7% in (a). The thick solid line in (a) is the same as the thick dashed line except using ISCCP TCC data. See text for the details describing the steps for determining the mean location of the sharp gradients and the corresponding mean LCC to the northeast and southwest sides.](image-url)
only 25% versus 40% for the ISCCP data. It seems that the underestimated gradient strength is due not only to underestimates in LCC on the stratocumulus side (73% versus 84% for ISCCP), but also to overestimates on the cumulus side of the transect (48% versus 44% for ISCCP). The former can be caused by very low LCCs near the coast, which reduces the average LCC to the NE side. It will become clear that the overestimates in LCC on the cumulus side (see Fig. 7) are an equally important factor.

The histogram for IP-12L (Fig. 9a)—in particular, the relatively large occurrence frequency of gradient locations in the convective region and the relatively small frequency in the stratocumulus region and transition regions—is not consistent with ISCCP observations (Fig. 13 in TJ11). The ISCCP distribution peaks at 23°N. The use of a 20% threshold greatly reduces the occurrence frequency in the convective region (Fig. 9c). The histogram with the smaller threshold resembles that of ISCCP observations (note that ISCCP TCC was used in TJ11). The mean gradient strength remains about the same (24.2%) as that with the 30% threshold. The average LCC on the cumulus side is higher (52.5%) while that on the stratocumulus side is also higher (76.7%). These results suggest that overestimates in the transition and convective regions are more likely responsible for the underestimated mean LCC gradient, compared to the ISCCP result mentioned earlier. That is, the simulated transitions in IP-12L are not as sharp as in the observations, which will be further examined later. Also noted is that the mean gradient location is moved from 19° to 24.7°N with the lower threshold (Fig. 9c).
As in TJ11, the mean gradient location and histogram remain the same (29°N) with the different thresholds for CAM5 (Figs. 9b,d). But the mean gradient strength increases from 30% to 42% as the threshold changes from 20% to 30%. With either threshold, the hourly LCC gradient locations occur as south as 20°N, but the majority occurs between 25° and 33°N. The histograms do not resemble that of ISCCP observations. Considering that the mean gradient location is at 20°N in ISCCP observations, the most frequent transitions simulated in CAM5 occur too close to the coast as in the seasonal mean transition discussed earlier. Additionally, the mean gradient location from SPCAM v3.0 (see TJ11) occurs also farther off the coast (at 25°N) than in CAM5.

The differences in the transition characteristics between IP-12L and CAM5 can also be revealed from the latitude versus LCC frequency diagram using the hourly model output (Fig. 10). Compared to the C3M (Fig. 10c), IP-12L seems to reproduce the essence of the observed LCC histograms although neither model reproduces all features on the observed histograms. The CAM5 has a bimodal structure, with shallow cumulus (5%–25%) occurring south of 25°N and stratocumulus (>80%) near the coast of California (north of 29°N). This result is in stark contrast with that of CAM3.0 shown in TJ11. The latter is similar to that of IP-12L shown in Fig. 10b. LCC was partly parameterized in CAM3.0 based on the dependence on the LTS (Klein and Hartmann 1993) with empirical coefficients characterizing the climatological condition. This formulation led to the fairly continuous transition with narrow ranges in LCC at all latitudes in the histogram, similar to the ISCCP and C3M observations.

For IP-12L, the simulated LCC variability is wider and the peak frequency is lower at all latitudes, compared with the C3M observations. One of the common features between IP-12L and C3M is that there is a minimum peak frequency at around 20°N. This result may
suggest that the LCC variability is the widest there along the transect. A noticeable difference is that the peak frequency appears at 80% LCC in C3M observations, instead of 100% LCC in both IP-12L and CAM5 north of 25°N. This is likely due to the high-resolution measurements of the satellites. The wide range of the simulated LCC variability may be the reason for the frequent occurrences of LCC jump in the transition and convective regions shown in Fig. 9 although the annual- and seasonal-mean LCCs agree with the C3M observations well (Fig. 1; Part II).

Next, the statistics of two decoupling measures are computed in order to understand the relationship between the LCC transition and boundary layer decoupling (Fig. 11). A subcloud-based decoupling measure (Jones et al. 2011) uses the difference between the cloud-base height \( z_{CB} \) and lifting condensation level (LCL) obtained from a parcel lifted from the surface. This decoupling measure is defined as
\[
D_{zb} = z_{CB} - z_{LCL}
\]
The details of computing \( D_{zb} \) are given in Part II. A profile-based decoupling measure compares the values of total water mixing ratio \( q_t \) near the surface to those at cloud-base height (Wyant et al. 1997). That is, \( \Delta q_t = (q_t)_{sfc} - (q_t)_{CB} \), where subscripts sfc and CB denote the surface and cloud base, respectively. Based on the observations of decoupling boundary layers, Jones et al. (2011) found that there is a linear relationship between these two decoupling measures. We choose \( \Delta z_b = 0.3 \text{ km} \) and \( \Delta q_t = 1.0 \text{ g kg}^{-1} \) as thresholds, based on the findings in Jones et al. (2011), to compute the statistics following the method in TJ11 for LCC jump. In this calculation, the location where the decoupling measure first exceeds the chosen threshold, starting from the coast, is chosen as the decoupling location. These thresholds are slightly lower than those of typically observed decoupling boundary layers, which can be justified by large spatial coverage of model grid cells, as compared to high-spatial-resolution measurements in Jones et al. (2011).

Histograms of the decoupling occurrences using the two decoupling measures (Fig. 11) resemble that of LCC gradient using the 20% threshold (Fig. 9c) except that \( D_{qt} \) has a narrower distribution around its mean latitude of 25.6°N. The mean decoupling latitude for \( \Delta z_b \) is about 2.5° farther away from the coast, as compared to that of the mean LCC gradients. The mean value of \( \Delta z_b \) is 0.15 km on the stratocumulus side but 0.5 km on the cumulus side, while the mean value of \( \Delta q_t \) is 0.7 g kg\(^{-1}\) on the stratocumulus side and 3.2 g kg\(^{-1}\) on the cumulus side. On the other hand, the decoupling measures for CAM5 (Figs. 11b,d and 12a) do not show results similar to those of IP-12L as LCC gradient does. This is due perhaps to the determination of cloud-base heights and physical formulation of shallow convection in CAM5. For instance, LCL is used to determine cloud base height, which results in \( \Delta z_b = 0 \). These results suggest that IP-12L produces reasonable decoupling statistics (see also Fig. 12b), which are consistent with those of stratocumulus-to-cumulus transition shown in Fig. 9 and LCC histogram shown in Fig. 10b. The quantitative aspect of this result is subject to the thresholds chosen for the analysis.
d. Transition characteristics in different seasons

We also analyze the transition characteristics in the other three seasons when the stratocumulus clouds are less active (Part II). The 20% LCC-jump threshold is used in this analysis. A consistent result between CAM5 and IP-12L is the significantly reduced LCC jumps across the mean transition latitudes for the other three seasons [September–November (SON), December–February (DJF), and March–May (MAM)] (Fig. 13). This is related to the reduced LCC on the stratocumulus side (55%–57% versus 78% in JJA for IP-12L; 46%–50% versus 68% in JJA for CAM5) rather than the increased LCC on the cumulus side (Table 1). Relatedly, the mean
transitions for these three seasons are located slightly closer (0.4°–2.2°) to the coastline than in JJA for both IP-12L and CAM5 (Table 2). The seasonal changes of $\Delta z_b$ and $\Delta q_t$ (Fig. 14) are also consistent with those of LCC gradients shown in Fig. 13 for IP-12L. The mean decoupling gradients across the mean latitudes are smaller than in JJA for both measures (Table 1). This is related to both the increases of decoupling strengths on the stratocumulus side and the decreases of decoupling strengths on the cumulus side.

4. Summary and conclusions

An analysis of simulated cloud regime transitions along a transect from the subtropical California coast to the tropics for the June–August (JJA) season has been presented in this study. CAM5, superparameterized CAM (SPCAM), and an upgraded SPCAM (SPCAM-IPHOC) are used to perform global simulations by imposing climatological SST and sea ice distributions. The seasonal-mean properties are compared with recent observations of clouds, radiation, and precipitation and multimodel intercomparison results. A main conclusion of this study is that there are qualitative agreements in the characteristics of cloud regimes along the transect among the three models, with SPCAM-IPHOC producing the most realistic cloud-regime transition. But there are some significant differences in the characteristics of stratocumulus-to-cumulus transition and the intensity of overturning circulations.

CAM5 simulates the magnitudes of precipitation and shortwave radiative fluxes well but the stratocumulus-to-cumulus transition occurs too close to the coast of California. SPCAM-IPHOC simulates longwave radiative fluxes and precipitable water well, but with systematic biases in shortwave radiative fluxes. The broad, stronger ascending band in SPCAM is related to the large biases in the convective region but the characteristics of the stratocumulus region are still more realistic and the transition occurs slightly more off the coast than in CAM5. Overestimates of upper tropospheric clouds in the convective and stratocumulus-to-cumulus transition regions and cloud–radiation interactions (related to the biases in cloud-top height) in SPCAM can explain the discrepancies in the radiative fluxes from the CERES observations. SPCAM-IPHOC simulates the variations of all four radiative fluxes along the transect well, suggesting that cloud regime transitions are represented more realistically in this model than the other two models.

The vertical structures of the transect show substantial differences in dynamics, thermodynamics, and cloud distributions among the three models. Obviously, these models simulate features that resemble the Hadley circulation with dominant ascent over the convective region and subsidence in the free troposphere over the transition and stratocumulus regions. One of the substantial differences is the widths and strengths of the convective region. For example, the double-peak vertical distribution of ascent in the convective region is not well captured by any of the three models. Consequently, there is a large disagreement in relative humidity of the upper troposphere in the convective and transition regions, where CAM5 is nearly saturated in the convective region (too cold in the upper troposphere) and both versions of the MMF have large areas of moderate RHs (40%–60%). This disagreement is also related to that of the cloud distributions among the three models, as compared to C3M observations. CAM5 produces very large cloud fractions in the upper troposphere but is accompanied

### Table 1. Seasonal variations of the mean decoupling measures $\Delta z_b$ and $\Delta q_t$, and LCC on the cumulus (Cu) and stratocumulus (Sc) sides of the mean LCC-gradient–decoupling latitude for CAM5 and IP-12L simulations.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Simulation</th>
<th>Cu</th>
<th>Sc</th>
<th>Cu</th>
<th>Sc</th>
<th>Cu</th>
<th>Sc</th>
<th>Cu</th>
<th>Sc</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\Delta z_b$ (km)</td>
<td>IP-12L</td>
<td>0.515</td>
<td>0.153</td>
<td>0.481</td>
<td>0.215</td>
<td>0.467</td>
<td>0.212</td>
<td>0.449</td>
<td>0.211</td>
</tr>
<tr>
<td>CAM5</td>
<td>0.0</td>
<td>0.0</td>
<td>2.99</td>
<td>0.826</td>
<td>2.924</td>
<td>0.888</td>
<td>2.837</td>
<td>0.759</td>
<td></td>
</tr>
<tr>
<td>$\Delta q_t$ (g kg$^{-1}$)</td>
<td>IP-12L</td>
<td>3.182</td>
<td>0.615</td>
<td>47.0</td>
<td>56.8</td>
<td>46.9</td>
<td>55.9</td>
<td>48.4</td>
<td>55.2</td>
</tr>
<tr>
<td>CAM5</td>
<td>0.545</td>
<td>0.452</td>
<td>25.9</td>
<td>50.0</td>
<td>26.9</td>
<td>47.4</td>
<td>29.1</td>
<td>45.9</td>
<td></td>
</tr>
<tr>
<td>LCC (%)</td>
<td>IP-12L</td>
<td>52.5</td>
<td>76.7</td>
<td>52.5</td>
<td>76.7</td>
<td>52.5</td>
<td>76.7</td>
<td>52.5</td>
<td>76.7</td>
</tr>
<tr>
<td>CAM5</td>
<td>32.7</td>
<td>67.5</td>
<td>32.7</td>
<td>67.5</td>
<td>32.7</td>
<td>67.5</td>
<td>32.7</td>
<td>67.5</td>
<td></td>
</tr>
</tbody>
</table>

Table 2. The mean LCC gradient (>20%) location (°N) for the four seasons of both IP-12L and CAM5 simulations. Also shown are the mean locations of stability parameters exceeding threshold values $\Delta z_b > 300$ m and $\Delta q_t > 1$ g kg$^{-1}$ for the four seasons of IP-12L simulation.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Parameter</th>
<th>JJA</th>
<th>SON</th>
<th>DJF</th>
<th>MAM</th>
</tr>
</thead>
<tbody>
<tr>
<td>IP-12L</td>
<td>LCC</td>
<td>24.7</td>
<td>26.7</td>
<td>25.1</td>
<td>26.1</td>
</tr>
<tr>
<td>$\Delta z_b$</td>
<td>22.2</td>
<td>26.4</td>
<td>25.3</td>
<td>24.8</td>
<td></td>
</tr>
<tr>
<td>$\Delta q_t$</td>
<td>25.6</td>
<td>29.7</td>
<td>28.7</td>
<td>28.1</td>
<td></td>
</tr>
<tr>
<td>CAM5</td>
<td>LCC</td>
<td>29.5</td>
<td>31.7</td>
<td>31.2</td>
<td>30.4</td>
</tr>
</tbody>
</table>
by small condensate amounts. The overestimates in cloud fraction and cloud-top heights in SPCAM are accompanied by large liquid water content. This combination can cause large biases in both LW and SW radiative fluxes, which lead to the strong large-scale ascent. These cloud biases were, however, smaller in SPCAM v3.0 (see TJ11). Other substantial differences among the three models examined in this study are related to the thickness of the boundary cloud layer and its water content despite the fact that the gradual rise of the moist boundary layer from the stratocumulus to deep convective regions is well simulated. SPCAM-IPHOC produces the thickest boundary layer but it overestimates cloud fraction and liquid water content (compared to C3M observations),

Fig. 14. As in Fig. 13, but for the two decoupling measures $\Delta z > 0.3$ km and $\Delta q > 1$ g kg$^{-1}$ for IP-12L only. Transition locations are identified from (b) 95.3% to (a) 99.6% of the time.
whereas CAM5 underestimates these quantities, especially in the transition and convective regions. SPCAM also underestimates cloud fraction, but its liquid water content is overestimated.

The high-frequency transition characteristics have been examined by identifying sharp gradients in LCC in hourly model output of CAM5 and SPCAM-IPHOC. The mean transition location is located at 19°N for SPCAM-IPHOC, as compared to 29°N for CAM5. For SPCAM-IPHOC, the mean sharp gradient location is the same as that diagnosed from ISCCP TCC data. The mean gradient strength across the transition location is, however, only 25% in LCC versus 40% for the ISCCP data. This is contributed by underestimates in LCC on the stratocumulus side and overestimates on the cumulus side of the transect. It is shown that there is relatively high occurrence frequency of sharp gradients in the convective region and the relatively small frequency in the stratocumulus and transition regions with the 30% LCC-jump threshold, which is not consistent with ISCCP TCC observations. The 20% LCC-jump threshold, as expected, greatly reduces the occurrence frequency in the convective region, but does not change the mean gradient strength. Analysis of two decoupling measures shows consistency in the mean locations and the histograms of decoupling locations with those of LCC sharp gradients. CAM5, however, lacks such a consistency, suggesting a need for further refinement of its boundary layer cloud parameterization.

The analyses performed in this study will be valuable for evaluating the performance of improved boundary layer cloud parameterizations in conventional GCMs and MMFs. The LCC transition characteristics, especially their statistics, provide a useful diagnostic of the success of representations of cloud processes in GCMs because dynamics, thermodynamics, and cloud distributions are tightly linked in such a sharp transition region over the northern Pacific. High-frequency variations of other cloud micro- and macrophysical properties such as cloud-top height and LWP for distinct boundary layer cloud types (e.g., Xu et al. 2008) will be helpful to further identify model deficiencies. Although the present version of SPCAM-IPHOC captures some major features of this transect, further improvement of its boundary layer cloud parameterization is needed. This may be achieved with improved cloud microphysics and its interactions with aerosols.

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