1. Introduction

Because of the smaller heat capacity of soil compared to water, the amplitudes of the diurnal cycle of surface total available turbulent (latent and sensible) heat flux and skin temperature tend to be greater over land than ocean. This likely amplifies lower-atmospheric heat energy in the afternoon, which often increases buoyant force, as measured by convective available potential energy (CAPE; Pielke 2001). As a result, continental precipitation is most frequently observed from noon to late afternoon (Wallace 1975; Carbone et al. 2002; Matsui et al. 2010; Kikuchi and Wang 2008) and has been traditionally used to explain why continental convection is more vigorous than maritime (Williams and Stanfill 2002, hereafter WS02).

Lucas et al. (1994) investigated aircraft-measured vertical velocity within deep convection over ocean and land and found that the convective updraft cores of maritime systems are only one-third or one-half the size of those of continental systems, although CAPE is similar between the two environments. WS02, Zipser et al. (2006), and Orville and Henderson (1986) similarly showed that continental convective precipitation
systems tend to have more vigorous radar reflectivities and much higher lightning flash rates per storm as observed by the Tropical Rainfall Measuring Mission (TRMM) satellite. These convective clouds with frequent lightning suggest enhanced electric charge separation associated with mixed-phase cloud microphysics processes, accretion of supercooled cloud droplets onto ice crystals, and collision between the crystals and graupel particles (Williams et al. 2002, 2005; WS02; Takahashi 1978).

WS02 pointed out that the width of a thermal plume could be associated with the boundary layer height based on the classic similarity theory of Morton et al. (1956). This simple parcel theory showed that a buoyant thermal parcel originating from point sources has a constant expansion (~10° expansion angle) toward the top of the boundary layer, leading to the following simple relationship between thermal width $W$ and boundary layer depth $D$:

$$ W = 0.3525D. $$

From this equation, boundary layer depths of 1000, 2000, 3000, and 4000 m lead to thermal widths of 352, 705, 1057, and 1410 m, respectively. Although these estimates appear to be slightly higher than typical observations (Stull 1988), it does provide a basic physical explanation of how a deep continental boundary layer could generate wider updraft velocities in deep convection (Lucas et al. 1994).

WS02 also argued that land–ocean contrasts in cloud microphysics and dynamics should be associated with cloud-base heights. For example, maritime environments generally have a smaller surface sensible-to-latent heat flux ratio (~0.1), more surface relative humidity (~80%), and a lower and warmer cloud-base height (~500 m), all of which tend to enhance the warm rain process and create weaker updrafts and suppress supercooled water in deep convection. On the other hand, continental environments tend to have a much higher surface sensible-to-latent heat flux ratio (0.2–1), less surface relative humidity (20%–60%), and higher and colder cloud-base heights (1000–4000 m), all of which tend to suppress warm rain processes and enhance supercooled water in the mixed-phase zone.

Larger concentrations of activated cloud condensation nuclei (CCN) associated with aerosols can increase the number concentration of cloud droplets while decreasing their sizes, which reduces the efficiency of warm rain processes. If further lifted above the 0°C-isotherm level, the supercooled water will form ice crystals through various nucleation processes, enhancing latent heat release via the condensation of supercooled water and deposition of ice crystals in convective cores, all of which invigorate deep convection (e.g., Rosenfeld and Woodley 2000; Khain et al. 2008; Tao et al. 2007, 2012).

Although not yet fully investigated, large nonhygroscopic aerosol particles can increase the concentration of ice-forming nuclei (IFN), which likely enhances heterogeneous ice nucleation and latent heat release in deep convection (Tao and Matsui 2015). Both CCN and IFN are more largely concentrated over land than ocean (Demott et al. 2010), which could explain continental convective invigoration.

Robinson et al. (2011) recently showed examples of overland convective invigoration through a set of idealized cloud-resolving model (CRM) simulations with various-sized flat islands surrounded by ocean. Regardless of the different sets of microphysics schemes and large-scale forcing, scattering of CRM-generated microwave brightness temperature ($T_b$) tends to become larger as the island size increases for different large-scale forcing and shows reasonable agreement with TRMM-observed microwave scattering (Williams et al. 2005; Zipser et al. 2006). They concluded that the dominant mechanism for convective invigoration over islands is mesoscale dynamics (pressure gradient) induced by the thermal contrast (the so-called “thermal patch” effect) between the island and surrounding ocean and argued that boundary layer, surface humidity, and aerosol impacts are not significant for convective vigor over islands.

Further investigation requires more observational evidence over the spectrum of convection types, from shallow to deep convection, as well as large-scale, high-resolution numerical experiments to better understand the physical mechanisms associated with land–ocean contrast. To this end, this study provides a climatological view of the contrast between oceanic and continental convective precipitating clouds from long-term TRMM satellite multisensor statistics (Matsui et al. 2009). Unlike previous studies, this one extends the observational analysis from shallow to deep convective precipitating clouds in terms of spatial variability as well as tropical land–ocean composites (section 2). This study is also the first attempt to investigate and understand how current global storm-resolving models can reproduce signals of land–ocean contrast in relation to satellite observations (section 3). Finally, the capabilities, limitations, and physical processes associated with the land–ocean contrast in convective systems are contrasted between the TRMM observations and the two global numerical cloud models (section 4).

2. Observed TRMM climatology

a. T3EF database

This study utilizes the TRMM Triple-Sensor Three-Step Evaluation Framework (T3EF; Matsui et al. 2009)
Although it contains subtropical zones, which is simply denoted as the tropics in this study, a land–ocean mask is utilized to identify land and ocean grids. Overall, spatial variability of precipitating cloud types and signal strength as well as land–ocean-grouped statistics over the entire tropics are compared for each 2.5° grid box to show the cloud-resolving models (Stephens et al. 2004; Masunaga et al. 2005), the phase of the Madden–Julian oscillation (Lau and Wu 2010), and for evaluating cloud-resolving models (Stephens et al. 2004; Masunaga et al. 2008; Matsui et al. 2009).

If a PR $H_{ET}$ is found, a precipitating cloud type is assigned to the column, depending on the VIRS Tb$_{IR}$ and PR $H_{ET}$ thresholds. Warm cloud tops (Tb$_{IR}$ > 260 K) and shallow echo-top heights ($H_{ET} < 4$ km) are assigned to the shallow warm (SW) category. A category with slightly colder cloud tops (Tb$_{IR}$ > 245 K) and higher echo-top heights ($4 < H_{ET} < 7$ km) was previously defined as congestus (Masunaga et al. 2005; Matsui et al. 2009). The Tb$_{IR}$ threshold of 245 K is based on that in Machado et al. (1998) for separating deep and nondeep clouds. However, this broad range of cloud-top temperatures also encompasses slightly deeper clouds than the traditional definition for congestus (~260 K; Johnson et al. 1999). Thus, in this study, this second class is denoted as midwarm. The midcold category represents either stratiform precipitation (Masunaga et al. 2005) or congestus overlapped by cirrus clouds (Stephens and Wood 2007) and has cold cloud-top temperatures (Tb$_{IR}$ < 245 K) and the same storm tops as midwarm ($4 < H_{ET} < 7$ km). The deep category includes deep stratiform areas and deep convection with cold cloud-top temperatures (Tb$_{IR}$ < 260 K) and significantly higher echo-top heights [$H_{ET} > 7$ km; see the schematics and discussion in Fig. 1 in Matsui et al. (2009)]. Since this study extends into subtropical and mountainous regions, a new category, shallow cold (Tb$_{IR}$ < 260 K and $H_{ET} < 4$ km), is introduced. This new category represents shallow cold precipitation in subtropical regions or in the mountains as well as warm precipitating clouds overlapped by cirrus clouds.

The horizontal extent of cloud systems (i.e., cloud clusters) is also an important physical parameter for understanding tropical precipitation processes (Mapes and Houze 1993; Machado et al. 1998). Masunaga et al. (2005) found that horizontal precipitation and cloud correlation lengths, measured from TRMM $H_{ET}$ and Tb$_{IR}$, consistently exceed 100 km in the midcold and deep categories, while they are limited to 8–18 km in the shallow warm and midwarm categories. Although the database does not directly characterize their horizontal extents, the $H_{ET}$–Tb$_{IR}$-based categories statistically suggest that the midcold and deep categories are more organized and clustered cloud–precipitation systems, while the shallow warm and midwarm categories are more isolated types.

Figure 1a shows joint Tb$_{IR}$–$H_{ET}$ diagrams over a 14-yr period for the entire tropics as well as their land–ocean difference. Based on the defined categories, the population of tropical precipitating cloud is 13.8% shallow warm, 17.9% shallow cold, 24.8% midwarm, 28.4% midcold, and 14.7% deep. Land–ocean differences in the Tb$_{IR}$–$H_{ET}$ diagram reveal larger proportions of the midwarm and deep categories over land but a much lower proportion over ocean.
larger proportion of the shallow warm category over ocean. Midcold clouds are separated into relatively warmer (ocean) and colder (land) $T_b$ IR modes, respectively. These statistics essentially highlight the climatological land–ocean contrast in terms of cloud–precipitation types.

Figure 1b shows the spatial variations of the normalized frequencies for each category averaged over the 14-yr
period by constructing the $T_{bIR} - H_{et}$ diagram at each $2.5^\circ \times 2^\circ$ grid. Thus, frequencies of five categories are summed up to 100% on each grid box. Shallow warm is the dominant precipitating cloud category over the southern portion of the Indian Ocean and the eastern portions of the Pacific and Atlantic Oceans. The total precipitation rate as well as the proportion of precipitating columns in these regions is very small, since most of these low clouds have no drizzle signal or are undetectable by the PR instrument (Matsui et al. 2004). The majority of shallow precipitation occurs over ocean (Schumacher and Houze 2003), while shallow precipitation frequency over land is limited to coastal regions.

Shallow cold is more frequently found near the subtropical boundaries because of the presence of wintertime midlatitude frontal systems and also over mountainous regions, such as the Rocky Mountains in North America, the western slopes of the Andes Mountains, and the Tibetan Plateau, where it is the most dominant precipitating cloud type (>70%).

The midwarm category mainly occurs over central North Africa over land and off the west coast of Namibia over ocean. Midwarm is most likely observed in regions where hot, dry continental air from deserts engages with warm moist air masses, such as the semidesert regions in Chad and Sudan, off the west coast of Namibia, the eastern part of the Arabian Peninsula, and the center of Australia. Note that this class was previously defined as congestus (Masunaga et al. 2005; Matsui et al. 2009); however, the climatological map of relative frequencies differs from that estimated from CloudSat and TRMM data with different thresholds (Wall et al. 2013). Thus, the midwarm class encompasses a broader range of precipitating clouds over tropical oceans than the congestus class (Masunaga et al. 2005; Stephens and Wood 2007).

The midcold category is most prevalent over the Pacific warm pool, the North and South Pacific convergence zones, the eastern portion of the Indian Ocean, and along the ITCZ over ocean. It also appears over central Africa, South America, India, and Southeast Asia over land. Because it has a large population with high rain intensities with larger clustering (Masunaga et al. 2005), the midcold category characterizes tropical precipitation variability. Therefore, the midcold frequency map closely resembles the precipitation climatology map (Wang et al. 2014). The highest (blue shade) midcold frequencies are observed off of the west coasts of Burma, Sumatra, Borneo, and Central America. These regions have the largest annual precipitation rates from nocturnal stratiform precipitation (Mapes et al. 2003).

Finally, the deep category appears most over land, such as West Africa, India’s Gangetic basin, and Argentina’s steppe regions where the most intense storms are typically observed (Zipser et al. 2006). These deep convective storms are commonly driven by strong CAPE, wind shear, and large-scale upward motion due to summertime monsoonal circulations and strong surface insolation. The highest frequencies (black shade) are found over Lake Chad in West Africa. In this region, up to 50% of precipitation pixels have storm heights greater than 7 km and cloud-top temperatures colder than 260 K.

c. Microphysical properties associated with categorized reflectivity CFADs

The second step in T3EF is to construct contoured frequency with altitude diagrams (CFADs; Yuter and Houze 1995) of PR reflectivity separately for the categories defined in the first step (section 2b) in order to investigate precipitation microphysics (Matsui et al. 2009). Reflectivity CFADs were constructed for the entire tropics as well as their land–ocean difference by binning the reflectivities into 1-dBZ bins at each height increment (250 m; Fig. 2a). Note that the sharp increase in reflectivity near the ground evident in all classes is most likely due to surface clutter.

Reflectivity intensities for the shallow warm and shallow cold categories are the weakest. Modal and maximum reflectivities are ~25 and 45 dBZ, respectively. Land–ocean differences in their CFADs show that both of these shallow categories have narrower reflectivity distributions over land than over ocean, probably due to a lack of moisture as well as larger concentrations of aerosols. The midwarm and midcold categories have larger modal and maximum reflectivities than the shallow warm and shallow cold categories, and their land–ocean contrast shows that the land reflectivities tend to be more widely distributed, suggesting a larger variability of precipitation particle sizes (and rainfall intensity) over land.

Shallow warm and midwarm show continuously increasing reflectivities toward the ground and suggest raindrop growth via coalescence processes, while the midcold category has a subtle brightband signal around 5 km, suggesting the presence of melting ice particles. Overlying solid particles above 7 km are invisible to the PR but not to infrared Tb and high-frequency microwave Tb [as shown in the next section and in the contrast between Figs. 3 and 4 in Matsui et al. (2009)].

The most dramatic transitions in the reflectivity distributions are in the deep category where there are three distinct zones: the solid phase [i.e., above 8 km, around $-20^\circ$C (Hashino et al. 2013)], where the presence of
solid precipitation particles statistically generates narrow reflectivity profiles; the mixed phase (i.e., between 5 and 8 km), where the aggregation and melting of frozen particles dramatically increase reflectivity distributions; and the liquid phase (i.e., below 5 km), where liquid raindrops dominate the radar backscattering signal. In this lowest, liquid-phase zone, profiles of near-constant reflectivity distribution suggest raindrop size distributions are close to an equilibrium state through collision–coalescence growth and

![Images of reflectivity profiles and spatial variability](image-url)
breakup processes (McFarquhar and List 1991) in a statistical sense.

Note that the CFADs for this class include data from various echo-top heights (i.e., 7–20 km). The reason for the relatively narrow echo distributions around 7 km is the dominant sampling of convection with echo-top heights of 7–10 km (e.g., Fig. 1a). Alternatively, convection with echo-top heights greater than 10 km is statistically very small, which often frustrates the interpretation of the CFADs. It should be understood that CFADs from a large sample volume would be slightly different from those from a single convective episode (e.g., Li et al. 2010).

Tao and Matsui (2015) decomposed a reflectivity CFAD from a cloud simulation with a bulk microphysics scheme (see their Fig. 8). Their study suggests that the majority of the mode reflectivities in the solid-phase zone are dominated by snow aggregates from the deep stratiform portion of their simulated MCS, while the infrequent occurrence of intense echoes is predominantly contributed by hail within the convective cores. In this way, strong reflectivities from a small area of convective cores and weak reflectivities from a large area of stratiform precipitation characterize the radar CFAD above the melting layer in what would be the deep category.

The land–ocean difference in CFADs also shows a clear contrast between maritime and continental environments. In the solid-phase zone (>8 km AGL, below -20°C), the land echo distributions tend to stronger values (red shading), while the ocean PR echo frequencies are more concentrated below -24 dBZ, which suggests a larger convective core fraction in the continental deep convective clouds. Below 8 km, continental echoes are more broadly distributed than oceanic, as illustrated by the central blue mode shade surrounded by the red shading. This indicates that continental precipitation has a wider particle size spectrum (i.e., more smaller and more larger) than oceanic.

To better understand the large variability in the deep category, the spatial variability of echo-top height and maximum column echo are examined. PR CFADs are constructed for each 2.0° × 2.5° grid box over the tropics; the 95th percentile is then used to estimate the maximum PR echo-top height and column echo for the deep category. Figure 2b shows that deep convection tends to have larger maximum column echoes (up to 50 dBZ) and taller echo-top heights (up to 15 km) over major continents than ocean. Continental convective invigoration is also apparent over the major islands in Southeast Asia, similar to the findings of Williams et al. (2004) and Robinson et al. (2011).

d. Ice scattering from microwave brightness temperatures

The third step of T3EF is to analyze the distributions of microwave Tb scattering. Scattering from high-frequency microwave channels is more directly associated with the path-integrated frozen hydrometeor amount than the surface precipitation rate, if there is enough ice aloft in the atmosphere. Otherwise, surface scattering signals dominate the scattering signals measured from the TMI. The 85-GHz TMI channels are fairly sensitive to smaller-sized frozen particles, which are often undetectable by the PR (Matsui et al. 2009). Because the TMI sampled mixed land–ocean areas over the tropics, a polarization-corrected brightness temperature (PCTb85) is used to compensate for the inhomogeneity in surface emissivity (Spencer et al. 1989) via

\[ PCTb_{85} = Tb_{85V} + a(Tb_{85V} - Tb_{85H}), \]

where \( Tb_{85V} \) and \( Tb_{85H} \) are the Tb from the vertical and horizontal polarization channels at 85 GHz, respectively. Based on Matsui et al. (2009), a fixed value of \( a = 0.8 \) generally works well over the tropics with the exception of some regions. This is close to the value (0.818) used in Spencer et al. (1989).

Figure 3a shows cumulative probability distributions of PCTb85 (the bin size is 10 K) for all categories integrated over land and ocean. For each class, the microwave scattering index (MSI) in this study is defined as

\[ MSI = \frac{PCTb_{85\mid 95\%} - PCTb_{85\mid 5\%}}{PCTb_{85\mid 95\%} - PCTb_{85\mid 5\%}}, \]

where PCTb85|5% and PCTb85|95% are the 5th percentile and 95th percentile of the cumulative distributions for each category in order to remove statistical outliers. Figure 3 shows cumulative distributions of PCTb85 and MSI for the entire tropics as well as their land–ocean differences.

It is quite discernible that the probability distributions trend toward lower PCTb85 values as the cloud types progress from shallow warm to midwarm to shallow cold to midcold to deep categories. MSI ranges from 29.82 to 100.2 K, which suggests that the amount of frozen particles increases substantially toward deep categories. Land–ocean differences show that the deep category is positive below 220 K and negative above and that the midcold category is positive below 240 K and negative above. This suggests that continental convection tends to have more solid ice processes than oceanic. Also note that there are substantial positive and negative variations in other categories, but these are most likely due to land surface signals.
To assess spatial variability, PCTb₈₅ probability distributions and MSI are computed for each 2° × 2.5° grid box to show the geographical distribution of ice scattering (Fig. 3b). MSI for the shallow warm category ranges from 0 to 30 K over ocean and from 10 to 40 K over land, which appears to be natural variability (gray shade) in the background microwave emission. Over the Tibetan Plateau, the MSI for the shallow warm category
is anomalously large (up to 60 K) most likely because of the presence of surface snow (Pulliainen and Hallikainen 2001). The polarization correction most likely failed to mask the surface signals because of the lack of column water vapor over the Tibetan Plateau. Note that the scattering signals for shallow warm and midwarm are also affected by the slant angle of the TMI sensor viewing path, which often includes signals from neighboring cells with cold precipitation processes.

The MSI for midcold ranges from 10 to 70 K. A large MSI is associated with the regions where midcold is most frequently observed (Fig. 1b). The deep category has the largest variability of MSI, up to 110 K. Over ocean, the MSI for deep generally ranges up 80 K, but in very limited regions such as the tropical warm pool, it can reach 90 K. The MSI for the deep category is clearly larger over land, including from central to southern Africa, India, Southeast Asia, northern Australia, and North and South America. Even over the islands in Southeast Asia, the MSI is typically 10–20 K larger than the surrounding ocean, although the air mass should be maritime over land. This suggests that continental deep convection produces larger ice water paths than oceanic even within the same tropical air mass.

e. Mechanisms and robustness of land–ocean contrast

Overall, Figs. 1–3 reconfirm that continental deep convection tends to have taller echo-top heights, larger maximum column reflectivities, and larger microwave scattering signatures than does maritime deep convection in agreement with previous studies (Williams et al. 2004; WS02; Zipser et al. 2006). Figure 4 summarizes the potential pathways for the invigoration of continental deep convection based on the aforementioned previous studies presented in section 1. There are potentially four factors that could invigorate deep convection over land: 1) amplified CAPE; 2) high surface sensible heat fluxes that lead to elevated cloud-base heights or the height of the lifting condensation level (HLCL), which then reduce warm cloud depth while also deepening the PBL depth and result in enhanced convective width and updraft velocity; 3) the thermal patch effect wherein islands or land-cover patterns enhance mesoscale pressure gradients, wind convergence, and consequently updraft velocity; and 4) higher aerosol concentrations that can reduce warm rain process via cloud nucleation. As evidenced from the TRMM climatology, all of these factors potentially can lead to the continental vigor of cloud and precipitation processes, namely, the enhancement of deep convective processes and the concurrent suppression of warm rain processes (Fig. 1), along with an enhanced amount of supercooled water and more heavily rimed ice particles (Fig. 3), which result in deeper convection with more intense surface precipitation (Fig. 2).

Instead of using long-term climatology, this section briefly discusses the robustness of land–ocean contrast based on shorter time scales. Matsui et al. (2015) proposed the idea of convection-microphysics quasi-equilibrium (CMQE) states through a planetary view of tropical convection. They found that the characteristic spectrums of TRMM precipitation signals (radar echo and microwave brightness temperature spectra), precipitation rate, and microphysics states are nearly identical regardless of seasons and years as long as the sampling covered the entire tropics, regardless of the day-to-day, seasonal, or interannual variability of tropical dynamics.

Figure 5a shows a monthly time series of the difference in composite PR CFADs for land minus ocean at midlevels (i.e., 3 km for shallow warm and shallow cold, 3.5 km for midwarm and midcold, and 10 km for deep). There are some subtle but consistent land–ocean differences in the shallow warm, shallow cold, and midwarm categories, close to the climatology difference in Fig. 2a. The midcold category, however, shows seasonal cycles in its land–ocean contrast. The most significant and consistent differences appear in the deep category, wherein continental deep clouds always show radar reflectivity invigoration within the solid-phase zone regardless of different months and years as long as the sampling is over the entire tropics.

Figure 5b shows a monthly time series of the difference in PCTb85 frequency between land and ocean in a similar manner to Fig. 5a. The PCTb85 distributions show clear land–ocean differences for all categories. Consistent positive signals for the shallow warm and shallow cold signals are due to land surface signals, which are mostly likely not associated with differences in the microphysical characteristics between land and ocean environments. Midwarm signals are also inconsistent throughout the time series so that there is no time-consistent land–ocean PCTb85 because of the lack of appreciable amounts of ice in this class. At times, midwarm over ocean has PCTb85 probability densities that are more narrowly distributed, but the results are relatively sporadic. The midcold category has more distinct land–ocean differences in PCTb85, though the PR CFAD results are nearly identical between land and ocean. This could be because of a difference between land and ocean in the tiny precipitating ice to which the PCTb85 is sensitive and/or because the TMI slant beam path includes deep category signals. The deep category has the most distinct and consistent differences, with the overland microwave
signal always having stronger scattering because of the larger amount of precipitating ice in the mixed-phase zone.

These results confirm that the continental invigoration of deep convection is always present, regardless of the month and year. As suggested by the CMQE hypothesis (Matsui et al. 2015), the land–ocean contrast could be evident at even finer temporal resolutions (e.g., daily) provided there is a sufficient number of convective samples. Alternatively, land–ocean contrast can also be investigated at short time periods if enough sampling can be obtained from high-resolution global simulations.

3. Land–ocean contrast in global storm-resolving models

a. Global storm-resolving models

Using the new observational benchmark for convective land–ocean contrast through T3EF, the main objective of this section is to evaluate for the first time the land–ocean contrast simulated by global high-resolution model simulations over a relatively short time period. The impact of aerosols, however, is not investigated, as it is not predicted in the models. The simulations that will be tested are from the NASA Multiscale Modeling Framework (MMF) and the Nonhydrostatic Icosahedral
Cloud Atmospheric Model (NICAM). Both simulations are designed to resolve organized convection with horizontal grid spacings of 4 and 3.5 km, respectively.

The NASA MMF is based on the Goddard Earth Observing System, version 4 (GEOS4), with the convection and microphysics parameterizations replaced by explicit 2D CRM simulations using the Goddard Cumulus Ensemble (GCE) model (Tao et al. 2009, 2014). The MMF uses the Community Land Model (CLM; Bonan et al. 2002) to predict the land surface turbulent heat flux and skin temperature in each GEOS4 grid. The MMF was initialized with European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim; Dee et al. 2011) and integrated over the entire month of June 2008.

NICAM was developed to study the capabilities of fully 3D global CRMs (Tomita and Satoh 2004; Satoh et al. 2008, 2014). The NICAM experiment in this study was designed to study Typhoon Fengshen from its genesis stage to its mature stage (Hashino et al. 2013). The winds, temperature, relative humidity, and geopotential heights in the NICAM simulation were initialized with the 0.5° ECMWF Year of Tropical Convection (YOTC) analysis at 0000 UTC 15 June 2008 (Moncrieff et al. 2012) and integrated for one week only. The surface variables such as sea surface temperature (SST), sea ice cover, and soil moisture are initialized with 1° National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalyses.

The MMF and NICAM use the Goddard single-moment four-class ice (4ICE; Lang et al. 2014) and the NICAM Single-Moment Water 6 (NSW6; Tomita 2008) cloud microphysics schemes, respectively. Both 4ICE and NSW6 are based on the Lin et al. (Lin et al. 1983) and Rutledge and Hobbs (Rutledge and Hobbs 1983) schemes, but NSW6 omits the wet growth of hail in order to reduce computation cost. The 4ICE scheme has been developed based on a series of improvements using observations (Tao et al. 2003; Lang et al. 2007, 2011). The most recent updates (Lang et al. 2014; Tao et al. 2016) include newer diagnostic snow and graupel...
size distributions and a new frozen drops–hail category for severe thunderstorms as well as other improvements. Table 1 summarizes the MMF and NICAM setup and physics options.

NICAM has a realistic land mask, terrain, and land-cover specification, permitting thermal patch effects due to sea breezes and/or land-cover heterogeneity, while the NASA MMF only generates homogeneous surface energy and turbulent fluxes driven by the GEOS4 CLM. Thus, both the MMF and NICAM could be used to examine the impact of variations in the HLCL and large-scale CAPE, while only NICAM is capable of producing a realistic thermal patch effect (Fig. 4).

b. Satellite simulators

Model-simulated geophysical parameters (including cloud and precipitation microphysics information) from the MMF and the NICAM simulations are converted into TRMM signals through the Goddard Satellite Data Simulator Unit (G-SDSU; Matsui et al. 2013, 2014) and the Joint Simulator for Satellite Sensors (Hashino et al. 2013). Both multi-instrumental simulators employ identical microwave simulators (Kummerow 1993), radar simulators (Masunaga and Kummerow 2005), and IR simulators (Nakajima and Tanaka 1986). Details on the computation methods are described in previous studies (Matsui et al. 2009, 2014).

Briefly, the effective dialectic function is calculated from the Maxwell–Garnett assumption, and single scattering properties are calculated from the Mie assumption for both microwave and radar simulators. The microwave simulator can account for slant-path beams, mimicking the 3D slant-path observations in conical-scanning microwave radiometers, such as the TMI. These forward models are consistent with the physics assumptions (including the particle size distributions, phase, and effective densities) within the microphysics schemes used in the MMF and NICAM. VIRStb_{IR}, PR H_{ET}, PR CFAD, and TMI PCTb_{85} are simulated in each of the MMF columns and the NICAM output and sampled in an identical manner to the TRMM observations.

c. Simulated T3EF diagrams from the MMF and NICAM

Figure 6a shows Tb_{IR}–H_{ET} diagrams derived from TRMM observations, the MMF simulation, and the NICAM simulation for the entire tropics over both land and ocean for June 2008. Figure 6b shows the corresponding land–ocean differences. The June TRMM Tb_{IR}–H_{ET} diagram for the combined land–ocean average has a very similar pattern to the climatology (Fig. 1a). Also similar to the climatology, monthly land–ocean differences in the Tb_{IR}–H_{ET} diagram reveal larger proportions of the midwarm and deep categories over land but a much larger proportion of the shallow warm category over ocean.

The structure of the MMF Tb_{IR}–H_{ET} diagrams is somewhat closer to the TRMM observations. However, its joint PDF is skewed more toward the shallow warm and shallow cold categories and misses a large portion of midwarm. The shallow cold category is nearly twice as much as for the TRMM observations. The land–ocean contrast for the MMF appears to be quite good in agreement in terms of its larger proportion of maritime shallow warm (blue shading in zone 1) and continental midwarm (red shading in zone 2). The MMF, however, produces a higher portion of vigorous convection over ocean as depicted by the blue shading in the deep category, unlike the TRMM observations. Also, its maritime midcold peak is skewed toward shallow cold (zone 5, near echo-top heights of ~4 km). As a result, midcold appears at a higher frequency over land.

The structures of the NICAM Tb_{IR}–H_{ET} diagrams appear to be somewhat different from the TRMM observations as well as from those of the MMF. The shallow warm category is over twice the TRMM-observed frequency, while the deep distribution extends to much higher H_{ET} (>10 km) than do the TRMM observations or the MMF simulations, which occur predominantly
under 10 km. Similar to the MMF results, there are no distinct peaks in the midwarm and midcold zones. NICAM produces the correct land–ocean contrast in the form of higher frequencies over land for the deep category and higher frequencies over ocean for the shallow warm, though they are too strong and too weak, respectively. The continental midwarm peak is also quite evident, but the echo-top heights are too shallow relative to the TRMM observations and as a result appear as shallow warm.

Since the NICAM integration time is much shorter than the TRMM observations and the MMF simulation, the spatial distributions of these precipitating cloud types are not discussed in the main article. However, the reader is encouraged to view the detailed spatial maps and discussion in the supplemental material (supplement A).

Figure 7a shows PR CFADs for the deep category from the TRMM observations, the MMF simulation, and the NICAM simulation for the entire tropics over both land and ocean for the same time period. Both the MMF and NICAM reproduce reasonable PR CFAD structures. At upper levels (>8 km AGL), the MMF echoes are more narrowly distributed because of the specified snow aggregate and graupel size distributions and effective densities (Lang et al. 2014). In contrast, the NICAM CFAD is more broadly distributed and tends to overestimate PR echoes at all levels.

Figure 7b shows the corresponding land–ocean contrast in PR CFADs for the deep category. The monthly TRMM observations show a clear land–ocean contrast in the PR echo distributions, nearly identical to the climatology (Fig. 2). The MMF simulation has a mixture of results. Above 10 km, continental echoes are shifted to larger values as observed; however, below 10 km, the oceanic distributions are broader (i.e., the red shading is surrounded by blue), which is opposite to the
It is probably due to the invigoration of oceanic deep convection in the MMF simulation (Fig. 6b). The NICAM simulation shows signs of continental invigoration above 5 km similar to the observations, but the signal is weaker than the TRMM observations and less coherent throughout the vertical range. In the warm precipitation zone below 5 km, as with the MMF, the results are counter to TRMM with broader oceanic distributions. The results are discussed more in the next section.

Figure 8a shows cumulative frequencies of PCTb85 (with a bin size of 10 K) for the deep category from the TRMM observations, the MMF simulation, and the NICAM simulation. The MMF-simulated PCTb85 is distributed over much colder temperatures than the observations. NICAM also overestimates the PCTb85 depression; however, it performs better than the MMF. This indicates that the MMF and NICAM overproduce solid precipitation particles in deep convection environments, which has been a common problem in CRMs because of unresolved microphysics and dynamics (e.g., Varble et al. 2014).

The TRMM land–ocean difference (Fig. 8b) in PCTb85 is positive below 240 K and negative above and suggests that continental convection tends to have more solid ice processes than oceanic. The MMF land–ocean difference in PCTb85 is more exaggerated; positive frequency differences (i.e., a higher land frequency) reach nearly 6.5% from ~170 to 180 K. There are small negative deviations above ~220 K, but they are rather weak. The MMF therefore does not reproduce continental invigoration in its PR signals, but it does in its TMI signals. The positive PCTb85 frequency deviations below ~220 K in NICAM are comparable to the TRMM observations but the negative deviations (i.e., higher oceanic frequencies) are lacking.
To verify the background model dynamics, the updraft velocities and thermodynamic states from the MMF and NICAM simulations are investigated. Figure 9 shows CFADs of vertical velocity over the entire tropics as well as their land–ocean difference. The MMF CFAD exhibits two subtle downdraft modes that peak around 2–4 km and 12–14 km of altitude, while peak updraft velocities appear near 10 km of altitude. The MMF CFAD shows that 99.9% of all vertical velocities (color shades) are less than 5 m s\(^{-1}\). At a frequency of 10^{-5}% (outer edge of darkest gray shade), updrafts reach up to 20 m s\(^{-1}\). The land–ocean difference shows a wider spread of updraft velocities (i.e., a higher portion of strong updrafts and strong downdrafts, red shading) over land, especially from 2 to 10 km of altitude. This might explain the MMF’s land dominance in the second and third categories of the \(\text{Tb}_{\text{IR}} - H_{\text{ET}}\) composites (Fig. 6b).

The NICAM CFAD exhibits two distinct modes of stronger downdrafts near ~14 and from 2 to 4 km. Peak updraft velocities occur slightly higher in NICAM (~12 km) than in the MMF (~10 km) and are more intense at the 10^{-5}% frequency, up to 27 m s\(^{-1}\). The difference between land and ocean shows a wider distribution of updraft velocities over land, especially from 12 to 18 km of altitude. This explains the continental vigor in deep convection in the NICAM simulation. It requires more case-by-case comparison with the observed quantities in future study (e.g., Heymsfield et al. 2010).

Figure 10a shows the environmental thermodynamic states over land and over ocean, namely, the CAPE and the HLCL in clear-sky conditions (due to the data availability) from the Aqua Atmospheric Infrared Sounder (AIRS) level 2 data and all-sky conditions from the MMF and NICAM simulations. For larger CAPE (>2000 J kg\(^{-1}\)), the MMF’s frequency curves over land and ocean are nearly identical to each other; the same is true for NICAM. In comparison to the observations, the MMF’s oceanic CAPE curve is closer to the AIRS pattern, while over land the NICAM’s CAPE distribution is much closer to the AIRS pattern than the MMF’s. However, the AIRS retrievals are severely limited over land because of cloud contamination for the larger CAPE conditions, and AIRS sampling took place during the local early afternoon. Furthermore, it is technically difficult to derive CAPE values only prior to the onset of deep convection from both the AIRS observations and the MMF/NICAM simulations. Regardless of the uncertainties, it is confirmed that CAPE between land and ocean over tropical regions shows a relatively small contrast in each of the MMF and NICAM simulations, consistent with previous observational studies (WS02; Williams and Renno 1993; Lucas et al. 1994).

HLCL clearly shows a strong land–ocean contrast, especially for AIRS and the MMF. Maritime conditions...
tend to have a significantly shallower HLCL because of the dominance of surface turbulent latent heat fluxes, while HLCL values are much deeper over land because of the dominance of turbulent sensible heat fluxes. The MMF has a higher frequency of very deep HLCLs over land than does NICAM. This suggests that NICAM is characterized with moister boundary layer conditions than the MMF over land, which results in NICAM having shallower HLCL values. This then leads to the NICAM biases associated with the lack of separation between shallow warm and midwarm clouds (Fig. 6b). On the other hand, drier boundary layer conditions in the MMF likely suppressed warm rain processes but increased production of ice crystals, resulting in the larger ice scattering signals over land (Figs. 4, 8), despite having similar CAPE over land and ocean.

Figure 10b shows the lowest-layer relative humidity (RH) and surface evaporative fraction (EF; latent heat flux over total latent and sensible heat flux) from the MMF and the NICAM simulations. Over ocean, the MMF and NICAM show very similar distributions and are largely skewed toward larger RH and EF values. Over land, the MMF has broader distributions of RH and EF, with NICAM having narrower, moister distributions. This suggests that NICAM tends to have larger surface latent heat fluxes that induce higher boundary layer humidity and lower HLCLs than in the MMF. This subsequently lowers the cloud altitude for the continental midwarm type into the shallow warm class, resulting in an overestimation of the shallow warm class in the NICAM simulation.

4. Summary and discussion

An initial attempt at evaluating the land–ocean contrast in convection and microphysics processes from two global storm-resolving models has been conducted. To some extent, the MMF simulations reproduced the observed pattern of land–ocean contrast for the shallow warm and midwarm types (Fig. 6b). This is most likely because the MMF reproduced a realistic land–ocean contrast in HLCL (Fig. 10a). These results suggest the importance of the pathway associated with high sensible heat fluxes over land in Fig. 4 (WS02). Drier surface
fluxes also elevate boundary layer heights, which would favor the creation of more midwarm clouds as shown by the $Tb_{IR}$–$H_{ET}$ diagrams (Fig. 6b). The NICAM simulation tends to have much wetter surface soil conditions, which induces a larger evaporative fraction and higher relative humidity. As a result, it lowers cloud altitudes from midwarm to shallow warm types, resulting in its continental midwarm (i.e., congestus) mode forming more to the shallow warm category (Fig. 6b). As a result, the NICAM simulation overestimates the shallow warm class over the tropics (Fig. 6a). Supplemental material also shows extensive and excessive distributions of shallow warm clouds over desert region (supplement A).

Although the magnitude and extent were not quantitatively consistent with the observations (Fig. 8a), both the MMF and NICAM simulations generated more PCT$b_{85}$ ice scattering in continental convection (Fig. 8b), suggesting they produced more solid ice particles in their deep continental convective mode despite the large CAPE distributions being very similar between land and ocean (Fig. 10a). However, neither the MMF nor NICAM generated a realistic land–ocean contrast in their radar signals as depicted by PR CFADs (Fig. 7b). Structural differences in the simulated CFADs between land and ocean are less coherent in the vertical direction, unlike those from the TRMM PR, which show clear, coherent patterns through solid, mixed, and liquid-phase precipitation.

The results are also interesting because the MMF modeling setup does not account for the thermal patch effect. Robinson et al. (2011) concluded that mesoscale wave dynamics due to the thermal patch effect is the primary mechanism for continental convective
vigor over islands. At least, despite homogeneous surface fluxes in the MMF and nearly identical CAPE distributions between land and ocean, stronger scattering signals were simulated in continental deep convection primarily because of the drier surface conditions (i.e., higher HLCL; WS02). Of course, these results do not reject the finding in Robinson et al. (2011), because the MMF simulation does not reproduce more frequent deep convection (as measured by Tbl and HET) over land, while the NICAM simulation does. From these simulations, it can be concluded that 1) drier surface fluxes contribute to the generation of midwarm types of clouds and glaciation of deep convection and 2) the thermal patch effect is important for generating more frequent deep convection and taller storm-top heights in continental deep convection.

Quantifying the surface sensible heat flux impact and thermal patch impact requires that the physics, grid configurations, and initial and boundary conditions between the MMF and NICAM be more similar. The overall results indicate that even these advanced global storm-resolving modeling systems have room to improve their microphysics and dynamics in order to replicate the nature of the observed land–ocean contrast in convection, as suggested by previous evaluation studies (Masunaga et al. 2008; Inoue et al. 2010; Satoh et al. 2010; Kodama et al. 2012; Roh and Satoh 2014).

There are some pathways for improving the weaknesses in the simulations associated with the surface conditions, resolution, and the sophistication of the microphysics. The spinup of soil moisture for initial conditions in NICAM and MMF could potentially be improved for more realistic CAPE but would require a more careful setup for future experiments (Mohr 2013). Although traditionally known as “cloud-resolving models,” a typical grid spacing of 1–4 km and ∼60 vertical levels will only resolve cloud systems greater than ∼5–40 km (Pielke 2013). Khairoutdinov et al. (2009) showed a clear improvement in simulating shallow/congestus populations when the horizontal grid spacing was reduced to 200 m and vertical levels increased to 256. It is also important to couple with realistic aerosol simulations to obtain CCN and IFN in order to investigate whether aerosols really help to characterize a realistic convective land–ocean contrast in future (Saleeby and van den Heever 2013).

It must be emphasized that analyzing global storm-resolving models provides a more comprehensive pathway for more universal understanding, including the land–ocean contrast. However, this approach requires a lot more computing resources, but such an improvement is needed to fulfill the desire to simulate a more realistic land–ocean contrast (Lucas et al. 1994; Williams et al. 2002, 2005; WS02; Liu and Zipser 2005).

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