Sensitivity of seasonal climate and diurnal precipitation over Central America to land and sea surface schemes in RegCM4

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ABSTRACT: Multi-annual simulations over the Central America CORDEX domain are conducted with the latest version of regional climate model RegCM4 driven by ERA-Interim reanalysis fields. The RegCM4 system can reproduce both the annual cycle and the spatial patterns of mean summer precipitation over Central America and Mexico. Regional circulation features are also reproduced, although the intensity of the Caribbean Low-Level Jet is underestimated and it is located too far south. Over most land areas, RegCM4 surface air temperatures are lower than observations by 1 to 3°C, which however may also be related to biases in the reanalysis forcing data. The model can realistically simulate the amplitude of the convective diurnal cycle in areas where the convective triggering is dominated by non-local gravity wave effects. However, the simulation of the phase of the diurnal cycle of convection is less satisfactory, with the peak precipitation occurring earlier than observed, a common fault in atmospheric models. Sensitivity experiments are carried out to investigate the model sensitivity to land surface and a prognostic diurnal sea surface temperature scheme. Use of the Community Land Model (CLM) instead of the Biosphere-Atmosphere Transfer Scheme (BATS) results in a warmer and drier land surface and a better simulation of the seasonal average spatial pattern of precipitation. However, with BATS, RegCM4 has a more realistic simulation of the mid-summer drought over the region. The impact of the prognostic sea surface temperature (SST) scheme is generally small. In general, neither of these surface physics upgrades results in a clearly superior model performance.

KEY WORDS: RegCM4 · CORDEX · Central America · Diurnal precipitation · Climate · Diurnal SST scheme · CLM

1. INTRODUCTION

The climate of Central America is characterized by complex spatial and temporal features. Rainfall over Central America comes in synoptic events ranging in intensity from easterly waves to hurricanes, but much of it is associated with moisture surges from the Pacific or Caribbean that impinge upon its topography (Hastenrath 1967, Waylen et al. 1996, Peña & Douglas 2002). Moreover, mesoscale circulation features such as the Caribbean Low-Level Jet (CLLJ) also affect rainfall (Wang 2007, Wang & Lee 2007, Muñoz et al. 2008, Cook & Vizy 2010). On the western side of the Central American isthmus, the eastern Pacific inter-tropical convergence zone (ITCZ) brings precipitation as it migrates northward during the northern summer season (Hastenrath 1967). A particularly interesting feature of Central America’s precipitation patterns is that parts of the region display a bimodal precipitation annual cycle, with maxima occurring in June and in either September or October, separated by a lull in July and August, with tropical cyclone activity in the Caribbean following a similar pattern (Wang & Lee 2007). This period of lower rainfall is called the mid-summer drought (MSD) (Magaña et al. 1999).
Relatively few regional climate modeling studies have been performed over domains that include Central America, and most have limited their analyses to aspects of the seasonal mean climate. For instance, Hernandez et al. (2006) tested MM5 over a small Central American domain for a single year (1997). They showed large positive biases in precipitation compared to observations, particularly over the Caribbean Sea, although stations over land showed better agreement; MM5 simulated precipitation totals within 25% of the observed. Over a similarly small domain, Karmalkar et al. (2008) used the Hadley Centre PRECIS model to simulate climate for the 20th (1961–1990) and 21st century (2071–2100) at high resolution (0.22° grid-spacing) for the dry winter season (December-January-February, DJF). They noted model deficiencies in simulating both temperature and precipitation at high altitudes, with negative biases in both cases. Martinez-Castro et al. (2006) completed a sensitivity study of the Caribbean climate to domain size, resolution and convective schemes in RegCM3 for the 1993 summer period, finding substantial sensitivity to all 3 factors explored. Tourigny & Jones (2009a,b) used the Rossby Centre regional atmospheric model (RCA3) at 0.33° grid spacing and a series (1979–2005) of 13 mo long re-initialization integrations (with a seasonal forecasting framework) forced by ECMWF re-analysis to examine the simulated climate over Mexico, Central America, the Caribbean, and northern South America. They found that the model captured both the precipitation annual cycle over Mexico and the Caribbean and the precipitation anomalies in most areas associated with ENSO events. However, the monthly precipitation minimum in July and August that is representative of the MSD climate was not simulated by their model. Clearly, more work is needed to assess the performance of regional climate models (RCMs) in reproducing the complex climatic features of Central America, which are affected by both remote processes related to large scale atmospheric and oceanic teleconnection patterns, such as ENSO and the North Atlantic Oscillation (e.g. Hastenrath 1967, Enfield 1996, Giannini et al. 2000), and local processes due to complex topographical, land use and coastline features (Hastenrath 1967, Waylen et al. 1996, Peña & Douglas 2002).

Further emphasizing the need for regional climate assessments over Central America are recent studies suggesting that Central America could be a climate-change hot-spot during the 21st century due to consistent model projections of future drying over the region (e.g. Giorgi 2006, Neelin et al. 2006, Rauscher et al. 2008, 2011). Recognizing the importance of the Central America region within the climate change context, the newly implemented international program Coordinated Regional Downscaling Experiment (CORDEX) (Giorgi et al. 2009; see also http://wcrp.ipsl.jussieu.fr/SF_RCD_CORDEX.html), includes this region as one of the focus domains. These CORDEX experiments will provide an opportunity to evaluate the performance of RCMs over the region and to produce a new generation of RCM-based climate change projections. The first step in the CORDEX protocol is to assess models using lateral boundary conditions from reanalysis of observations. Here we present a detailed analysis of the performance of an RCM (the newly developed RegCM4; Giorgi et al. 2012, this Special), with a focus on the seasonal and diurnal cycle characteristics of precipitation over Central America. The analysis is conducted for the northern hemisphere summer months (June to September) since most of Central America experiences its rainy season during this period (Hastenrath 1967).

One aspect of regional climate that is receiving increased attention is the diurnal cycle of precipitation. While it has not been conclusively demonstrated, it is claimed that a good simulation of the diurnal cycle is necessary for representing the mean climate and its variability on longer time scales (e.g. Yang & Slingo 2001). A number of studies have analyzed the diurnal cycle of precipitation using observations and/or global and regional climate models (Wallace 1975, Dai et al. 1999, Dai 2001, Yang & Slingo 2001, Collier & Bowman 2004, Dai & Trenberth 2004, da Rocha et al. 2009). Although these studies mostly agreed that in many tropical regions, precipitation peaks in late afternoon/early evening over land and late night/early morning over ocean, they noted that there are regional differences because of land–sea breezes, mountain–valley circulations, and offshore ocean regions affected by gravity wave-triggered convection (Mapes et al. 2003). Errors in the diurnal precipitation cycle may affect the intensity and frequency of simulated precipitation. For example, many convection schemes allow convection to begin too early in the day (related to the representation of the convection trigger; Dai & Trenberth 2004, Lin et al. 2000), and as a result over-estimate light precipitation events. While such errors may be masked in monthly mean precipitation statistics, they can nonetheless contribute to biases in other fields, such as cloudiness (Dai & Trenberth 2004). Moreover, any model deficiencies in physical parameterizations (e.g. boundary layer, convective parameterization) that are related to the surface heating can negatively
impact the simulation of the diurnal cycle. Closer to the study area of this paper, da Rocha et al. (2009) recently examined the diurnal cycle of precipitation using RegCM3 over South America and found that the phase of RegCM3 matches observations (by capturing the afternoon peak) over the continental South American convergence regions. For Central America, due in part to the topography of the isthmus, the diurnal precipitation cycle is complex and shows both sub-regional and seasonal variations (Curtis 2004).

While many previous studies of the diurnal cycle have focused on deficiencies in the convective parameterization physics itself, such as its sensitivity to entrainment of dry air or triggering function (e.g. Betts & Jakob 2002), the atmosphere−surface interaction can also be important for forcing diurnal variability as well as for determining the mean climate. Thus, in addition to evaluating the performance of RegCM4, this paper investigates the model sensitivity to different land surface schemes available in the model and to the implementation of a simple prognostic scheme that represents the diurnal variation of SST.

2. DATA, MODEL AND METHODS

2.1. Data

Three observational datasets are used to evaluate the model seasonal precipitation: the monthly mean products from the Tropical Rainfall Measuring Mission (TRMM 3B43; Huffman et al. 2007), the Climate Research Unit (CRU; Mitchell & Jones 2005) and the Global Precipitation Climatology Project (GPCP; Adler et al. 2003) datasets, while the TRMM 3B42, which is a 3 hourly dataset, is used to validate the diurnal cycle of precipitation. The TRMM 3B42 is a high resolution (0.25° × 0.25°) near-global dataset, produced by optimally merging multi-satellite estimates to produce 3-hourly precipitation fields. TRMM 3B43 is a monthly dataset produced by combining the 3-hourly merged high-quality/infrared (IR) estimates from 3B42 for the calendar month with monthly accumulated and analyzed rain gauge data. The TRMM 3B42 data were chosen over the radar-only TRMM 2A25, since the temporal sampling of precipitation from radar-only measurements is poor, with sampling frequencies sometimes less than once per day, especially towards the equator. Unless other estimates (e.g. from microwave or infrared data) are incorporated, only a long term climatology of the diurnal cycle can be produced such as in Biasutti et al. (2011), who noted that the seasonality of the diurnal cycle was a strong caveat of their own analysis. Seasonal rainfall statistics are also evaluated using 2 additional data sets. GPCP is a relatively coarse gridded (2.5° × 2.5°) monthly rainfall analysis produced by merging gauge measurements and satellite estimates of rainfall (Adler et al. 2003). The CRU dataset includes 0.5° resolution gridded data derived from land stations for both surface air temperature and precipitation (Mitchell & Jones 2005). Finally, where appropriate, the model results are also compared with the corresponding fields from ERA-interim reanalysis (Simmons et al. 2007) that is used to drive the RegCM4.

2.2. Model setup and analysis

The climate model used here is the latest version of the International Centre for Theoretical Physics (ICTP) RCM (RegCM4), which is described by Giorgi et al. (2012). RegCM4 has a hydrostatic dynamical core (Grell et al. 1994) on an Arakawa B horizontal grid, and a terrain following sigma coordinates. In the present setup, the model employs the radiation scheme of CCM3 (Kiehl et al. 1996), a modified version of the planetary boundary layer scheme of Holtslag et al. (1990), the resolvable scale precipitation scheme of Pal et al. (2000) and a mixed convection scheme using the MIT convection parameterization of Emanuel (1991) over ocean areas and the scheme of (Grell 1993) over land areas. This configuration has proven to be most effective over a number of CORDEX domains (Giorgi et al. 2012). The control (CTRL) model integration uses Biosphere-Atmosphere Transfer Scheme (BATS; Dickinson et al. 1993) to describe the land surface processes. This scheme has been used within the RegCM system for a number of years. In addition, 2 sensitivity studies are performed which impact the surface coupling to the atmosphere over the land and ocean, respectively. (1) In the first experiment, the BATS land surface scheme is replaced by the Community Land Model (CLM) version 3.5 (Oleson et al. 2008), which is a more advanced scheme including a more comprehensive description of the surface energy and water cycles. (2) The second experiment focuses on the ocean with the implementation of a diurnal cycle SST scheme of Zeng & Beljaars (2005), which calculates the diurnal evolution of the ocean skin temperature based on energy budget considerations. The 2 sensitivity experiments are hereafter referred to as the CLM and DCSST integrations, respectively.

Six hourly fields from ERA-interim reanalysis (Simmons et al. 2007) are used to provide the initial condi-
tions and lateral boundary forcing for the regional model in all experiments. The ERA-interim reanalysis was interpolated onto a 1.5° × 1.5° horizontal grid resolution and 37 levels in the vertical. SSTs are obtained from the National Oceanic and Atmospheric Administration (NOAA) optimal interpolation weekly SST data (Reynolds et al. 2002). Ten grid points in each direction are allocated for each lateral buffer zone where an exponential nudging is used to combine the model fields and the boundary conditions (Giorgi et al. 1993). The model domain follows the CORDEX specifications and covers a large region encompassing Central America (15° to 145°W and 25°S to 42°N, Fig. 1), however the analysis focuses on the inner regions of the domain covering the Central American continental area and adjacent ocean regions (0° to 35°N, 30° to 130°W). Following CORDEX, the model horizontal grid spacing is 50 km with 18 sigma-pressure levels used in the vertical as in the standard configuration of the model. All the integrations start at 00:00 h UTC on 1 January 1997 and run continuously for 6 yr until 1 January 2003. The first year is excluded from the analysis to allow for model spin-up.

A harmonic analysis method is used to measure the coherence of the diurnal cycle between the simulations and observations (Wallace 1975, Dai 2001, Collier & Bowman 2004). The harmonic analysis of a time series is the decomposition of a periodic function into a sum of trigonometric functions as shown:

\[
R(t_k) = \hat{R} + \sum_{i=1}^{N/2} A_i \cos(2\pi t_k/24) - \phi_i
\]

where \( t_k = (00, 03, 06, 09, 12, 15, 18, 21) \), the time in hours (UTC), \( A_i \) and \( \phi_i \) represent the amplitude and the phase of the \( i \)th harmonic respectively. \( R(t_k) \) is the estimated precipitation for each 3 h interval. \( \hat{R} \) is the average of the eight 3-hourly samples. The portion of variance explained by the \( i \)th harmonic wave is determined by \( A_i^2/\sigma^2 \), where \( \sigma \) is the standard deviation of the eight 3-hourly mean samples. First, a long term (seasonal) mean for the 8 time steps is computed. Then, a Fourier analysis is used to compute the amplitude and phase of the first harmonic (diurnal cycle). The phase of the first harmonic indicates the timing of maximum precipitation in the diurnal cycle.

3. RESULTS AND DISCUSSION

3.1. Seasonal features

3.1.1. Precipitation

The comparison of mean precipitation in the 3 RegCM4 experiments with observations confirms that all simulations produce the observed precipitation maximum over the eastern Pacific and Atlantic ITCZ regions, and over northern South America (Fig. 2). While the observed precipitation patterns are well reproduced, both the CTRL and DCSST experiments overestimate precipitation amounts (with up to 8 to 10 mm d\(^{-1}\) for the CTRL and 6 to 8 mm d\(^{-1}\) for the DCSST) off the west coast of Central America and over the north-northeastern coasts of South America. Conversely, precipitation is somewhat underestimated, by up to 2 mm d\(^{-1}\), over areas of the Gulf of Mexico, Cuba, Florida, the Yucatan Peninsula and the eastern coasts of Central America and western Caribbean islands.

A prominent and somewhat surprising finding of Fig. 2 is the strong response of the simulated precipitation over the equatorial Pacific and Atlantic oceanic regions in the CLM simulation, in which the biases over these regions are significantly reduced compared to the CTRL run, especially off the west coast of Central America and the northeastern coasts of South America.

In general, CLM tends to inhibit precipitation over land when compared to BATS, both over Central and South America. This is consistent with previous tests of CLM within the RegCM framework. For example, Steiner et al. (2009) found that the use of CLM reduced simulated precipitation over the West Africa monsoon region and attributed this to soil–precipitation feedbacks. A more recent study with RegCM and CLM3.5 with a domain over the USA also showed a warm/dry bias over the centre of the country, linked to soil moisture (Tawfik & Steiner 2011).
Concerning the use of the diurnal cycle SST, its main effect, although not large, is to increase precipitation along the ITCZ core, both over the Pacific and the Atlantic. The simulated diurnal cycle of SST is of the order of several tenths of a degree, and this is evidently sufficient to trigger increased convection particularly during the day. We also note the substantial differences between the TRMM and GPCP observations during this 5 yr averaging period, with TRMM indicating a more sharply defined and more intense ITCZ core. In addition to the use of different data sources and underlying retrieval algorithms, these differences could also be related to the order-of-magnitude contrast in the resolution of the 2 products. This observational uncertainty makes an unambiguous assessment of the model performance more difficult.

3.1.2. Annual cycle and MSD

The area-averaged (land-only grid points) simulated and observed annual cycle of rainfall for the Central America and Mexico subregions (shown in Fig. 1) is presented in Fig. 3. Over Central America, the observations show a bimodal annual cycle with peak precipitation in June and September. Precipitation shows a relative minimum in July and August, during the MSD phenomenon. This bimodal pattern, which is more pronounced in the TRMM than the CRU data, is reproduced well by the RegCM4_CTRL and RegCM4_DCSST simulations, although there is a difference in the simulated and observed amount of peak precipitation in September. The CLM simulation captures the September peak but misses the
June one and, more generally, shows an underestimation of precipitation over land.

The reduction in precipitation in the CLM simulation appears to lead to an overall reduction in diabatic heating, as diagnosed by vertically averaged temperature differences between the CLM and CTRL simulations (not shown). Consistent with this negative heating anomaly in the CLM simulation relative to the CTRL run, there is a Gill-type response (Gill 1980) with overall easterly surface— and westerly upper level—winds to the west of the negative heating, and a pair of upper-level cyclones in the subtropics. Following the steady, linearized vorticity equation (Gill 1980), a negative heating anomaly and associated column shrinking is balanced by equatorward motion to the west of the forcing. A low-level anticyclone and an upper-level cyclone to the west of the cooling comprise the baroclinic structure that characterizes a Gill response to a negative heating anomaly. Particularly in June, the location of these circulation anomalies (Fig. 4) appears to favor further precipitation reduction, with maximum anticyclonic anomaly at 925 hPa over the west coast of Mexico. Associated with this anticyclonic anomaly, there is enhanced vertically integrated moisture divergence over Central America and off its western coasts, and enhanced moisture convergence further south near 5°N over the eastern tropical Pacific. Therefore, particularly in June, the reduction in precipitation in the CLM run initiated by the land surface response feeds back to the regional circulation, further reducing precipitation, even over adjacent oceanic regions. The response over the ocean regions may be interpreted as a Gill response to the negative heating anomaly over the tropical American land regions.

Over the Mexico sub-domain, the bimodal nature of precipitation is not present, since part of this region is affected by the North American monsoon, which has a precipitation annual cycle that is out of phase with the MSD (Adams & Comrie 1997, Small et al. 2007). TRMM places a precipitation maximum in September, while CRU places it in August. The CLM simulation more closely follows the observed annual cycle of CRU, although the maximum in the model occurs in July. As TRMM, the CTRL run correctly shows a precipitation maximum in September, but overestimates precipitation. Overall, the annual cycle comparison suggests that there is no single configuration of RegCM4 that is best for both regions, although the CTRL configuration appears to better reproduce the observed intra-seasonal fluctuations (i.e. the MSD) in the annual cycle.

As a measure of the MSD, Fig. 5 shows the spatial pattern in precipitation difference between July/August and June/September (monthly data used) following Small et al. 2007) for TRMM, and the CTRL, DCSST and CLM runs. The blue color in the plots indicates the precipitation deficit associated with the MSD. The TRMM data exhibit 2 main areas with MSD. One is located over southern Mexico, western Central America, and the Gulf of Tehuantepec, while a less pronounced region is found over the Gulf of Mexico and the northern Caribbean. Conversely, a positive precipitation anomaly is found over the northern equatorial Pacific and over the western coasts of Mexico near the core region of the North American monsoon system, which, as noted previously, has a rainy season that is out of phase with the MSD.

All the RegCM4 simulations are able to reproduce the negative precipitation anomaly during the MSD
over Central America, and the positive anomaly over the north-equatorial Pacific and over the west coast of Mexico. The model, however, does not adequately capture the presence of the MSD when using the CLM scheme. Overall, the use of the BATS scheme appears to produce a better agreement with observations in terms of the intra-seasonal anomalies analyzed in Fig. 5.

The MSD precipitation anomaly off the western coastal regions of Central America is associated with a low level (at 925 hPa) anticyclonic circulation anomaly over the region, i.e. a northerly anomaly over the Gulf of Mexico, easterly anomaly over the Caribbean and southerly anomaly over eastern Pacific along the west coast of the continent (Fig. 6). The easterly anomaly over the Caribbean suggests that the CLLJ is enhanced during the MSD. This relationship between rainfall and the CLLJ is in agreement with the results of Wang (2007) who found a negative correlation between rainfall over Central America and CLLJ at the inter-annual time scale. All RegCM4 simulations show a similar circulation anomaly pattern compared to the reanalysis, although with a reduced magnitude. The inter comparison of the 3 simulations shows that the anticyclonic anomaly (northerly over the Caribbean and southerly over the eastern Pacific and stronger CLLJ) is better reproduced by the RegCM CTRL and DCSST run, with the CLM run showing a much weaker anomaly over the Caribbean. The weaker MSD in the CLM run may be related to the Gill-type response in the CLM run noted above, which serves to reduce precipitation in the first peak in June, which would reduce differences between July-August and June-September precipitation.

3.1.3. Temperature

In all 3 simulations the model is able to capture the spatial pattern of surface air temperature as driven by the local topography (Fig. 7). However, several differences between the model integrations, reanalysis, and observations are found. Differences between reanalysis products and surface observations can derive from errors in reanalyses systems (from model and/or observational error), with Mooney et al. (2011) recently documenting pointwise root mean square errors in ERA-Interim ranging from 0.2 to 0.8 K for a series of mid-latitude weather stations, but also from the fact that the interpolated/gridded observational dataset also contains observational errors, especially in the parts of the region where station data is relatively sparse. First, it is noted that ERA-Interim produces lower temperatures than CRU over the Amazon Basin and the Sierra Madre Occidental of Mexico. The CTRL run also shows a similar cold bias distribution, in the order of 1 to 3°C, largely agreeing with the ERA-Interim reanalysis. The cold bias over mountainous regions is at least partially related to the well known valley bias in
station-based observational networks, and has also been noted in evaluations of other RCMs over Central American domains (Karmalkar et al. 2008). Over the Amazon Basin the density of observing stations is low, making the CRU data rather uncertain. The DCSST simulation has a limited impact on temperature, mostly in the order of a few tenths of a degree. A greater sensitivity is found in the CLM run; the simulated temperatures are higher over most parts of Mexico and northern South America compared to the CTRL run, and this helps in reducing the cold model bias by 1 to 1.4°C compared to the CRU data.

3.1.4. Low level circulation

The low level circulation features (925 hPa) in the ERA-Interim reanalysis and the 3 RegCM4 simulations are shown in Fig. 8. The dominant feature at this level is the easterly jet core over the Atlantic which is associated with the subtropical Atlantic High. The easterlies reach their maximum (up to 11 m s⁻¹, i.e. the CLLJ; Amador 1998) between the Caribbean highlands and South America. These easterlies turn into a southeasterly flow over the Gulf of Mexico, bringing moisture to the southeastern US (Wang & Lee 2007, Muñoz et al. 2008, Cook & Vizy 2010) via the Great Plains low level jet. On the Pacific side, the southerly cross-equatorial flow from the southern Pacific, the northerly flow from the northern Pacific (which is associated with the Pacific subtropical high) and the easterly flow from the Atlantic converge off the west coast of Central America. The easterly flow from the Atlantic and the southwesterlies from the Pacific also converge over the northern part of South America. This indicates that the main areas of moisture convergence and precipitation during the northern hemisphere summer are off the west coast of Central America and northern South America.

The location of CLLJ is simulated reasonably well by the 3 model versions, although the strength of the simulated jet is weaker compared to the ERA-Interim reanalysis, and it is located slightly too far south, with the core around 12° N. A vertical cross section of the zonal winds at 75° W (not shown) reveals that the CLLJ is slightly stronger in the CLM simulation than...
in the CTRL run, perhaps contributing to the lower rainfall amounts over Central America in the CLM simulation. This is consistent with Wang (2007), who found (for August and September) a significant negative correlation between the CLLJ and rainfall over Central America, the eastern Caribbean Sea, and northwestern South America. The slight strengthening of the jet in the CLM run may be related to the higher surface temperatures over South America, since heating can strengthen the meridional pressure gradient over the Caribbean and hence the jet (Muñoz et al. 2008, Cook & Vizy 2010).

3.1.5. Surface energy budget

Since the effect of the DCSST run is small, the analysis of surface fluxes concentrates on the comparison of the CTRL and CLM runs. As shown in Fig. 2, and discussed in Section 3.1, precipitation is substantially reduced in the CLM run compared to the CTRL run. This change can be largely attributed to land-atmosphere feedbacks, as the latent heat flux in the CLM simulation is much lower than in the CTRL simulation (Fig. 9d). Due to the lower precipitation and cloud cover, net shortwave radiation at the surface increases in the CLM simulation, while net longwave radiation at the surface also increases due to higher surface temperatures.

CLM has a well-known dry bias in comparison to other land surface models, which has been partially attributed to its tendency to simulate mean global evapotranspiration with low contributions from transpiration and high contributions from soil and canopy evaporation (Oleson et al. 2008, Lawrence & Chase 2009). Reduced evaporation and drier soils can lead to a reduction in precipitation recycling (Schär et al. 1999) and a decrease of the efficiency of precipitation processes operating in a region (e.g. Betts et al. 1996). Similar feedbacks have been demonstrated in previous RegCM4_CLM couplings (Steiner et al. 2009). This decrease in surface moisture availability, and hence precipitation, is substantial enough to modify the circulation of the CLM simulation in a way that

Fig. 6. Mid-summer drought low-level (925 hPa) wind speed, 1998–2002. (a) ERA-Interim, (b) RegCM4_CTRL, (c) RegCM4_DCSST, and (d) RegCM4_CLM. Filled contours: magnitude of the wind
appears to enhance the overall dryness of the CLM experiment, as discussed in the previous subsection.

3.2. Diurnal cycle

3.2.1. Significance of the first harmonic

In order to ascertain how much the daily precipitation cycle is explained by the first harmonic, the percentage of the total daily variance explained by it is shown in Fig. 10. During JJAS, the diurnal cycle from the 1st harmonic accounts for >70% of the daily mean variance over most of the continental domain, suggesting that the first harmonic can be used as a first order measure of the characteristics of the diurnal precipitation cycle. In fact, a good agreement between the simulated and observed explained variance is found over land areas. In the TRMM dataset, a land–sea contrast can be observed in terms of the magnitude of this percentage, with lower values of percentage of variance explained by the first harmonics over ocean regions. Conversely, in the model the land–sea variation in the magnitude of variance explained by the first harmonic is not evident. This suggests that the model has a substantial diurnal cycle (compared to semi-diurnal and higher order harmonics) over the oceans, whereas in TRMM semi-diurnal
and higher order harmonics have a larger share in shaping the diurnal cycle over the ocean.

3.2.2. Amplitude of the first harmonic

Fig. 11 shows the JJAS seasonal mean normalized amplitude of the first harmonic of the diurnal cycle. The diurnal amplitude is generally larger over land than ocean areas in both TRMM and the RegCM4 experiments, consistent with previous studies using observations and models (e.g. Wallace 1975, Yang & Slingo 2001, Collier & Bowman 2004). Comparing the RegCM4 simulations and TRMM data, the diurnal amplitude is generally smaller in all RegCM4 simulation than in the TRMM data, except over the eastern coast of Central America. However, the CLM run appears to produce larger diurnal amplitudes than the CTRL run. This underestimation of the diurnal cycle amplitude by models is not common, especially in general circulation models (GCMs). In fact many GCM’s tend to overestimate the amplitude of the precipitation diurnal cycle (e.g. Collier & Bowman 2004).

3.2.3. Phase of the first harmonic

Fig. 12 shows the spatial distribution of the time of the peak diurnal precipitation from the TRMM observations and RegCM4 simulations. The time of peak precipitation is converted from UTC to local solar time by subtracting lon/15 from UTC, where lon is a 15 degree wide window centered at 15°W, 30°W, 45°W, ..., 135°W longitudes. Observed precipitation peaks in the early morning or early afternoon over most ocean areas and in the evening over most land areas. There are significant spatial gradients in the peak precipitation hours. For example, precipitation peaks in early morning along the coastal oceanic regions of Central America and later in the morning or early afternoon further from the coasts into the Gulf of Mexico and the equatorial ocean areas. The model reproduces some of these patterns, such as the
earlier peak times over ocean than over land and the coast-to-open sea gradients. In general, however, the model produces peak precipitation times earlier than observed. Over land, the peak precipitation time in the simulations is around 15:00 to 17:00 h, while the observed values are mostly around 20:00 h or even later.

The discrepancy between TRMM and model simulations in the timing of maximum precipitation is partly due to the fact that the TRMM data also include precipitation estimates from the IR spectrum. Lin et al. (2000) have shown that the timing between the maximum surface precipitation and the minimum outgoing longwave radiation or coldest cloud top surface (which is the basis of IR estimates in TRMM 3B42) can have a lag time of 2 to 3 h. In addition, Nesbitt & Zipser (2003) also found that TRMM estimates using radar and microwave imagers place the precipitation maxima between 17:00 and 19:00 h over Mexico, and suggested that the precipitation estimates from radar are superior to those based on IR or microwave brightness temperature. However it should be noted that the temporal sampling of precipitation from radar measurements alone (such as TRMM 2A25) is poor because some points sample at a frequency of less than once
per day, especially towards the equator. Therefore in order to obtain a time series of high temporal resolution for studying the diurnal cycle, some other estimates should be incorporated (e.g. from microwave or IR data) such as in TRMM 3B42; otherwise, only a climatology of the diurnal cycle can be produced (Biasutti et al. 2011). For example, Biasutti et al. (2011) generated a high-resolution climatology of 3-hourly gridded precipitation for Central America by combining the precipitation estimate in winter and summer with the assumption that in the tropics the characteristics of the diurnal cycle of convection remain similar. However, the phase of the diurnal harmonic in Yang & Slingo (2001) shows a significant difference between winter and summer, especially over Mexico, where the winter precipitation maxima occur around midnight.

Over open ocean areas, on the other hand, the simulated precipitation peaks in the early morning, while in the TRMM the peak is in the late morning or early afternoon. The best agreement between observations and simulations occurs over the Gulf of Mexico and the coastal areas. The peak precipitation time is also related to the convection trigger. For example, the Grell scheme — which is activated when sufficient buoyant energy lifts parcels above the free convection level — tends to generate precipitation in the early afternoon over land (Dai et al. 1999). Intercomparison of our simulations suggests that the use of CLM produces a slightly earlier precipitation peak than BATS, and the use of the DCSST scheme produces a later peak, than the standard configuration; however, these effects are not pronounced.

The diurnal cycle of precipitation can be determined by both local and non-local effects, with convection propagating into the study area in mesoscale-organized systems (Yang & Smith 2006). Propagating systems are revealed in a time–longitude cross section of JJAS seasonal mean precipitation anomalies (for anomalies greater than the daily mean) around Central America, averaged between 10° N and 20° N for the TRMM and RegCM4 simulations (Fig. 13). In Fig. 13 the diurnal cycle is repeated twice for clarity. The TRMM data shows that in this region mesoscale
Convective systems tend to initiate at 82.5° W at noon, then propagate westward, and reach 95° W at 03:00 h with an average speed of $\approx 25$ m s$^{-1}$.

All 3 RegCM4 simulations reproduce this westward propagation, even though there is a phase shift and a reduced intensity, especially around 90° to 91.5° W. A break in the propagation of rainfall anomalies in the simulation is observed during the early evenings and this is more pronounced in the CLM run. This break in the propagation rainfall anomaly at about 91°W can be linked to the Yucatan Peninsula and Guatemalan highlands, since the breaks are aligned with a secondary peak in topographic height, which is calculated by averaging the topography over the 10° to 20° N latitude band (Fig. 13).

### 4. CONCLUSIONS

In this paper we present a sensitivity study with the newest version of the RegCM regional climate modeling system, RegCM4 (Giorgi et al. 2012) run over the Central America CORDEX domain. This region is characterized by complex climatic features resulting from large-scale flow interactions and local processes related to topography and land–sea contrasts, and to date only few regional modeling studies have focused on it. On the other hand, it is a region identified as one of the most prominent climate change hot-spots (Giorgi 2006). It is thus important to test RCMs over this domain and to identify major model deficiencies and areas in need of improvement. Specifically, the present study tests the sensitivity of the model to land surface (CLM vs. BATS) and ocean SST (Standard vs. diurnal SST) parameterizations with a focus on the model representation of precipitation from the seasonal to the diurnal scale.

Overall, the model reproduces the spatial and seasonal patterns of precipitation over the region. The model captures the occurrence of the MSD over Central America, i.e. a break in the precipitation amount during July and August, followed by a recovery in September. Regional circulation features are also reproduced, although the intensity of the Caribbean...
Low-Level Jet is underestimated, and it is located too far south.

The model shows a substantial sensitivity to the land surface scheme, through the interaction of the surface energy and water budgets with the overlying circulations and convection. In general, the use of CLM leads to reduced precipitation, associated with reduced evaporation and increased surface sensible heat flux. This leads to increased agreement with observations over some areas and decreased agreement over others. The choice of land surface scheme affects not only precipitation over land, but also over oceans through a teleconnection involving regional circulations. The response over the ocean regions may be interpreted as a Gill response to the negative heating anomaly over the tropical American land regions. Conversely, the effect of the diurnal SST scheme is relatively minor.

Concerning the diurnal cycle of precipitation, the model reproduced reasonably well the amplitude of the observed cycle and its land–ocean contrast, but showed a systematic bias in the phase of the cycle consisting of an earlier placement of the peak precipitation time in the model compared to observations. This bias was present when using both the BATS and CLM land surface modules, and is therefore probably related to other components of the model physics, most likely involving the deep convection schemes. More research is needed to improve this problem, which is common to many global and regional climate models.

As a summary assessment, the present version of RegCM4 appears to provide a good representation of the basic features of the monsoon climate of Central America, although some systematic biases do persist, particularly at sub-diurnal scales. Neither of the land surface schemes tested appears universally superior over this region, and in particular it is highlighted that the more complex CLM did not lead to a consistently improved model. Given the overall satisfactory performance of the model over this region, and keeping in view its main deficiencies at sub-daily scales, in future studies we plan to apply the RegCM4 modeling system to project climate changes over this region following the CORDEX protocol (Giorgi et al. 2009).

Fig. 12. Phase (in local solar time) of the first harmonic analysis for (a) TRMM, (b) RegCM4_CTRL, (c) RegCM4_DCSST, and (d) RegCM4_CLM
Fig. 13. Time–longitude cross-section of mean JJAS precipitation anomaly greater than the daily mean rainfall (averaged over 10°–20° N). (a) TRMM, (b) RegCM4_CTRL, (c) RegCM4_DCSST, (d) RegCM4_CLM. The shading at the top of each plot represents the scaled terrain height averaged between 10° and 20°N. LST: local solar time.

LITERATURE CITED


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