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**A High Resolution Coupled Ocean-Atmosphere Simulation of the Regional Climate
Over Central America**

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14 **Abstract**

15 This study analyzes a relatively high resolution (15km grid spacing), regional coupled
16 ocean-atmosphere simulation configured over Central America. The simulation is forced with
17 global atmospheric and oceanic reanalysis for a period of 25 years (1986-2010). The spatial
18 resolution and the time period of the Regional Climate Model (RCM) simulation are both
19 unprecedented for the region. The highlights of the RCM simulation include the verifiable
20 seasonal cycle of mesoscale features like the low level jets, the mid-summer drought and the
21 seasonal tropical cyclone activity both in the Pacific and in the Atlantic Oceans. Similarly, the
22 seasonal cycle of the robust surface ocean currents in the eastern Pacific and the Costa Rica Dome
23 is also well captured in the RCM simulation. The RCM simulation also resolves the seasonal cycle
24 of the Panama-Colombia Gyre, the Gulf of Papagayo and the Gulf of Tehuantepec Gyre. In many
25 instances we find the RCM improves upon the global reanalysis forcing the simulation, indicating
26 the potential value of dynamic downscaling. Furthermore, the co-evolving components of the
27 atmosphere and ocean in the RCM is an added benefit to the atmosphere only and ocean only
28 global reanalysis forcing the simulation. However, the model displays significant biases that
29 manifest in precipitation, precipitable water, SST and winds which could potentially be improved.
30

1 Introduction

The Central American Isthmus (CAI) is by far one of the most challenging areas for climate modeling. It is a narrow strip of land, which is approximately 100km wide, beset with narrow range of mountains and surrounded by oceans along its eastern and western coast (Fig. 1). The isthmus is periodically affected by natural threats engendered by a prolonged season of Tropical Cyclone (TC) activity from both the Caribbean Sea and the eastern Pacific Ocean. For example, Hurricane Mitch of 1998, devastated Honduras with over 11,000 fatalities (Morris et al. 2002). Similarly, Afonso (2011) and Spencer and Urquhart (2018) indicate that there is significant human migration from Central American countries to the United States as an aftermath to landfalling TCs in the region. Besides TCs, the region is well known for many Low Level Jets (LLJs) that are caused by gaps in the mountain ranges of Central America. These LLJ's can sometimes produce low level vorticity, which in conjunction with either the Inter-Tropical Convergence Zone (ITCZ) and or the prevalent monsoon trough can lead to tropical cyclogenesis in the tropical eastern Pacific Ocean (Heather and Bourassa 2014).

Maldonado et al. (2018) suggest that apart from the intense, relatively infrequent events of landfalling TCs, the higher frequency meteorological events of lower intensity can also cause severe impacts in the region. For example, the synoptic disturbances emanating from the ITCZ affect the Pacific side of Central America (Hidalgo et al. 2015; Quiros-Badilla and Hidalgo-Leon, 2016), the rainfall production from cold surges sometimes produce extreme precipitation in the dry winter and early summer climate of the north and Caribbean coast of Costa Rica and Honduras (Schultz et al. 1998; Retana 2012), the westward propagation of tropical disturbances in the summer (Amador et al. 2016), and the mid-summer drought (Magana et al. 1999).

The eastern Pacific Ocean in the west and the Gulf of Mexico and the Caribbean Sea to the east of the CAI is part of the Intra-Americas Seas (IAS). The IAS is host to the second largest body of warm water ($\geq 28.5^{\circ}\text{C}$) on Earth and also hosts the second largest diabatic heating center of the tropics during the boreal summer (Wang and Enfield 2001). Overlying the IAS, on the Atlantic side is the North Atlantic Subtropical High (NASH), which display a distinct seasonal cycle (Davis et al. 1996). The equatorward flank of the NASH comprise of the easterly trades of the Atlantic, which has a strong bearing on the moisture flux into the CAI as these trade winds accelerate in the Caribbean Sea to form the Caribbean LLJ (CLLJ; Amador and Magana 1999; Poveda and Mesa 1999; Wang 2007; Gimeno et al. 2012; Misra et al. 2014). Similarly, in the Pacific side, resides the North Pacific subtropical high, which also carries the north easterly trades but away from the CAI. Xie et al. (2008) suggest that variations in the tropical Atlantic Ocean can affect tropical Pacific by way of the atmospheric bridge across Central America. For example, the cold phase of the Atlantic Multidecadal Oscillation is observed to reduce the amplitude of the seasonal cycle over the equatorial Pacific (Timmermann et al. 2007). Similarly, Xie et al. (2008) indicate that cooling of the North Atlantic SST suppresses atmospheric convection over Central America and tropical eastern Pacific Ocean. Likewise, the mid-summer drought over Central America observed in July-August amidst their wet season before it recovers in September is a result of the variations of NASH initiated by anomalies in the north tropical Atlantic Ocean (Mapes et al. 2005; Xie et al. 2008). More recently, Corrales-Suastegui (2021) suggest that the differential warming between the tropical Atlantic and eastern Pacific Ocean can lead to further severity of the mid-summer drought in a future climate.

The Costa-Rica Dome located off the west coast of Central America is the shoaling of a generally strong and shallow thermocline in the region. It is a ‘permanent’ feature that is centered

around 9°N and 90°W, about 300 km off the Gulf of Papagayo between Costa Rica and Nicaragua (Wyrtki 1964). As a result of this ridging of the thermocline, the Costa Rica Dome is a region of high primary productivity as nutrients are brought up to the surface by wind mixing and upwelling (King 1986). However, there is significant variability of this dome both with its location and magnitude as a result of the wind variations associated with the ITCZ and the surface ocean currents (Fiedler 2002). Seasonally, the Costa Rica Dome is shallower in the boreal summer-fall seasons when the ITCZ is over the dome and local winds are weak. In the boreal winter season, the Costa Rica Dome is deeper when the ITCZ is south of the dome and the northeasterly trades are stronger. However, it may be noted that like any other tropical thermal dome, the Costa Rica Dome is associated with the cyclonic turning of zonal surface currents following geostrophic balance. In the case of the Costa Rica Dome, these zonal surface currents in the east Pacific are the North Equatorial Counter Current to the south and the North Equatorial Current to the north of the dome.

Given this rich amalgam of atmospheric and oceanic features of Central America, we are motivated to conduct a regional coupled ocean-atmosphere integration at an unprecedented spatial resolution of 15km grid spacing to assess the fidelity of the model in simulating distinct atmospheric and oceanic features. In contrast, recent regional downscaling experiments over Central America are coarser in spatial resolution and are downscaling just the atmospheric component (e.g., Cavazos et al. 2019; Corrales-Suastegui et al. 2020). A coupled regional climate model integration over this region was last conducted in Xie et al. (2008), which was over an 8 year period at 0.5° grid spacing. In this study, the regional climate model integration is conducted by forcing at the lateral boundaries with global atmospheric and oceanic reanalysis for a period of 25 years from 1986-2010. In the following section we describe the regional climate model. In Section

3, we discuss the details of the model integration along with the validation datasets used in the study. We present the results in Section 4 followed by concluding remarks in Section 5.

2 Model Description

The model used for this study is the Regional Spectral Model-Regional Ocean Modeling System (RSM-ROMS). RSM-ROMS is a regional, coupled ocean-atmosphere model that has been extensively used for regional climate modeling studies (Li et al. 2014; Li and Misra 2014; Ham et al. 2016; Misra et al. 2018). The atmospheric component of RSM-ROMS is RSM that was first introduced in Juang and Kanamitsu (1994). RSM uses the spectral method to compute advective derivatives using sine and cosine functions in two dimensions with wall boundary conditions (Tatsumi 1986). RSM uses the semi-implicit time integration scheme. The RSM has 28 terrain following sigma levels, with irregular spacing in the vertical that is identical to the NCEP-DOE reanalysis (R2; Kanamitsu et al. 2002), and the top of the atmosphere in RSM is at approximately at 2 hPa. The physics package used in RSM for this study is outlined in Table 1. The scale selective bias correction (Kanamitsu and Kanamitsu 2007, Kanamitsu et al. 2010) is used to prevent synoptic scale drift during the integration of RSM. Additionally, we have also increased the width of the sponge zone from the 8 grid points around the lateral boundaries used in the previous studies (e.g., Misra et al. 2018) to 16 grid points in this study. This was done after extensive model integrations, which showed that RSM displayed the most skill in reproducing the observed climate at a sponge width of 16 grid points (not shown).

ROMS is the oceanic component of RSM-ROMS. ROMS is a free surface, terrain following primitive equation ocean model (Haidvogel et al. 2000; Shchepetkin and McWilliams 2005). The equations are solved using the split explicit time-stepping scheme. The primitive equations are discretized in the vertical by using stretched, terrain following coordinates with 30

levels. The ROMS contain several subgrid scale parameterizations, which include local closure schemes based on the level 2.5 turbulent kinetic energy equations (Mellor and Yamada, 1982) and generic length-scale parameterization following Umlauf and Burchard (2003). For this study, we used the second order biharmonic horizontal diffusion in ROMS (Ezer et al. 2002). The nonlocal closure scheme is based on the K profile, boundary layer formulation developed by Large et al. (1994). The K profile scheme has been expanded to include both surface and bottom oceanic boundary layers.

RSM and ROMS are coupled with a coupling interval of 1-hour. The resolution of RSM and ROMS for this study is set at 15km and is identical for both model components. Therefore, no flux coupler is used, and the atmospheric fluxes and SST are exchanged directly without any interpolation between the two model components at every hour of the model integration. Furthermore, there is no flux correction used in the model integration of RSM-ROMS.

3 Design of Experiment

The model domain used for this study is shown in Fig. 1. As mentioned earlier, the horizontal grid spacing of RSM-ROMS is 15km with the center of the grid at 85°W and 10°N with 149 and 178 grid points in the zonal and meridional directions, respectively. The model domain extends from 3.49°S to 23.29°N and 68.01°W to 96.37°W, which includes the sponge zone. The orography used in RSM is interpolated to the domain grid from the Global 30 Arc-Second Elevation (Danielson and Gesch 2010), which equates to a resolution of approximately 1km (Webster et al. 2003). Similarly, the ocean bathymetry used in ROMS is from 2-minute gridded global relief data from NOAA National Geophysical Data Center (2006). The orography over land and the bathymetry in the ocean for the model domain and resolution is shown in Fig.

1. Over land, the post-processed topography used in RSM-ROMS does not resolve the Cordillera Central Mountain ranges of Costa Rica and Panama and underestimates the elevation of the Sierra Madre range over southern Mexico and Guatemala. The orography is spectrally filtered to reduce the grid scale numerical noise in RSM-ROMS, which we hope to rectify in our future studies over this region.

The initial and lateral boundary conditions of the atmosphere are supplied to RSM from R2. The lateral boundary conditions of RSM are provided at intervals of 6 hours. The ocean initial and lateral boundary conditions are provided from Simple Ocean Data Assimilation version 2. 2. 4 (SODA; Carton and Giese 2008). RSM-ROMS is integrated from 1st January 1986 through 31st December 2010.

The model verification and analysis are conducted in a domain that emphasizes the CAI. Therefore, the regional domain over which the model results are analyzed is limited to 0°-22°N and 75°W-95°W. In order to verify the RSM-ROMS integration we make use of the daily rainfall data from the Integrated Multi-satellitE Retrievals for Global Precipitation Mission version 6 (IMERG; Huffman et al. 2019). This rainfall data is available globally at 0.1° grid spacing from 01 June 2000 to the present. Although, this data set does not overlap with the integration period of RSM-ROMS (1986-2010), it still provides us a robust 20 year climatology covering both land and the oceans at a resolution that is comparable to RSM-ROMS. In addition, we use the Climate Prediction Center (CPC) daily rainfall data following Xie et al. 2007 and Chen et al. 2008. This rainfall data is available only over land and from 1979 to the present (which overlaps with the integration period of RSM-ROMS) and is at 0.5° grid spacing. We also make use of the total precipitable water from the NASA Water Vapor Project (NVAP; [https://eosweb.larc.nasa.gov/sites/default/files/project/nvap/readme/ASDC NVAP Overvie](https://eosweb.larc.nasa.gov/sites/default/files/project/nvap/readme/ASDC_NVAP_Overview)

[w_2016.pdf](#); doi:10.5067/NVAP-M/NVAP_CLIMATE_Total-Precipitable-Water_L3.001) to verify the precipitable water from the RSM-ROMS integration. The NVAP dataset is available globally from 1988 to 2009 at 1° grid spacing. For the verification of the model simulated SST, the NOAA optimally interpolated SST dataset is used. This dataset is available from 1981 to the present at 0.25° grid spacing at daily interval. The Climate Forecast System Reanalysis (CFSR; Saha et al. 2010) data is used for the validation of the upper air simulations. CFSR is available globally at 0.5° grid spacing on 17 mandatory pressure levels. We also verify the upper ocean heat content measured as the depth of the 20°C isotherm and the wind stress from the model simulation against the SODA reanalysis data (Carton and Giese 2008). The best track data of TCs both in the Caribbean Sea and in the eastern Pacific Ocean from the model simulation are verified against HURDAT2 (Landsea and Franklin 2013). However, the tracks of the TCs are plotted from both these datasets at one-day interval as some of the variables required to track them (mean sea level pressure, winds at 850hPa, topography height, 400hPa temperature) following the tracking algorithm of Ullrich and Zarzycki (2017) were unfortunately stored at one day interval in RSM-ROMS.

4 Results

a) The Seasonal Cycle of Rainfall

The seasonal cycle of the observed rainfall shows distinct dry (Figs. 2a and b) and wet (Figs. 2c and d) halves of the year over the CAI, coinciding with the boreal winter/spring and summer/fall seasons, respectively. The Inter-Tropical Convergence Zone (ITCZ) over northeastern Pacific Ocean appears very strong in June-July-August (JJA; Fig. 2c), which also coincides with the seasonal peak of rainfall over the CAI. The corresponding seasonal cycle of the

rainfall from RSM-ROMS is shown in Figs. 2e-h and its systematic errors in Figs. 2i-l and Table 2. From these model figures and Table 2, we observe that although RSM-ROMS shows the seasonal peak of the ITCZ in the JJA season (Fig. 2g), it underestimates the seasonal ITCZ rainfall in the spring (Fig. 2j), summer (Fig. 2k), and fall (Fig. 2l) seasons while overestimating in the winter season (Fig. 2i). However, the seasonal variability in the latitudinal width of the near rain free zone over the eastern Pacific cold tongue region is nicely replicated in RSM-ROMS. Over the Caribbean Sea, RSM-ROMS systematically overestimates (Figs. 2i-k) except in the September-October-November (SON) season, when the model shows underestimation along the Caribbean coast of Central America (Fig. 2k; Table 2).

Over land, RSM-ROMS uniformly overestimates across CAI in the December-January-February (DJF) season (Figs. 2e and i, and Table 2). The systematic errors in DJF over Honduras and Nicaragua and over northwestern Colombia are however significantly larger (Figs. 2e and i). In the March-April-May (MAM) season, the systematic errors of rainfall from RSM-ROMS over the CAI is significantly reduced (Figs. 2f and j, and Table 2). The systematic seasonal mean JJA rainfall errors of RSM-ROMS over the CAI are underestimated north of Costa Rica while over Panama and Costa Rica it is highly overestimated (Figs. 2g and k, and Table 2). However, in the SON season, the systematic errors of rainfall from RSM-ROMS become much larger compared to JJA with negative bias (underestimation) north of Panama and positive bias (overestimation) south of Costa Rica (Figs. 2h and i, Table 2).

b) The Seasonal Cycle of Precipitable Water

The observed seasonal cycle of the precipitable water (Figs. 3a-d) closely follows that of rainfall (Figs. 2a-d), which is well depicted in the RSM-ROMS simulation (Figs. 3e-h). The north-south gradient of precipitable water across the domain throughout the year is quite similar in both

observations (Figs. 3a-d) and RSM-ROMS (Figs. 3e-h). Furthermore, the maxima of the precipitable water in the ITCZ region of the eastern Pacific throughout the year is also captured in the RSM-ROMS simulation. However, over the tropical Pacific the bias in RSM-ROMS is negative throughout the year (Figs. 3i-l; Table 3). A widespread dry bias of the atmospheric column develops across the domain in JJA (Fig. 3k; Table 3) and SON (Fig. 3l; Table 3) seasons.

c) The Seasonal Cycle of SST

The observed SST in Figs. 4a-d show a robust seasonal cycle in both the eastern Pacific Ocean and in the Atlantic Ocean. For example, the coastal oceans of Central America both along the Pacific and the Atlantic sides exhibit coldest SST in DJF (Fig. 4a) and warmest SST in JJA (Fig. 4c). Furthermore, the cold tongue in the eastern equatorial Pacific Ocean widens and becomes colder in the JJA (Fig. 4c) and in the SON (Fig. 4d) seasons relative to DJF (Fig. 4a) and MAM (Fig. 4b) seasons. This seasonal cycle of the SST is reasonably captured in the RSM-ROMS simulation (Figs. 4e-h). The systematic errors of SST (Figs. 4i-l) indicate that the errors are least in DJF (Fig. 4i) and most severe in JJA (Fig. 4k) and SON (Fig. 4l) seasons. In general, the RSM-ROMS simulation is dominated by warm bias over both the Caribbean Sea and in the Pacific Ocean except in the cold tongue region where it exhibits a cold bias throughout the year.

An iconic feature of the region is the Costa Rica Dome, which shoals from west to east between the westward North Equatorial Current (NEC) to the north and the eastward North Equatorial Counter Current (NECC) to the south with the Costa Rican Counter Current (CRCC) along its eastern boundary (Fig. 5). This is observed in the SODA reanalysis monthly climatology from the depth of the 20°C isotherm (a proxy for the depth of the thermocline; Fiedler 2002), with the dome appearing in the Gulf of Papagayo in January-March, moving offshore during April-July, intensifying in July-November, before it begins to diminish in December-January (Fig. 5).

This seasonal cycle of the Costa Rica Dome is also depicted in RSM-ROMS (Fig. 6). However, RSM-ROMS displays a shallower thermocline depth and weaker surface ocean currents throughout the year over this region of the Pacific Ocean. Furthermore, SODA reanalysis displays an anticyclonic gyre between the Gulf of Papagayo and Gulf of Tehuantepec far more strongly in the winter months (Jan-Feb-Mar) than in RSM-ROMS. Perez et al. (2000) using two years of sea surface temperature data, remotely sensed altimetry, and scatterometer based wind stress datasets identified warm gyres forming in the winter months in this region. These gyres form as a result of the seasonal peak in strong off-shore low-level atmospheric gap winds in the Sierra Madre ranges that create anticyclonic wind stress curl producing warm core gyre in the region (Perez et al. 2000; Kessler 2006).

This bias of RSM-ROMS in the evolution of the Costa Rica Dome is also observed in the surface ocean currents. For example, the westward flowing SEC is much stronger and anchored closer to the equator in RSM-ROMS (Fig. 6) compared to SODA (Fig. 5) throughout the year. However, the seasonal cycle of eastward flowing NECC is far more reasonable in the RSM-ROMS simulation with its peak around 5°N in October and its nadir in March. The westward flowing NEC appears strong at ~10°N from January through April and then shifts further north and weakens slightly till October before the NEC begins to shift southward and strengthen (Fig. 5). The RSM-ROMS simulation in Fig. 6 shows the NEC at its peak around 10°N from January through April but then fades away till it begins to reappear in November. This bias has implications on the Costa Rica Counter Current in RSM-ROMS, as this current disappears from April to June before it begins to reappear as northwestward flow in July, which is unlike SODA reanalysis.

The upper ocean circulation in the Caribbean Sea is dominated by the Caribbean Current (Fig. 5). This is a warm, persistent surface current that flows along the northern coast of South

America and then along the eastern coast of Central America. Another surface ocean feature of note in the Caribbean Sea is the highly variable Panama-Colombia Gyre over the Colombian Basin. At the resolution of SODA reanalysis (at 0.5°), this gyre is barely resolved and is most apparent in July-August-September-October season (Fig. 5). The climatological surface ocean circulation in RSM-ROMS in Fig. 6 shows the persistent Caribbean Current in the northern Colombian Basin, western Cayman and Yucatan Basin and a rather strong cyclonic recirculation associated with Panama-Colombian Gyre in southwestern Caribbean Sea. The seasonal variation of the Panama-Colombian Gyre in RSM-ROMS with intense cyclone in the boreal winter months to multiple cyclones of similar size embedded in a larger but weaker cyclonic circulation in the Caribbean Sea is in qualitative agreement with previous observational findings (Andrade and Barton 2000; Fratantoni 2001). The depth of the thermocline in the Caribbean Sea is, however, consistently shallower in RSM-ROMS (Fig. 6) compared to that in SODA reanalysis (Fig. 5). The warm SST bias of RSM-ROMS both in the Caribbean Sea and eastern Pacific Ocean in Fig. 3, combined with the shallower thermocline depth (Fig. 6), especially in the summer and fall seasons suggests that the upper ocean is highly stratified. In other words, RSM-ROMS could benefit from stronger mixing in the upper ocean to reduce some of its bias both in the Caribbean Sea and in tropical Pacific Ocean.

c) The seasonal cycle of the low level jets

One of the grand challenges of modeling the climate of this region are the multitude of LLJs that have significant implication on the hydroclimate of the region. In this regard, the regional domain of the RSM-ROMS includes four prominent LLJs: The Caribbean Low Level Jet (CLLJ), the Panama Low Level Jet (PLLJ), the Papagayo Low Level Jet (PaLLJ), and the Tehuantepec Low Level Jet (TLLJ). The climatology of the 925hPa wind field created from both

the CFSR and the RSM-ROMS integration are shown in Fig. 7. We will discuss the winds from R2 shown in Fig. 7 in sub-section g. All of these LLJs are apparent in the CFSR annual mean climatology of the 925hPa wind from their distinct maxima in the wind speed (Fig. 7b). In contrast, in the annual mean climatology of the 925hPa winds of the RSM-ROMS simulation, amongst the four LLJs, the CLLJ and the PaLLJ are most prominent (Fig. 7c). The PLLJ in the RSM-ROMS appears subtly with a weaker wind maximum while the TLLJ does not appear as distinct (Fig. 7c). The annual mean systematic errors in Fig. 7d show that the RSM-ROMS clearly underestimate all of these LLJs relative to CFSR.

We observe in Figs. 7e-h that overall, the RSM-ROMS is able to fairly capture the seasonal cycle of the four LLJs. For example, the dual maximum of the CLLJ with primary and secondary peak in the boreal summer and winter is reasonably captured by the RSM-ROMS simulation (Fig. 7e). Although, RSM-ROMS underestimates the magnitude of the CLLJ throughout the year except in September (Fig. 7e). Likewise, the PLLJ shows a dual maximum, with primary peak in January and secondary peak in July, which the RSM-ROMS is able to simulate (Fig. 7f). PaLLJ in Fig. 7g, shows that it peaks in February in CFSR. RSM-ROMS displays significant month-to-month variability, although, the magnitude of PaLLJ is comparable to that of CFSR with slight underestimation of the seasonal peak in February. CFSR indicates that the TLLJ reaches its annual peak in November with a secondary maximum in July. The RSM-ROMS shows a similar feature with a primary peak in December and secondary peak in July (Fig. 7h). But the TLLJ is underestimated by RSM-ROMS relative to CFSR throughout the year (Fig. 7h).

d) Interannual variability

The interannual variations of precipitation in CAI is largely dictated by the variability of the LLJs (Duran-Quesada et al. 2017). Amongst the LLJs, the interannual variations of the CLLJ

during the JJA season dictate to a large extent the interannual variability of the summer hydroclimate over CAI (Wang et al. 2007, 2008; Misra et al. 2014). This variation of the CLLJ is associated with the variations of the Atlantic warm pool in the IAS (Wang et al. 2007, 2008), the variations of the meridional pressure gradient across the Caribbean Basin modulated by the tropical Pacific variability (Munoz et al. 2008) and the zonal gradient of SST between the tropical Atlantic and eastern Pacific Oceans (Enfield and Alfaro 1999). In the winter although, CLLJ has a seasonal peak, it does not serve as large of a source of moisture to CAI as in the summer, since the associated moisture flux vectors are comparatively weaker (Munoz et al. 2008). Furthermore, Maldonado et al. (2018) suggest that in the winter, moisture convergence in CAI is less important to dictate the hydroclimate than during the rest of the year. In Fig. 8a we show the CLLJ index¹ time series from CFSR and RSM-ROMS for the JJA season. The years of high and low CLLJ index are highlighted in Fig. 8a. These high and low index seasons are chosen if the CLLJ index exceeds or is less than the corresponding seasonal climatological mean by $0.75 \times$ standard deviation. The fractional coefficient of 0.75 was chosen to achieve reasonable sample of seasons for high and low index seasons to conduct a meaningful statistical significance test on the corresponding precipitation anomalies. The CLLJ index in JJA season shows far higher variability in RSM-ROMS than CFSR (Fig. 8a). The variations of the time series of the CLLJ index between CFSR and RSM-ROMS in Fig. 8a are understandably inconsistent with each other (meaning high and low index seasons are not coinciding between CFSR and RSM-ROMS) for a number of reasons. These include the notion that climate simulation is not an initial value problem, contributions of the bias (e.g., in SST, sea level pressure gradients that dictate the variability of low-level winds) and internal variability in RSM-ROMS. Furthermore, the bias and variations in

¹ defined as area averaged 925hPa wind magnitude over 12.5°N-17.5°N and 76°W-80°W

the R2 reanalysis that is forcing the RSM-ROMS may also contribute to the inconsistencies in the variations of the CLLJ between RSM-ROMS and CFSR in Fig. 8a.

The composite difference of JJA seasonal mean rainfall between the high and the low CLLJ index years shown in Fig. 8b from CPC rainfall indicate that most of CAI has a dry anomaly, consistent with the moisture flowing across the Isthmus into eastern Pacific to feed the ITCZ when the CLLJ is comparatively strong (Wang et al. 2007; Misra et al. 2014; Duran-Quesada et al. 2017; Corrales-Suastegui et al. 2020). It may be noted that we use land based CPC rainfall for this comparison because the time period of this rainfall dataset overlaps with that of CFSR and RSM-ROMS simulation. The RSM-ROMS simulation in Fig. 8c is able to show similar seasonal mean rainfall anomalies as the observations over Guatemala, southern Mexico, Nicaragua, Honduras, and Costa Rica (Fig. 8b). However, the positive rainfall anomalies in RSM-ROMS over Panama and northeast Colombia in Fig. 8c are unsubstantiated in the observations (Fig. 8b).

e) The Mid-Summer Drought

The iconic feature of the region is the mid-summer drought. This was diagnosed both in the observations and in RSM-ROMS when a relative minimum of rainfall in July and August, was observed between biannual peaks in June and September (Magana et al. 1999). We defined intensity of the mid-summer drought as the difference between the average of the biannual peak and the average of the minimum in the intervening two months (Figs. 9a and b). Their frequency of occurrence is identified in Figs. 9c and d as fraction of seasons with mid-summer drought in the 25 years (1986-2010).

The observations show that the Yucatan region and the Pacific side of the northern part of the CAI (e.g., western coasts of Guatemala, Nicaragua, and San Salvador) exhibit largest intensity (Fig. 9a) and highest frequency (Fig. 9c) of the mid-summer drought. South of Nicaragua, the

frequency of the mid-summer drought drops significantly (Fig. 9c) while the intensity over Costa Rica and Panama are moderate (Fig. 9a). The western coast of Colombia displays high frequency (Fig. 9c), but the intensity is low (Fig. 9a). The RSM-ROMS simulation shows some of these features, like high intensity in the Yucatan Peninsula, Costa Rica, and Panama (Fig. 9b). Furthermore, the meridional gradient in the frequency of the mid-summer drought between north and south of the Nicaraguan western coast in Fig. 9d is also apparent like the observations (Fig. 9c). However, RSM-ROMS clearly underestimates the frequency of the drought across CAI. The spatial pattern of the intensity of the drought in RSM-ROMS is reasonable relative to the observations, but the magnitude shows systematic bias like the underestimation over Yucatan Peninsula, Honduras, Guatemala and overestimation over Costa Rica and Panama.

f) Tropical Cyclone activity

The tropical cyclone activity is shown as track density (which is number of TCs per $1^\circ \times 1^\circ$ grid) in Fig. 10. As noted earlier, Fig. 10 plots the tropical cyclone fix at one day interval over the integration period of 1986-2010. RSM-ROMS (Fig. 10b) shows reasonable distribution of the TCs in comparison to HURDAT2 (Fig. 10a) both over the Caribbean Sea and the eastern Pacific with some notable biases. For example, like HURDAT2, RSM-ROMS shows greater number of TCs in the Caribbean Sea than over the eastern Pacific. Furthermore, RSM-ROMS does not generate TCs below 5°N in the eastern Pacific. The distribution of the TCs in the Caribbean Sea with a cluster of high density between the Cayman and Yucatan Basins and over the Nicaraguan Rise in HURDAT2 is comparatively much weaker in RSM-ROMS. In addition, the large density of the TCs in the Bay of Campeche and in the Colombian Basin in RSM-ROMS is unsupported by

HURDAT2. Likewise, the density of TCs in the eastern Pacific between 10° to 15°N in RSM-ROMS is less than in HURDAT2.

g) Comparison with R2

in the 25 years (1986-2010) The key benefit of dynamic downscaling is its ability to resolve or permit the fine scale features of the regional climate because of the superior horizontal resolution of the regional model. This feature of downscaling in our study has thus far been fairly demonstrated in the fidelity of RSM-ROMS in depicting some mesoscale features like the Costa Rica Dome, the Panama-Colombian and the Papagayo-Tehuantepec Gyres, the atmospheric LLJs, interannual and sub-seasonal variations of rainfall over Central America and the TC activity in the region. The other value of dynamic downscaling is its potential to improve upon the large-scale forcing used to force the regional model. Therefore, in our case it would be incumbent to compare the systematic errors of R2 reanalysis with those from RSM-ROMS. It may however be added, that RSM-ROMS by way of its coupled ocean-atmosphere framework already adds value to downscaling in contrast to either R2 or SODA reanalysis, which are exclusively global atmospheric or oceanic reanalysis, respectively.

In Figs. 11a-d, we show the systematic errors of seasonal mean rainfall from R2. In comparison to RSM-ROMS (Figs. 2i-l), the systematic errors of R2 (Figs. 11a-d) are higher over the whole domain in spring and summer (Table 2). In contrast, R2 rainfall is superior in terms of the mean error in winter (Table 2). An interesting point, however, is that the sign of the systematic errors of rainfall in the tropical eastern Pacific Ocean is nearly opposite between RSM-ROMS and R2 except in the winter season. This may be a result of the air-sea coupling in RSM-ROMS being more active in tropical eastern Pacific Ocean that allows for more degrees of freedom, with the ocean co-evolving with the atmosphere. On the other hand, the systematic errors display largely a

wet bias over the Caribbean Sea in both R2 and RSM-ROMS, which could be suggesting that the Caribbean region rainfall in RSM-ROMS is being largely dictated by the large-scale atmospheric conditions of R2.

Similarly, the comparison of the systematic errors of precipitable water between R2 (Figs. 11e-h) and RSM-ROMS (Figs. 3i-l) indicate that the latter is an improvement over R2 in the boreal winter and spring seasons across the domain (Table 3). However, in the boreal summer and fall, the systematic errors of precipitable water are nearly identical in R2 and RSM-ROMS. It is apparent from this comparison that the systematic errors in precipitable water are far more consistent in RSM-ROMS in relation to R2, which is in contrast to the systematic errors of rainfall. This is because specific humidity is one of the prescribed variables at the lateral boundaries of RSM-ROMS and hence the evolution of humidity is constrained to that imposed at the lateral boundaries. However, rainfall is not constrained by the lateral boundaries of RSM-ROMS.

The comparison of the low-level winds at 925hPa in Fig. 7 reveals that despite the coarse resolution of R2, the LLJs are reasonably depicted including their seasonal cycle. The RSM-ROMS in most instance replicates this seasonal cycle of the LLJs in R2 except in the case of PaLLJ (Fig. 7g), where the jet in RSM-ROMS is systematically stronger and shows far more variability than R2. We believe this is one aspect of the RSM-ROMS simulation that could be further improved with a more realistic prescription of the orography than used in this study (Fig. 1).

One of the important forcing to the ocean is the surface wind stress, which is demonstrably significant for the mesoscale variations of the surface ocean currents in the region (Fiedler 2002; Sheng and Tang 2003). The easterly wind stress in the Caribbean Sea and in the northeastern Pacific Ocean are dominant both in the reanalysis and in RSM-ROMS (Fig. 12). Similarly, southerly and northeasterly wind stress over the equatorial eastern Pacific Ocean and over the Colombian Basin, respectively is also apparent in Fig. 12. Similarly, the wind stress curl shows

regions of strong upwelling and downwelling in both oceans that have distinct seasonality associated with the Costa Rica Dome, ITCZ in the eastern Pacific and the LLJs (Fig. 12). There are however subtle differences between R2 (Figs. 12e-h) and RSM-ROMS (Figs. 12i-l). For example, the upwelling associated with Costa Rica Dome and Panama-Colombia Gyre is stronger in RSM-ROMS (and closer to CFSR in Figs. 12a-d) compared to R2 (Figs. 12e-h). However, the downwelling over the eastern equatorial Pacific Ocean is stronger in RSM-ROMS compared to R2 or CFSR (Fig. 12).

5 Conclusions

In this study we have examined the fidelity of an unprecedented multi-decadal regional coupled ocean-atmosphere model (RSM-ROMS) simulation over Central America at 15km grid spacing. The last such regional coupled ocean-model atmosphere modeling study over the region was conducted by Xie et al. (2007), which was conducted at 0.5° grid spacing over a 7-year period. The CAI, by its geography has a very rich amalgam of variations that feature both the surrounding oceans and over the land surface, which warrants the use of such relatively high resolution coupled ocean-atmosphere models. The current state-of-the-art global climate models at a nominal spatial resolution ranging between 2° to 1° are highly inadequate for either simulating or projecting future climate over the region.

The highlights of the RSM-ROMS simulation over the CAI region, forced with R2 global atmospheric reanalysis and SODA ocean reanalysis at the lateral boundaries are as follows:

- The seasonal cycle of precipitation over CAI and the surrounding oceans is reasonable. It had the least error in the spring followed by summer with the wet bias in the winter and the dry bias in the fall being significantly larger. Interestingly, seasonal precipitation bias

over the eastern Pacific Ocean in R2 is opposite to those in the RSM-ROMS simulation suggesting the potential role of air-sea coupling in the basin.

- The seasonal cycle of precipitable water clearly shows the meridional gradient of high precipitable water over the equatorial latitudes to lower precipitable water in the subtropics from winter through fall in both observations and RSM-ROMS simulation. However, the systematic errors in simulated precipitable water by RSM-ROMS is very similar to those in R2, suggesting the strong influence of the lateral boundary conditions on the simulation.
- The seasonal cycle of the upper ocean in RSM-ROMS shows the gradual cooling and latitudinal expansion of the equatorial Pacific cold tongue like the observations. Furthermore, the warming of the northeastern Pacific Ocean and the Caribbean Sea from winter through fall is also well captured in the RSM-ROMS simulation. However, warm SST bias of RSM-ROMS in the oceans that contrasts with the Costa Rica Dome (measured by the depth of the 20°C isotherm) being perennially shallower than the SODA reanalysis is apparent.
- The seasonal cycle of the four LLJs (CLLJ, PLLJ, PaLLJ, and TLLJ) in the RSM-ROMS simulation follow those in R2 and is reasonably similar to CFSR. However, RSM-ROMS like the R2 underestimates these LLJs with respect to CFSR, except for PaLLJ where it is comparable to CFSR. We strongly believe that this feature could be further improved in RSM-ROMS by improving the representation of the topography. By adopting a more realistic topography in the RSM, these LLJs or gap winds could become stronger and may also affect the momentum flux in the oceans and thereby also increase the mixing in the oceans that currently seems to be weaker in RSM-ROMS.

- The interannual variations of the CLLJ in the summertime show verifiable precipitation response over the CAI in the RSM-ROMS simulation, especially over Honduras Nicaragua, and Costa Rica.
- The iconic feature of the mid-summer drought in the CAI in RSM-ROMS shows some similarities with observations. For example, the observed relative maximum of intensity over the Yucatan Peninsula, Guatemala, San Salvador, Costa Rica and Panama with a minimum over Honduras is well picked. Similarly, the observed meridional gradient of lower and higher frequency to the south and to the north of Nicaragua is also reasonably simulated in RSM-ROMS.
- The track density of the TCs in RSM-ROMS shows some important observed features like there is a greater distribution of TCs in the Caribbean Sea than over the eastern Pacific and there are no TCs generated south of 5°N in the eastern Pacific.

Despite these highlights, there is significant room for improving the RSM-ROMS simulation. The systematic errors of wet bias in the winter and dry bias in other seasons, underestimation of the magnitude of the LLJs, the seasonal shift the annual cycle of the Costa Rica Dome, underestimation of the frequency of the mid-summer drought, and the higher density of TCs in the Bay of Campeche and Colombian Basin are most apparent.

Declarations

Author Contributions: Corresponding author contributed to the study conception and design. Material preparation, data collection and analysis were equally done by V. Misra and C. B. Jayasankar. All authors have read and approved the final manuscript.

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686 **Table 1: Outline of the physical parameterization schemes used in RSM**

Atmospheric Model (Regional Spectral Model)	Reference
Gravity wave drag	Alpert et al. (1988)
Boundary layer	Hong and Pan (1996)
Clouds	Zhao and Carr (1997)
Deep convection	Moorthi and Suarez (1992)
Land surface Model	Ek et al. (2003)
Longwave radiation	Chou et al. (1996)
Shortwave radiation	Chou and Suarez (1994)
Shallow convection	Tiedtke (1983)

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689 **Table 2: Mean error of seasonal mean precipitation in R2 and RSM-ROMS (units:**
690 **mm/day)**

	DJF		MAM		JJA		SON	
	R2	RSM-ROMS	R2	RSM-ROMS	R2	RSM-ROMS	R2	RSM-ROMS
Tropical Eastern Pacific (only ocean points within 95°W to 78°W and 0° to 14°N, west of the isthmus)	4.53	4.48	3.27	-1.51	5.86	-4.25	5.28	-5.12
Central America isthmus (only land points within 95°W to 78°W and 6°N to 21°N)	0.03	6.19	0.54	-0.11	2.68	-1.81	0.95	-4.29
Caribbean Sea (only ocean points within 95°W to 75°Wand 9°N to 22°N, east of the isthmus)	0.84	7.93	0.75	1.82	6.07	3.39	4.06	-0.87

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Table 3: Mean error of seasonal mean precipitable water in R2 and RSM-ROMS (units: Kg/m²)

	DJF		MAM		JJA		SON	
	R2	RSM-ROMS	R2	RSM-ROMS	R2	RSM-ROMS	R2	RSM-ROMS
Tropical Eastern Pacific (only ocean points within 95°W to 78°W and 0° to 14°N, west of the isthmus)	-2.63	-1.45	-2.09	-1.32	-7.41	-7.39	-7.76	-7.73
Central America isthmus (only land points within 95°W to 78°W and 6°N to 21°N)	1.97	2.07	2.15	2.31	-7.43	-5.97	-5.72	-4.5
Caribbean Sea (only ocean points within 95°W to 75°W and 9°N to 22°N, east of the isthmus)	1.82	1.71	3.27	3.26	-4.49	-4.28	-4.79	-4.67

Figure 1: The ocean bathymetry (color bar on the right) and land topography (color bar at the bottom) of the regional domain of RSM-ROMS used in the study. Some bathymetric features and geographical locations referenced in the text are noted in the figure. The units are in meters.

Figure 2: The seasonal climatological rainfall (mm day^{-1} ; shaded) from (a, b, c, d) observations (IMERG) and (e, f, g, h) RSM-ROMS for (a, e) December-January-February, (b, f) March-April-May, (c, g) June-July-August, and (d, h) September-October-November seasons. The corresponding seasonal systematic errors (shaded only if they exceed 95% confidence interval) of RSM-ROMS is shown in (i, j, k, and l).

Figure 3: The seasonal climatological precipitable water (Kgm⁻²; shaded), (a, b, c, d) observations (NVAP), and (e, f, g, h) RSM-ROMS for (a, e) December-January-February, (b, f) March-April-May, (c, g) June-July-August, and (d, h) September-October-November seasons. The corresponding systematic errors (shaded only if they exceed 95% confidence interval) of RSM-ROMS is shown in (i, j, k, and l).

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717 **Figure 4:** The seasonal climatological SST ($^{\circ}\text{C}$; shaded), (a, b, c, d) observations (NOAA
 718 OISSTv2), and (e, f, g, h) RSM-ROMS for (a, e) December-January-February, (b, f) March-April-
 719 May, (c, g) June-July-August, and (d, h) September-October-November seasons. The
 720 corresponding systematic errors (shaded only if they exceed 95% confidence interval) of RSM-
 721 ROMS is shown in (i, j, k, and l).

Figure 5: The monthly mean climatology of the thermocline depth (measured as the depth of the 20°C isotherm) overlaid with surface ocean currents (ms^{-1}) from SODA. The prominent surface ocean currents in the region like the westward flowing North Equatorial Current (NEC) and South Equatorial Current (SEC), eastward flowing North Equatorial Countercurrent (NECC), and northwestward flowing Costa Rica Coastal Current (CRCC) of the Pacific and the Caribbean Current (CC) in the Atlantic are identified in the first panel.

Figure 6: The monthly mean climatology of the thermocline depth (measured as the depth of the 20°C isotherm) overlaid with surface ocean currents from the RSM-ROMS simulation. The prominent surface ocean currents (ms^{-1}) in the region like the westward flowing North Equatorial Current (NEC) and South Equatorial Current (SEC), eastward flowing North Equatorial Countercurrent (NECC), and northwestward flowing Costa Rica Coastal Current (CRCC) of the Pacific and the Caribbean Current (CC) in the Atlantic are identified in the first panel.

Figure 7: The climatological annual mean 925hPa winds from a) R2, b) CFSR, c) RSM-ROMS, and d) corresponding difference between RSM-ROMS and CFSR (c-b). The seasonal cycle of the 925hPa winds for e) Caribbean LLJ Index (averaged over 12.5°N-17.5°N and 76°W-80°W), f) Panama LLJ Index (averaged over 3°N-8°N and 76°W-80°W), g) Papagayo LLJ Index (averaged over 11°N-14°N and 82°W-90°W), and h) Tehuantepec LLJ Index (averaged over 13°N-18°N and 94°W-96°W). The shaded values in (a-d) is the magnitude (ms^{-1}) of the wind vectors.

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Figure 8: a) The Caribbean Low Level Jet (CLLJ) index for JJA from CFSR (black line) and RSM-ROMS (red line). The open and filled circles indicate high and low CLLJ index seasons. The solid horizontal lines are the corresponding mean of the CLLJ index from CFSR (black) and RSM-ROMS (red). The dotted horizontal lines on either side of the mean is (mean $\pm$ 0.75 $\times$ standard deviation) of the CLLJ index from CFSR (black) and RSM-ROMS (red). The corresponding difference of seasonal mean rainfall (mm day⁻¹) between high and low CLLJ index years for JJA from b) CPC and c) RSM-ROMS. Only significant values at 95% confidence interval according to t-test in (b, and c) is plotted.

Figure 9: The (a, b) intensity (I ; units: mm day^{-1}) and (c, d) frequency (F ; fraction of events in 25 years: 1986-2010) of the mid-summer drought events from (a, c) CPC and (b, d) RSM-ROMS simulation.

Figure 10: The track density plot of the TCs (i.e., number of TCs per $1^\circ \times 1^\circ$) from a) HURDAT2 and b) RSM-ROMS simulation over the time period of 1986-2010. It may be noted that the track density is based on once-a-day tracking data for both panels. The tracking algorithm used in (b) follows from Ullrich and Zarzycki (2017).

Figure 11: The seasonal mean climatological error of (a-d) rainfall (mm day^{-1}) and (e-h) precipitable water (kgm^{-2}) from R2 reanalysis used in the lateral boundary forcing of RSM-ROMS for (a, e) DJF, (b, f) MAM, (c, g) JJA, and (d, h) SON. The observations for rainfall and precipitable water is IMERG and NVAP, respectively.

Figure 12: The seasonal mean climatology of wind stress (Nm^{-2} ; vectors) and the curl of the wind stress (Nm^{-3} ; shaded) from (a-d) observations, (e-h) R2 reanalysis, and (i-l) RSM-ROMS simulation for (a, e, i) DJF, (b, f, j) MAM, (c, g, k) JJA, and (d, h, l) SON seasons.