The Intra-Americas Sea: Challenges and Opportunities to Understand North American Climate Variability and Predictability

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Abstract:
The Intra-Americas Seas (IAS), which includes the Gulf of Mexico and the Caribbean Sea, hosts part of the largest warm water pool in the Western Hemisphere and is a primary moisture source for precipitation in the Americas. This region undergoes significant variability from interannual to secular changes that impacts North American climate variability including the frequency of Atlantic tropical cyclones and southeast US tornadoes. Moisture evaporated from the IAS contributes to the severity of winter storms in the mid-Atlantic and northeast United States. The onset of North and South American monsoons also modify the trade winds with potential impacts on IAS surface fluxes and sea surface temperature (SST) variability. Most climate models display significant cold SST and dry rainfall bias over the IAS with a corresponding underestimation of their variability. In addition significant biases in IAS surface and deep ocean currents, and poor representation of the atmospheric low-level jets in the region are also apparent in the many of the climate models. Lack of observations in both the atmosphere and ocean in the IAS, which is one of the most poorly observed regions of the world, limits our ability to improve models. This paper provides recommendations to advance understanding of IAS climate variability and improve its predictability in the midst of a rapid deterioration of the observing network across the region.

1 Introduction

The Western Hemisphere Warm Pool (WHWP) located between North and South America and between the tropical North Pacific and Atlantic Ocean (Fig. 1a) is the second largest body of very warm water (≥28.5°C) on Earth and hosts the second largest diabatic heating center of the tropics during the boreal summer (Wang and Enfield 2001). The WHWP spans the northeast tropical Pacific west of Central America (also known as the Eastern Pacific Warm pool [EPWP]), the Intra-Americas Sea (IAS; i.e., the Gulf of Mexico and the Caribbean Sea) and the western tropical North Atlantic (the latter two regions also collectively referred to as the Atlantic Warm Pool [AWP]; Wang and Enfield 2001, 2003). The AWP serves as a host to nearly 80% of the WHWP and dominates the variability of the latter. The WHWP appears first in late boreal spring in the northeast Pacific followed by a warming in the Atlantic in boreal summer and fall. As a consequence of its geographic location and strong sea surface temperature (SST) variations superposed on a warm ocean basic state, the IAS plays a significant role in modulating atmospheric and terrestrial variability both locally within the IAS and over North America, particularly over the continental United States [CONUS], Central America and northern South America (Wang et al. 2006, 2008; Ruiz-Barradas and Nigam 2005; Mestas-Nunez et al. 2007; Misra et al. 2014; Poveda et al. 2014).

The IAS is a complex system influenced by large-scale atmospheric features including the North Atlantic Subtropical High (NASH), the monsoon systems of the Americas, and the (continental and oceanic) Inter-Tropical Convergence Zones (ITCZs). Seasonally persistent small-scale features such as atmospheric low level jets (LLJs; e.g. North American, Choco, Tehuantepec, Papagayo LLJs) arising from the complex topography (Fig. 1b) and seasonal migration of the NASH are also important in the region. Ocean processes impacting the region include the upper branch of the Atlantic Meridional Overturning Circulation (AMOC) that includes the Loop Current and the associated eddies in the Gulf of Mexico (GoM), prevalent subtropical cells, and the Caribbean current system with its ubiquitous mesoscale and sub-mesoscale eddies sustained by a complex bathymetry (Fig. 1b). Embedded within these defining features of the IAS are extreme events such as Atlantic, GoM, and Caribbean tropical cyclones (TCs), tornadoes in the CONUS, monsoon severe thunderstorms, large and intense meso-scale convective systems over the far eastern tropical Pacific and climate anomalies that arise largely in response to variability on time scales ranging from intraseasonal to secular changes in the IAS (Maloney and Hartmann 2000a, b; Wang and Lee 2007; Wang et al. 2007, 2008a; Klotzbach 2014; Crosbie and Serra 2014; Serra et al. 2014). The purpose of this paper is to highlight IAS variability and its role in regulating the climate of North America, followed by a discussion of the limitations of our current climate models and observational networks to effectively predict and monitor the observed IAS climate, respectively. We present recommendations and emerging opportunities to improve climate predictability and monitoring of IAS variability spanning from intra-seasonal timescales to secular changes.

2 The IAS Teleconnections

The WHWP that includes the AWP defined by the 28.5°C surface isotherm (Fig. 1a) has a distinct seasonal cycle (Wang and Enfield 2001, 2003; Enfield and Lee 2005;

Lee et al. 2007) with a maximum around September 3 (Misra et al. 2014). The choice of the 28.5°C isotherm to define the AWP stems from its close correspondence with the variations in the IAS mixed layer depth (Wang and Enfield 2003), its notable impact on organized convection (Graham and Barnett 1987), and the display of the strongest interannual variations of the area enclosed by this isotherm in the IAS (Misra et al. 2013). The variability of the WHWP is dominated by the AWP variability owing to its relative size and seasonal persistence compared to the EPWP component (Wang et al. 2008a). However the EPWP plays an important role in the North American Monsoon (NAM) variability (Adams and Comrie 1997) and also serves as a precursor to the development of the AWP (Enfield and Lee 2005; Lee et al. 2007).

The AWP-induced diabatic heating at its seasonal peak forces a Gill-type atmospheric response as well as extratropical stationary waves (Fig. 2) that produces rainfall variability over the CONUS (Fig. 3a), while modulating the subtropical highs in the southeastern Pacific and in the North Atlantic Oceans (Fig. 3b). AWP variations and their teleconnections to North American hydroclimate are also observed to be largely independent of the El Niño and the Southern Oscillation (ENSO) variations in the equatorial Pacific (Wang et al. 2006, 2008; Misra et al. 2013).

During the boreal summer season, easterly trade winds carry moisture from the Tropical North Atlantic (TNA) into the Caribbean Sea where the flow intensifies forming the easterly Caribbean low-level jet (CLLJ) due to a strong meridional pressure gradient established by the NASH (Fig. 1a; Amador and Magana 1999; Poveda and Mesa 1999; Wang 2007). As the CLLJ transits the Caribbean Sea, it splits into two branches: one turning northward and forming the Great Plains LLJ (GPLLJ) while the other branch continues westward across Central America into the eastern North Pacific (Fig. 1a). Modeling and observational studies indicate that a large (small) AWP is associated with weakening (strengthening) of the summertime NASH and strengthening (weakening) of the summertime continental low over the NAM region (Fig. 3b; Wang et al. 2008a). The observational studies also confirm that in response to the pressure changes, a large (small) AWP weakens (strengthens) the southerly GPLLJ (e.g. Wang 2007), which results in reduced (enhanced) northward moisture transport from the GoM to the region east of the Rocky Mountains and thus decreases (increases) the moisture available for summer rainfall over the central United States (Wang et al. 2006; Ruiz-Barradas and Nigam 2006; Mestas-Nuñez et al. 2007).

An anomalously strong CLLJ is also associated with reduced precipitation over the Caribbean and northern South America, including portions of Colombia and Venezuela (Fig. 3; Poveda and Mesa, 1999; Wang 2007; Cook and Vizy 2010; Martin and Schumacher, 2011b) and along the Pacific coast of Central America (Amador 1998). Concomitantly, a strong CLLJ is associated with above normal rainfall over the Caribbean coast of Central America, including Nicaragua and Costa Rica, particularly during the boreal summer (May–September), through orographic enhancement and large-scale low-level convergence at the jet exit (Waylen et al. 1996; Magaña et al., 1999; Amador et al. 2000; Magaña and Caetano, 2005). Such enhanced summertime convective activity on the upslope side of the terrain acts to deprive moisture to the Pacific coast of Central America and results in rainfall deficits there (Amador 1998; Cook and Vizy 2010; Martin and Schumacher 2011b).

The rainfall over the AWP alternates with the Amazon basin in South America as the seasonal heating source for the regional Hadley and Walker type circulations in the Western Hemisphere (Wang et al. 2006; Poveda et al. 2006). During the boreal summer/fall season, a strong regional Hadley-type circulation is established, with ascending motion over the AWP and subsidence over the southeastern tropical Pacific. A large (small) AWP during the boreal summer/fall results in a strengthening (weakening) of this regional Hadley-type circulation with enhanced descent (ascent) over the southeastern tropical Pacific (Wang et al. 2006, 2010, 2014; Lee et al. 2013b).

3 Secular change in the IAS

Among the challenges in interpreting observations of climate variability in the IAS region are the detection and attribution of low–frequency variability and long–term trends in SST. In particular, the IAS region is home to substantial internal multidecadal variability (Ting *et al.* 2011) that can mask forced long–term trends using instrumental records alone (Wang and Dong 2010) and often require historical hindcast model experiments to interpret (Ting *et al.* 2009). Moreover, future hydroclimate projections by state–of–the–art GCMs suggest that part of the IAS region including EPWP eastward through the Caribbean islands (including Central America, Mexico, and the southern United States) should anticipate a robust and severe reduction in precipitation (Neelin *et al.* 2006; Meehl *et al.* 2007; Taylor *et al.* 2012; Maloney *et al.* 2014), which is thought to be critically dependent on the magnitude and spatial patterns of ocean warming (Schubert *et al.* 2009; Xie *et al.* 2010; Rauscher *et al.* 2011; Lee et al. 2011).

Future projections by the Coupled Model Intercomparison Project phase 5 (CMIP5) models predict that under the Representative Concentration Pathway (RCP8.5), GoM SST will warm in a spatially uniform fashion in the multi–model ensemble mean at a rate of 3–4°C century^{-1} by the end of the 21st century, and 2–3°C century^{-1} in the Caribbean Sea (IPCC AR5, WG1, Ch. 12, Fig. 12.11 [Collins *et al.* 2013]). The projected trends in these marginal seas approximately mirror those for the open Atlantic Ocean regions immediately to their east.

A comparison of four gridded instrumental datasets (see supplementary material [SM] S.1) agree on a broad warming across the western Tropical Atlantic Ocean and the Caribbean Sea of ~0.4°C century^{-1} with the greatest warming found along the northern coast of South America in the southern Caribbean Sea (Fig. SF1). In stark contrast, SST trends within the GoM are spatially heterogeneous and vary substantially from one gridded product to another (Fig. SF1). Of particular relevance to recent and ongoing observational efforts is that SST changes in the GoM are highly divergent among the four data sets, varying by ~0.5°C relative to their 1951-1980 base periods. Published proxy reconstructions of water temperature corroborate the trends in the super ensemble of observations[1] (Fig. 4). The strongest proxy reconstruction trends are in the central Caribbean Sea (Fig. 4), where the super ensemble shows agreement with a positive trend. The weakest trends are in the GoM, but they are calculated from sediment cores with few points covering the 20th century (Richey et al, 2004; Richey et al., 2008). The record from the Cariaco Basin (off the north central coast of Venezuela; Black et al., 2008) is based on high time resolution through the 20th century, and it is the only proxy

[1] See SM S.1 on further details of super ensemble of observations

reconstruction in the basin that approaches the sign and magnitude of the trends in the super ensemble. The magnitude of these trends are based on species-specific calibrations that, in the case of the Caribbean Sea and GoM, seem to over-estimate long-term temperature change. It is important to note that records from Jamaica (Hasse-Schramm et al., 2003), Pedro Bank (central Caribbean, Haase-Schramm et al., 2003), and the Bahamas (Rosenheim et al., 2005) are from shallow subsurface records (28–67m). The general pattern of change in the instrumental SST products (smaller and/or less robust warming trends in the GoM than in the Caribbean Sea) is exactly opposite to the future trends projected by CMIP5 models.

It is important to point however that future projections by CMIP5 models may need to be downscaled to better understand the regional response of the IAS as they seem to have significant ocean circulation bias in the IAS (Liu et al. 2102, 2015). Liu et al. (2015) downscaled the CMIP5 model simulations under historical and two future emission scenarios using an eddy-resolving regional ocean model. They reported that the simulated volume transport by the western boundary current system in the IAS, including the Caribbean, Yucatan and Loop Currents, was reduced by 20-25% during the 21st century, consistent with a similar rate of reduction in the AMOC. Their modeling analysis also showed that the projected reduction of the IAS western boundary current system was linked to reduced upwelling and enhanced warming along the western boundary.

4 The monsoons and the IAS

The IAS by its proximate location to the monsoons of the Americas serves as a bridge to unify the South American Monsoon (SAM) and North American Monsoon (NAM) systems. Observations have shown an anti-correlation between boreal winter rainfall anomalies over the IAS region and those over the NAM and Southern Amazonia on interannual (Wang and Fu 2002) and multi-decadal scales (Arias et al. 2014). On interannual time scales these relationships are linked by the reversal of the cross-equatorial flow over the Americas associated with ENSO (Arias et al. 2014). Misra and DiNapoli (2013) also find that the anomalous meridional migration of the ITCZ in the western Atlantic Ocean dictated by the intensity of the seasonal rainfall activity in the equatorial Amazon during boreal winter is associated with the upper ocean heat content and surface temperature variability of the IAS in the subsequent seasons (Fig. 5). This teleconnection is established by the modulation of the atmospheric heat fluxes regulated by the overlying regional Hadley circulation, which gives rise to a robust negative correlation between austral summer rainfall over the equatorial Amazon and SST variability in the IAS during the subsequent boreal summer season.

On decadal time scales the anti-correlation of rainfall between the NAM and the SAM is also related to the intensification of rainfall over the IAS region, causing an early retreat of the NAM and a late onset of the rainy season over southern Amazonia. These variations seem to be a combined result of the westward shift of the NASH and the warm phase of the Atlantic Multidecadal Oscillation (AMO; Arias et al. 2014).

Previous studies on the sources of NAM moisture indicate that moisture at upper levels (above 700 hPa) originates to the east of the Sierra Madres Occidental and over the GoM, while low-level moisture of oceanic sources originates predominantly from the tropical Pacific Ocean and the Gulf of California (Schmitz and Mullen 1996; Adams and Comrie 1997). In addition to these oceanic sources, studies using a two-dimensional dynamic recycling model find that NAM terrestrial evapotranspiration accounts for approximately 40% of the total moisture sources to the NAM (Dominguez et al. 2006; Hu et al. 2014), similar to estimates of terrestrial sources for the May-July period based on moisture tracers in a GCM (Bosilovich 2003).

While the seasonal migration of the monsoon ridge is fundamental to the onset and demise of the NAM, monsoon precipitation is highly dependent upon intra-seasonal variability (ISV), especially in the southwestern United States at the northern edge of the monsoon. ISV in the IAS (see Serra et al. 2014 for a review) plays a particularly important role in supporting NAM rainfall events through gulf surges, or surges of moisture up the Gulf of California (e.g., Stensrud et al. 1997), and through IAS TC activity that recurves directly over the NAM region (Collins and Mason 2012; Ritchie et al. 2011; Corbosiero et al. 2009; Wood and Ritchie 2013). The link between tropical disturbances, gulf surges, the MJO, and NAM rainfall events has not been fully explored; however studies suggest that positive rainfall anomalies extend into the NAM region during the transition to the MJO westerly phase, when TC and easterly wave activity is also enhanced and shifted along the Mexican west coast (Lorenz and Hartmann 2006; Crosbie and Serra 2014; see SM [S.2])

The Mid-Summer Drought (MSD; Fig. 6) phenomenon characterized by a minimum in rainfall that separates two peaks in rainfall across the Caribbean and Central America during boreal summer season has been associated with the westward expansion and intensification of the NASH and associated CLLJ variability resulting in moisture flux divergence (Hastenrath 1976, 1978; Granger 1985; Magana et al. 1999; Giannini et al. 2000; Mapes et al. 2005; Wang 2007; Wang and Lee 2007; Misra et al. 2014). As the NASH expands westward during the summer months precipitation is suppressed via large-scale subsidence and increased stability (Knaff, 1997; Mapes et al. 2005; Wang and Lee 2007; Kelly and Mapes 2011). In addition, the enhancement of the easterly trade winds in the CLLJ leads to greater evaporative cooling and lower SSTs which further suppresses convection (Muñoz et al. 2008, Xie 2006, Martin and Schumacher 2011b).

5 The ocean circulation in the IAS

The IAS is comprised of a very complex ocean circulation system from the smallest spatial scales (e.g. ubiquitous presence of mesoscale and sub-mesoscale eddies; see SM [S.3]) to hosting the upper branch of the AMOC (e.g. Caribbean current system, GoM loop current and the southern branch of the Gulf stream system). These ocean current systems that bring relatively warm and saline water from the equatorial Atlantic to the GoM are fundamental to the regulation of the SST in the IAS (Jayne and Maotzke 2002; Chang and Oey 2010; Liu et al. 2012; Misra et al. 2015). Owing primarily to shortcomings in long-term observational network of the IAS, substantial uncertainty exists regarding the magnitude of inter-annual flow variability within the channels (Yucatan, Old Bahama [OBC] and Northwest Providence [NWP] channels) that connect to the Florida Straits at 27° N (Rousset & Beal 2014; see SM [S.4]).

The southern Caribbean upwelling system significantly contributes to the mean SST distribution and its seasonal and inter-annual variability in the Caribbean Sea. The upwelling follows a semi-annual cycle resulting in surface manifestation of relatively cold SST's, which peaks in December–March and July (Fig. 7) in response to semi-annual intensification of the along-shore CLLJ (Wang 2007; Wang and Lee 2007) and its associated changes in wind stress curl (Inoue et al. 2002; Andrade and Barton 2005; Cook and Vizy 2010). Jouanno and Sheinbaum (2013) also find that vertical turbulent mixing is important in regulating the surface cooling at the coast. The intense vertical mixing at the coast arises from vertical shear between the shallow and strong Caribbean Current and the subsurface Caribbean coastal undercurrent. This southern Caribbean upwelling region limits the southward extent of the AWP during its onset and peak phase (Lee et al. 2007) and modulates the meridional gradient of SST over the Venezuela and Colombia basins, with possible feedback to the trade winds (Chang and Oey, 2013). See SM (S.5) for further discussion on the rectification effect on the atmosphere from the cold SST's of this upwelling region.

It is also observed that a net freshwater gain (loss) in the TNA region coincides with the large (small) AWP regimes on multi-decadal timescales (Wang and Zhang 2013; Zhang and Wang 2012). Zhang et al. (2014) further confirm that this AWP induced freshwater flux plays a negative feedback role that acts to restore the AMOC from its anomalous state through basin-scale gyre circulation adjustments.

6 The IAS extremes

A number of studies have shown that IAS variability modulates Atlantic TC activity (Fig. 8; e.g. Molinari et al. 1997; Maloney and Hartmann 2000a,b; Higgins and Shi 2001; Wang and Lee 2007, 2009; Wang et al. 2007, 2008a, 2011; Aiyyer and Molinari 2008;

Jiang et al. 2012; Klotzbach 2014; Crosbie and Serra 2014). Calculated over a 34-year period (1979-2012), 90 named Atlantic TCs occurred in small AWP years versus 163 for large AWP years (Fig. 8). Wang and Lee (2007) argued that the observed teleconnection between IAS and Atlantic TC activity is a result of the IAS SST-forced changes of the vertical wind shear and moist static stability in the Atlantic TC Main Development Region (MDR). Several studies have also suggested an influence of the Atlantic Meridional Mode (AMM) on the Atlantic TC activity (Vimont and Kossin 2007; Kossin and Vimont 2007; Patricola et al. 2014).

Dynamically, the AWP-forced atmospheric circulation pattern is baroclinic within the tropical latitudes, with a large warm pool producing a cyclone in the lower troposphere and an anticyclone in the upper troposphere, both situated on the northern flank of the AWP (Wang et al. 2008a; Fig. 2). This anomalous circulation structure reduces the lower tropospheric easterly flow and the upper tropospheric westerly flow over the AWP, thus reducing the vertical wind shear in a way that favors atmospheric convection (or TC development; Wang et al. 2007; 2008a). Similarly, the modulation of large-scale environmental factors including tropospheric humidity, vorticity, vertical shear, and SST by the MJO and other modes of ISV can also manifest in IAS-Atlantic TC and extreme rainfall teleconnections (Maloney and Hartmann 2000a; Camargo et al. 2009; Martin and Schumacher 2011a; Jiang et al. 2012). However, precursor disturbances for TCs in the form of easterly waves are also modulated on intraseasonal timescales in the IAS region (e.g. Maloney and Hartmann 2001; Crosbie and Serra 2014; Rydbeck et al. 2014). Thermodynamically, the AWP increases convective available potential energy (CAPE) that provides the fuel for moist convection and thus facilitates the formation and development of TCs (Wang et al. 2008). AWP-forced extra-tropical stationary Rossby waves influence the barotropic atmospheric flow over North America and the Atlantic Ocean in boreal summer (Lee et al., 2009), which then affects the steering flow of North Atlantic TCs (Wang et al., 2011; see SM [S.6]).

Boreal spring (April and May) is the primary season for tornadoes in the US. The convergence of dry upper level air from higher latitudes and low-level warm moist air from the GoM result in a conditionally unstable environment to the east of the Rockies with a raised CAPE that make it conducive for tornadogenesis. Lee et al. (2013a) show that the April-May tornado outbreaks in the US are significantly correlated with moisture transport from the GoM (Fig. 9) and lower tropospheric vertical wind shear in the central and eastern US. Muñoz and Enfield (2011) find that the tornadic activity in lower Mississippi, Tennessee, and Ohio is particularly related to the interannual and decadal variability of the LLJs (Caribbean and North American) that dictate the moisture flux to the region.

The relatively warm SST in the GoM compared to surrounding areas can result in rapid cyclogenesis in the boreal winter season. For example, the Superstorm of 1993 (12-14 March) was comparable in strength to a category 1 hurricane, with winds reaching 70 kt along the storm path, and producing storm surges more than 10 ft along the northeast Gulf Coast (Schumann et al. 1995; Bosart et al. 1996). The GoM is a well-known source for winter cyclones (Dickinson et al. 1997). Among other atypical features of the environment that led to the rapid cyclogenesis event of Superstorm of 1993 (see Bosart et al. 1996), a well noted feature was the presence of Eddy Vasquez over the northwest GoM (Fig. 10), a warm water eddy shed from the loop current (Schumann et al. 1995).

This warm eddy not only provided the necessary air-sea fluxes to rapidly develop the extratropical cyclone but also provided a very strong surface temperature gradient (baroclinicity) from the cold continental shelf waters (Schumann et al. 1995).

7 The barrier layers of the IAS

The thick and persistent barrier layers or fresh water lens that form in the northwest tropical Atlantic Ocean is a result of the large volume of fresh water influx from some of the major continental rivers in northern South America (e.g., Orinoco and Amazon; Pailler et al., 1999). The barrier layers of the IAS are one of the most prominent structures in the world tropical and subtropical oceans (Mignot et al. 2007), which reflect the prevalence of the robust land-atmosphere-ocean coupling in the region. The mighty Amazon and Orinoco River systems together form the biggest river system in the world in terms of discharge (0.2 Sv). The Guyana current carries a significant portion of this water into the Caribbean Sea during boreal summer and fall (Hu et al., 1997), providing the largest term in the surface salinity balance of the region (Foltz and McPhaden, 2008; see SM S.7).

Because of the barrier layer's role in trapping heat within the upper ocean layer and reducing surface cooling, the barrier layer in the northwestern tropical Atlantic has the potential to affect Atlantic TC activity (Ffield 2007; Vizy and Cook 2010, Balaguru et al. 2012). When TCs pass over barrier layers, the reduced efficacy of vertical mixing in these highly stratified layers leads to reduced SST cooling, which then impacts TC evolution by maintaining strong air–sea fluxes. In the northwestern tropical Atlantic, Balaguru et al (2012) reported that the mean intensification factor of Atlantic TCs increases from 0.48 ms^{-1} over a 36-hour period in non-barrier layer regions to 0.98 ms^{-1} in barrier layer regions.

8 The limitations of numerical models simulating and predicting IAS climate

The surface temperature of the IAS and its variability is grossly underestimated by a majority of the CMIP models that participated in the IPCC AR4 and AR5 especially in the boreal summer and fall seasons (Fig. 11; Misra et al. 2009; Kozar and Misra 2012; Liu et al. 2012, 2013; Ryu and Hayhoe 2013). A majority of the CMIP5 models underestimate the area of the AWP in their 20th century simulation (Fig. 11). Surface heat budgets computed over the IAS reveal that the radiative fluxes of downwelling shortwave and longwave radiation dominate the maintenance of the AWP (Misra et al. 2013). However, the important and subtle role for air-sea turbulent fluxes in the variability of the AWP and its interplay with the variability of the overlying low-level easterlies, which is related to the strength and position of the NASH, cannot be overemphasized (Misra et al. 2009). Liu et al. (2012) and Misra et al. (2013) find that a majority of current Global Circulation Models (GCMs) and reanalysis products of the ocean and the atmosphere produce a qualitatively correct surface heat budget in the IAS. However, when compared to reanalysis, GCMs have a tendency to underestimate latent heat flux and downwelling shortwave flux, and overestimate sensible heat flux. Liu et al. (2012) also indicate that the upper ocean heat budget in the GoM is a complex balance of contrasting warming influence of upper ocean heat transports (e.g. Loop Current) and cooling influence of the net surface flux. In effect with significant GCM biases prevalent in surface fluxes and ocean circulation over the region (Liu et al. 2015), it is not surprising to note such large SST errors over the IAS in GCMs (Fig. 11).

The GCMs also display severe bias over equatorial Atlantic. Like some of the most fundamental features of the equatorial Atlantic Ocean – the east-west equatorial SST gradient and the eastward shoaling thermocline – cannot be reproduced by most global climate models (Fig. 12; e.g. Richter and Xie, 2008; Richter et al. 2014). It is concerning that like the previous feature in Fig. 11, this grave bias in the equatorial Atlantic has also persisted across generations of model development (Fig. 12). Xu et al. (2014) found that the southeastern tropical Atlantic warm SST bias can remotely force a significant cold SST bias and dry precipitation bias in the IAS region (see their Fig. 18).

The IAS region is characterized by significant deficiencies in the simulation of convection and related processes. For example, ISV in the east Pacific and Caribbean is overall poorly simulated in amplitude and spatial structure in climate models (Jiang et al. 2013), with poor performing models also tending to have common mean state biases in winds and other fields. Martin and Schumacher (2012) and Ryu and Hayhoe (2013) showed that cold SST bias of the IAS was the dominant cause of the negative rainfall bias across CMIP3 coupled historical simulations. In more recent CMIP5 historical simulations, large precipitation and SST biases across the Caribbean are still evident in both coupled (historical) and uncoupled (AMIP) simulations (see SM [S.8]). CMIP5 models produce too little mean precipitation in Central America and adjacent regions of the east Pacific and Caribbean during boreal summer in the multimodel mean, and also not enough precipitation in Caribbean island regions (Sheffield et al. 2013). Furthermore, the timing of NAM precipitation also exhibits substantial biases, with models generally producing too little precipitation in the NAM region in early and mid boreal summer, and excessive precipitation late in the monsoon period (Liang et al. 2008; Sheffield et al. 2014a), associated with difficulty in ending the monsoon (Geil et al. 2013).

Many of the biases described above may be related to deficiencies in the deep convection parameterizations of climate models, and are further exacerbated by the large SST bias in the IAS. For example, many deep convection parameterizations used in climate models do not exhibit adequate sensitivity to free tropospheric humidity (e.g. Derbyshire et al. 2004), or lack representations of cumulus momentum transport and the effect of mesoscale circulations on convective gustiness that affects surface fluxes. Model ISV in the IAS region can be improved by suppressing convection through enhanced moisture sensitivity or other means, although often the quality of the mean state degrades when such modifications are not done with care (e.g. Kim et al. 2011; Maloney et al. 2014a), suggesting that such changes can improve ISV for the wrong reasons (e.g. Maloney et al. 2014a; Hannah and Maloney 2014).

The recent effort to make seasonal predictions from multiple models like the North American Multi-Model Ensemble project (NMME; Kirtman et al. 2014) routinely available allows us to assess skill of predicting SSTs over the AWP, and terrestrial anomalies over the surrounding areas. Misra and Li (2014) find that the seasonal predictability of the AWP in the NMME models is promising compared to earlier generation GCMs (Misra et al. 2009), but also displays notable limitations. For example, NMME mean forecasts indicate that skill in predicting seasonal mean SSTs in the AWP is much less than the skill in predicting SST variability in the equatorial eastern Pacific associated with ENSO (Fig. 13). Anomaly correlations averaged over the AWP region are in the range of 0.5 – 0.6, much less compared to those over the eastern equatorial Pacific (Niño3.4 region). Prediction of terrestrial climate anomalies around the IAS relies on skillful prediction of IAS SSTs, and hence seasonal prediction over land areas cannot be improved until skill in predicting SST is improved. An assessment of the corresponding JJA seasonal mean precipitation anomalies confirms this (Fig. 14) in which prediction skill (anomaly correlation of the ensemble mean) generally does not exceed 0.3. See SM (S.10) for comparison of this skill with the boreal winter season.

9 The observing network in the IAS

A gradually and significantly degrading observational network for the atmosphere and ocean in the IAS region also exacerbates the modeling challenges described in the previous section. For example, the radiosonde network in the region is quite sparse (Fig. SF10; SM [S.9]) and temporally incomplete. For a region that displays a highly heterogeneous distribution of rainfall, the rain gauge network is highly inadequate (Fig. SF11). The complex spatial structure of rainfall in the IAS (with topographic, island, and LLJ effects) raises a challenge for observing and monitoring the climate of the region. For example, Figs. 6a and b show rainfall measurements from the satellite based Tropical Rainfall Measuring Mission (TRMM) and merged satellite and in-situ Global Precipitation Climatology Project (GPCP) data, respectively, and differences between the products are evident in both the location and magnitude of rainfall features across the region. Using TRMM precipitation radar data, Sobel et al. (2011) show that rainfall enhancement is more significant over larger (greater than 315 km^2) islands than smaller islands in the Caribbean and that smaller islands have a negligible or even negative change in rainfall intensity and frequency relative to surrounding oceans. However, while high-resolution satellite data such as TRMM and the newly launched Global Precipitation Measurement (GPM) satellites can provide useful information, the 0.25° TRMM resolution (~28km) provides fewer than 3 measurements for most of the smaller islands in the Lesser Antilles. For example, the island of Dominica (approximately 25km wide and 50 km long) has large meridional variations in rainfall intensity and frequency as measured during the DOMEX (Dominica Experiment) field campaign that are not captured by current satellite measurements of rainfall (Smith et al. 2012, Minder et al. 2013).

In-situ ocean observational networks in the IAS have also diminished considerably over time. ARGO floats, which have the operating depth of about 2000m cannot pass the Lesser Antilles to reach the Caribbean Sea. As a result, the IAS is currently as poorly observed as the polar oceans (Fig. SF12). This poor observational network leaves some basic issues unresolved including characteristics of the seasonal cycle of the oceanic variables in the IAS and its relation to the Loop Current and eddy shedding dynamics, and the different pathways of the flow that can impact the salt, temperature and fresh water transport by the AMOC through the western boundary. As mentioned earlier, there is significant disparity in the diagnosis of the observed long-term variations of SST in the IAS.

10 Discussion and Conclusions

The IAS produces robust local and remote climate variability on intra-seasonal timescales to decadal timescales, and also exhibits secular change. In the midst of weak ENSO forcing over North America especially during boreal summer and fall seasons, the the development of SST anomalies in the IAS region superimposed on a warm mean SST state (WHWP) can influence atmospheric and terrestrial variability over remote regions via different pathways, including modulation of low level moisture transport and changes in convection over the AWP, including remote teleconnections induced by these convection anomalies.

Coordinating with ongoing projects and leveraging technological advances could significantly ameliorate the observational gaps in the IAS and its adjacent nations. For example, in response to the need to monitor both short and long term natural hazards throughout the Caribbean and Mexico, the National Science Foundation (NSF) has funded two GPS-based atmospheric sounding initiatives: The Continuously Operating Caribbean Observation Network (COCONet) (Braun et al. 2012) and the Trans-boundary, Land and Atmosphere Long-term Observational and Collaborative Network (TLALOCNet) in Mexico. Mexican partners, including the National Autonomous University of Mexico (UNAM) will also add to this network (see SM S.11). Data products from these networks include estimates of column integrated tropospheric water vapor; surface meteorological variables including wind speed and direction, air temperature, humidity and precipitation; time series of daily positions and component velocities for each station (used to quantify tectonic changes in the region); and high-rate low-latency data from a subset of stations. Similarly, new techniques like Global Positioning System (GPS) Radio Occultation (RO) measurements can provide relatively accurate atmospheric sounding data with high vertical resolution especially above the atmospheric boundary layer (Anthes et al. 2008) that could supplement the sparse upper air network in the region.

Several important measuring and modeling projects for the GoM are currently in progress or are about to be financed either by the oil industry and/or government agencies from different countries (e.g. the Gulf of Mexico Research Institute). A coordinated approach to achieve some basic common goals among these projects could help sustain a multinational long-term observing and monitoring program at least for the GoM. In addition, new technologies including underwater gliders appear to be particularly ideal for setting up a relatively low cost long-term observing program for the IAS in which cross-sections across the Caribbean Current or in the Gulf of Mexico could be regularly sampled. Choosing some of these sections to coincide with satellite-altimetry tracks would provide invaluable information for calibration of altimetry against observed stratification and heat content. Glider observations seem especially well suited for studies of hurricane intensification, which could be supplemented with greater coverages by surface drifters and IRIDIUM-controlled ARGO floats. A pilot underwater glider project for the Caribbean Sea has already started since 2014 (http://www.aoml.noaa.gov/phod/goos/gliders). Coordinated intensive field observations can also be planned to provide high temporal and spatial resolution atmospheric and oceanic observations in limited regions to better understand local and regional physical process and to provide high quality datasets for improving models. High resolution modeling studies may also provide a bridge to improving global model physical parameterizations, if coordinated with such field programs.

One of the current grand challenges in seasonal-to-interannual prediction is rainfall during the warm season over North America, which is a difficult problem because much of the rainfall is associated with relatively small-scale structures (e.g., thunderstorms and mesoscale convective complexes) that are not adequately resolved by the current generation of seasonal-to-interannual prediction systems. As a consequence, large systematic errors occur in the predicted rainfall anomalies that seriously limit forecast quality. Fundamental predictability issues also exist, namely that remote forcing from remote climate variability (e.g. ENSO) is relatively weak and the rainfall signal-to-noise ratio over North America is comparatively small. Nevertheless, our current understanding indicates that current prediction systems are underperforming, even with the limited predictability (see SM [S.10]).

The IAS climate processes outlined in this paper provide opportunities to improve local and remote North American hydroclimate prediction across temporal scales. An opportunity also exists to provide a more holistic picture of Western Hemisphere climate and its prediction by bridging tropical South American climate variability with that of the IAS. Moving forward, the modeling and observational challenges of IAS have to be overcome for improved climate monitoring and prediction to become reality for the ~600 million population of the IAS region.

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Figure Captions

Figure 1: a) The mean June-July-August-September (JJAS) climatological SST (Reynolds et al. 2007; contoured) with the 28.5°C isotherm (marking the boundary of the WHWP) in bold black line with mean climatological JJAS 925hPa winds (from Climate Forecast System Reanalysis [Saha et al. 2010]), and mean climatological JJAS mean rainfall (from Chen et al. 2002; shaded). b) The orography of the Americas and the bathymetry of the neighboring oceans.

Figure 2: Summertime (JJA) atmospheric teleconnections linked to large minus small AWP in (top panels) observations, (middle panel) AGCM and (bottom panels) simple model experiments (see Lee et al. 2009). Left panels show baroclinic steam function and rotational wind anomalies (750 minus 250hPa and divided by 2) and the right panels show barotropic streamfunction and rotational wind anomalies (750 plus 250mb divided by 2). The observations are based on ERSST3 and NCEP reanalysis, the AGCM results are from Wang et al. (2008), and the simple model results are from Lee et al. (2009). The unit for stream function is 10^{-6} m^2 s^{-1}.

Figure 3: The correlation of the June-October AWP area anomalies (SST from Reynolds et al. 2007) with corresponding a) rainfall anomalies (shaded; rainfall from Chen et al. 2002) and b) mean sea level pressure (MSLP) anomalies (shaded; MSLP from Saha et al. 2010) and regression of June-September AWP area anomalies on corresponding 925hPa wind anomalies (winds from Saha et al. 2010). The significant values at 95% confidence interval according to t-test are contoured and vectors are shown in red.

Figure 4: Linear trend in SST from “super ensemble” consisting of HadlSST1, KaplanSST2, ERSST3b and COBE2 reanalysis of the iCOADS data set. Masked (white) areas are where the sign of the trend in all 4 analysis products does not match; colored areas are average trends where all four products match in sign. Circles show sites where proxy reconstructions have positive (red) centennial trends or insignificant (white) trends. Magnitudes of proxy trends are generally higher than the observational data set. Boxes indicate areas for which Caribbean and Gulf of Mexico time series are calculated (see SM S.1).

Figure 5: The correlation of the June-July-August mean SST anomalies (OISST) with preceding December-February rainfall anomalies (CRU) over equatorial Amazon (outlined). These correlations are computed over 1995-2004 period after the linear trends in rainfall and SST are removed and only significant values at 90% confidence interval are shown. Adapted from Misra and DiNapoli (2013).

Figure 6: Climatological annual mean rainfall (mm/day) computed between 1998 and 2013 from TRMM 3B43 (0.25° resolution) and GPCP 1-degree daily (1DD) data (1° resolution) for four regions in the IAS: a) Caribbean wide (10-25°N, 60-90°W), b) Central America (8-22°N, 83-95°W), c) the Greater Antilles (16-24°N, 65-87°W), and d) the Lesser Antilles (11-20°N, 58-65°W).

Figure 7: Mean climatological SST (°C) in the Caribbean Sea for January-February-March (a) January-February-March, (b) April-May-June, (c) July-August-September, and (d) November-December-October from NOAA Ocean Watch blended SST product (http://oceanwatch.pfeg.noaa.gov/thredds/Satellite/aggregsatBA/ssta/catalog.html). The

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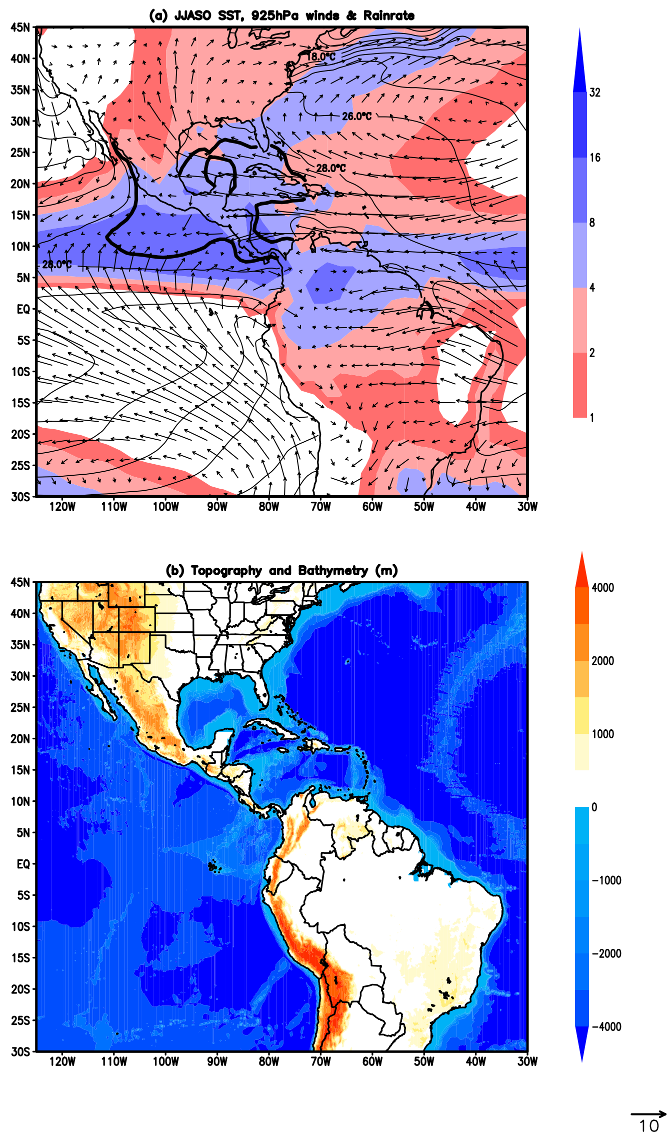
Figure 11: The average monthly Atlantic Warm pool areas (in x10^6 km^2) from 1909–2005 based on various CMIP5 20th century simulations and ERSSTv3 observations.

Each cell in the table is color coded (cool colors indicate a small AWP; warm colors indicate a large AWP) in order to show the average seasonal evolution of the Atlantic Warm Pool's areal extent. Adapted from Kozar and Misra (2012).

Figure 12: Annual mean SST biases in CMIP5 (a) and CMIP3 (b) model ensembles. The biases are referenced to observed Reynolds SST. After Xu et al. 2014.

Fig. 13: Anomaly correlation (AC) skill of SST prediction for seasonal means of June-July-August (JJA). AC is computed based on the NMME hindcasts over the 1981-2010 period. The NMME ensemble initialized in March was used.

Figure 14: Same as Fig. 13 but for seasonal mean June-July-August (JJA) precipitation anomaly.



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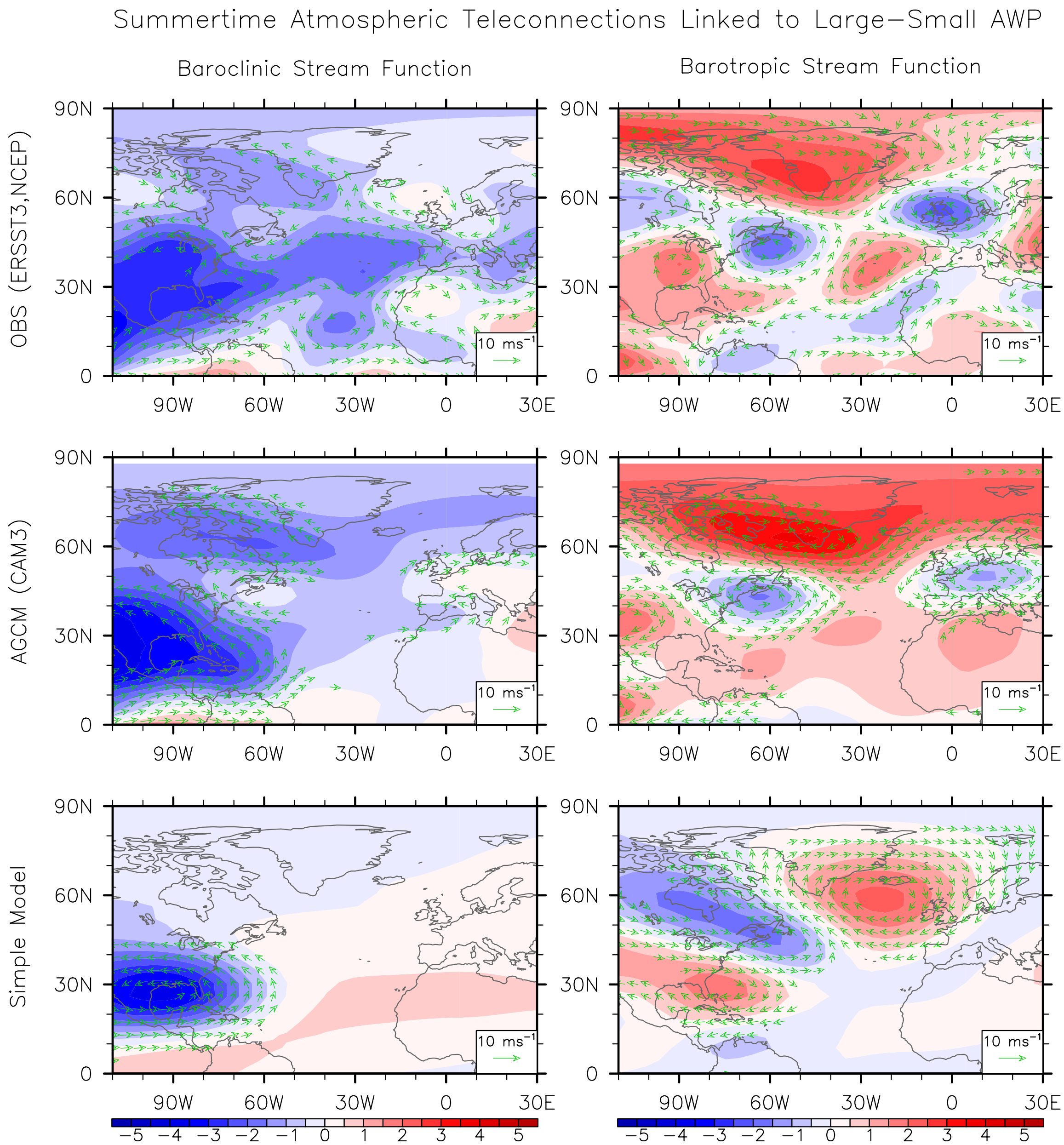

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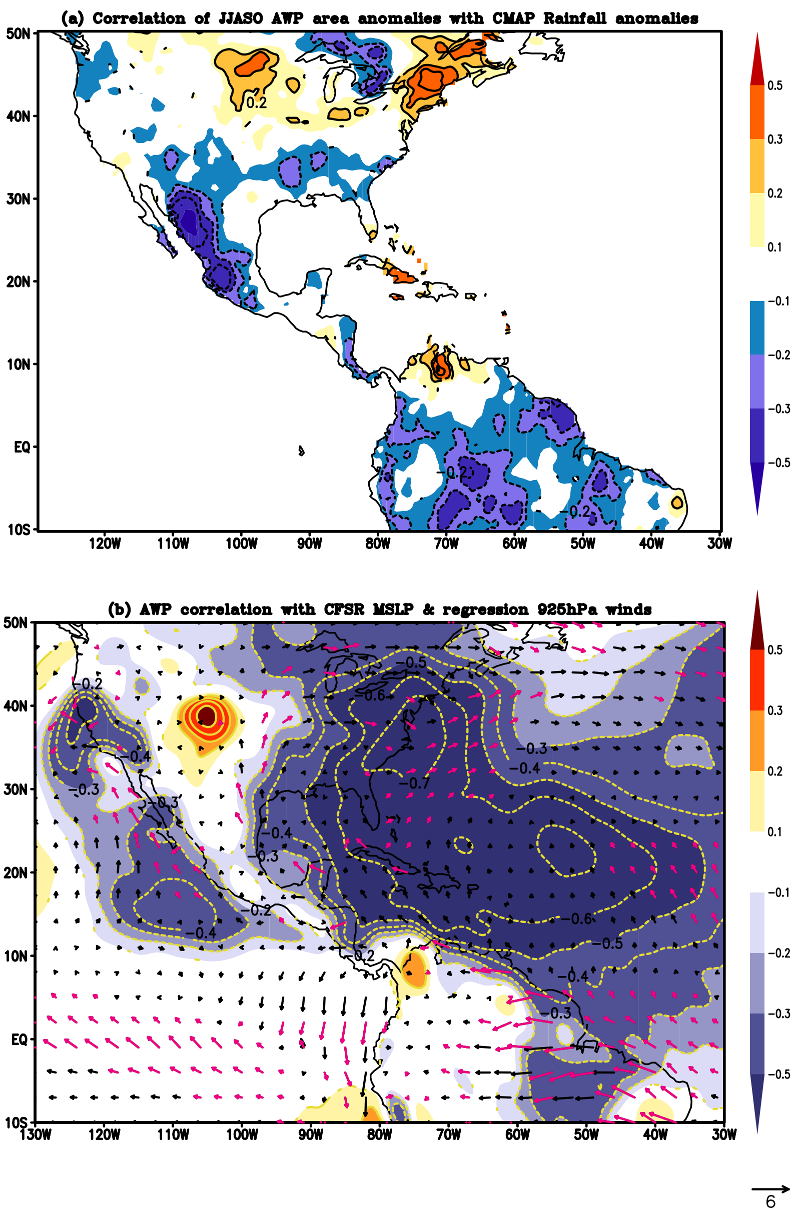

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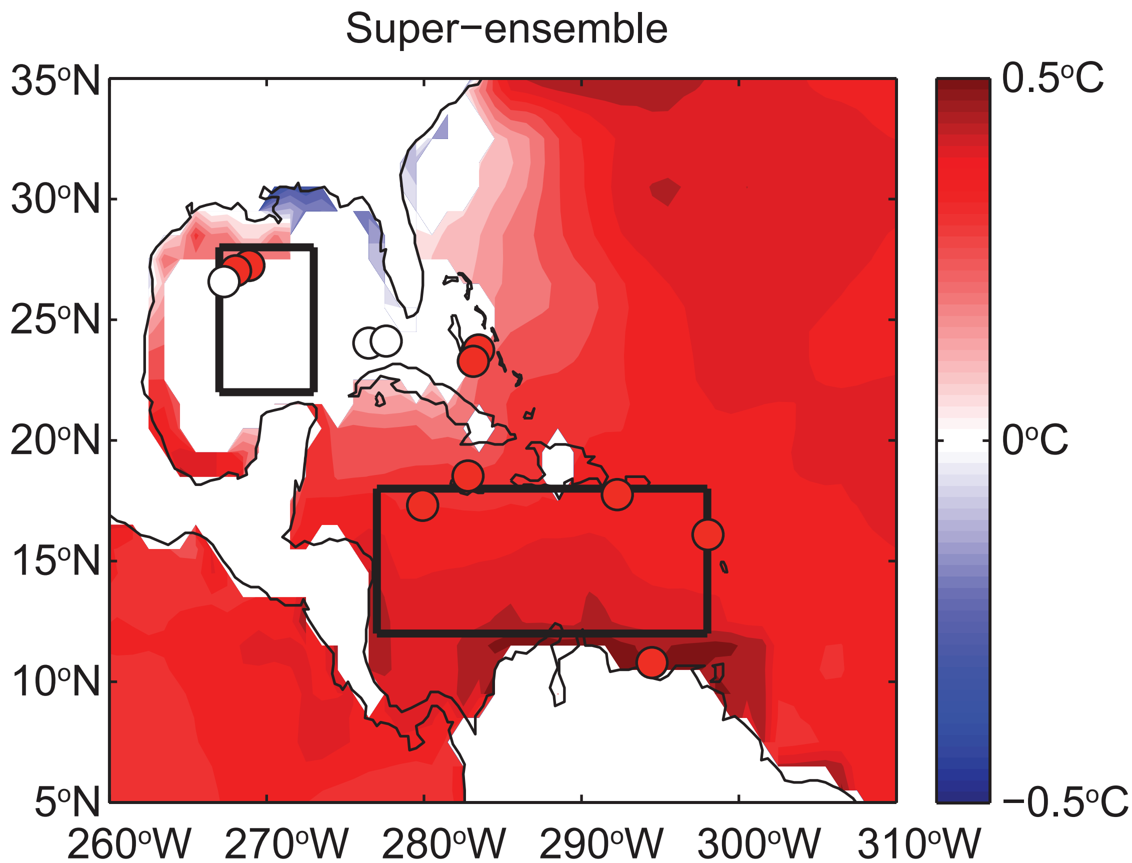

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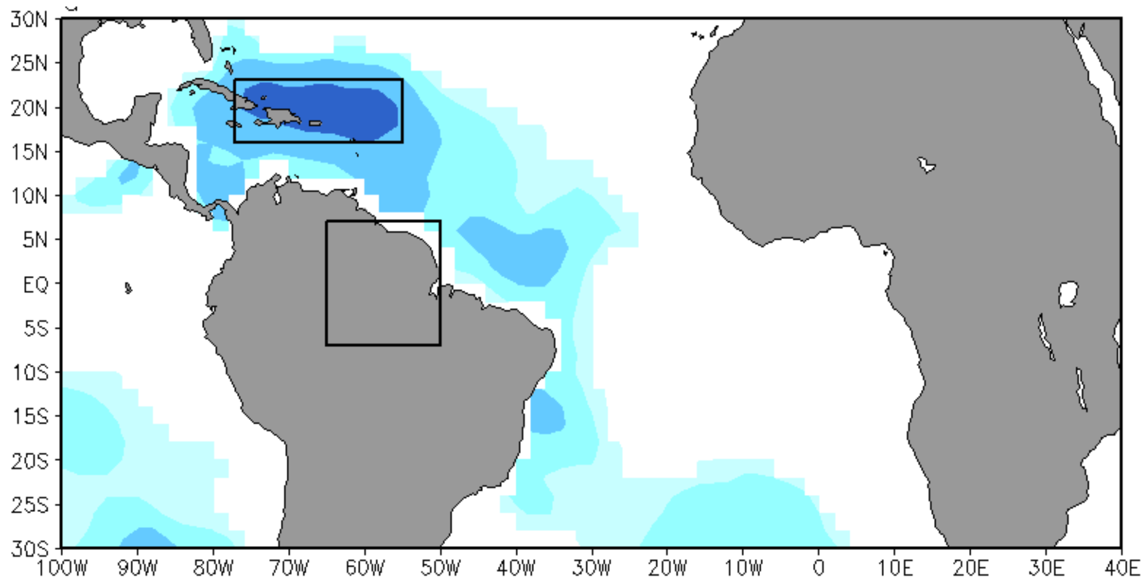

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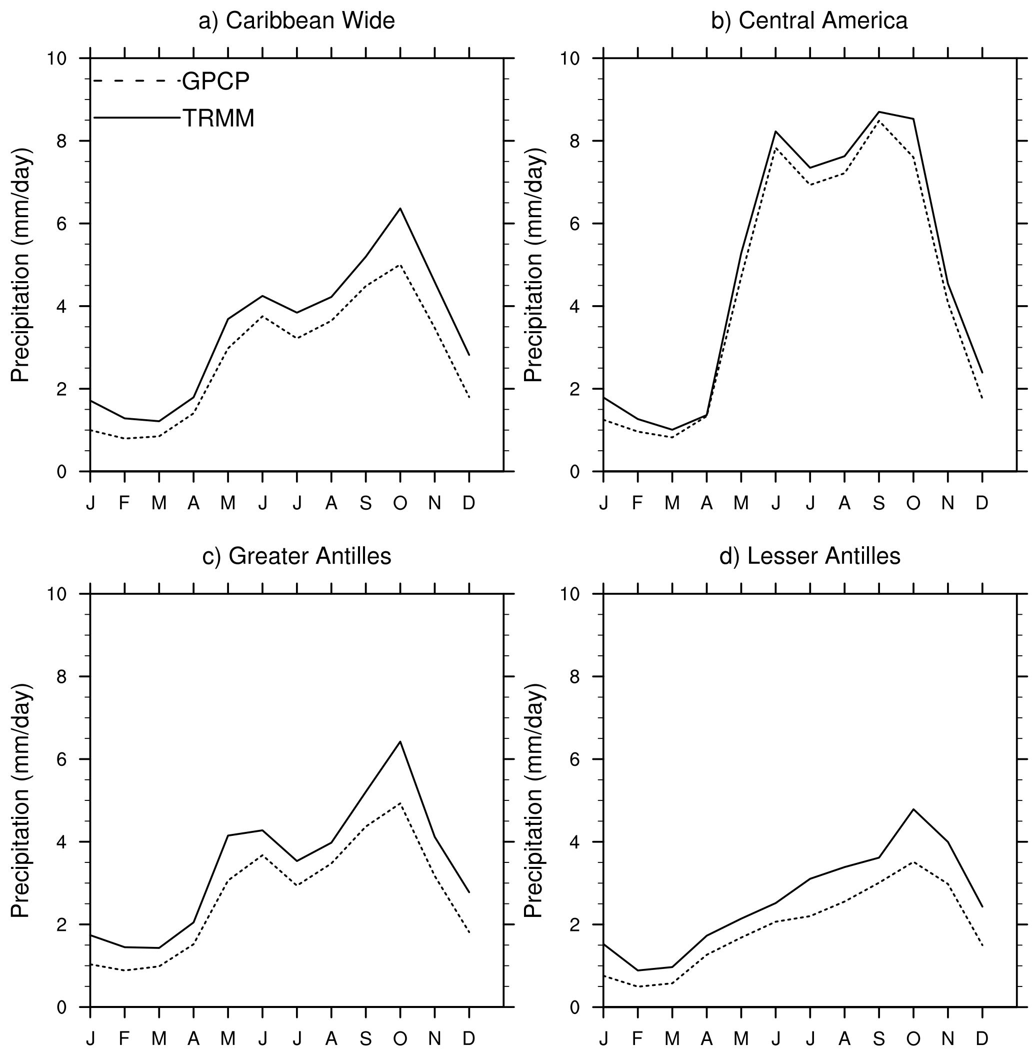

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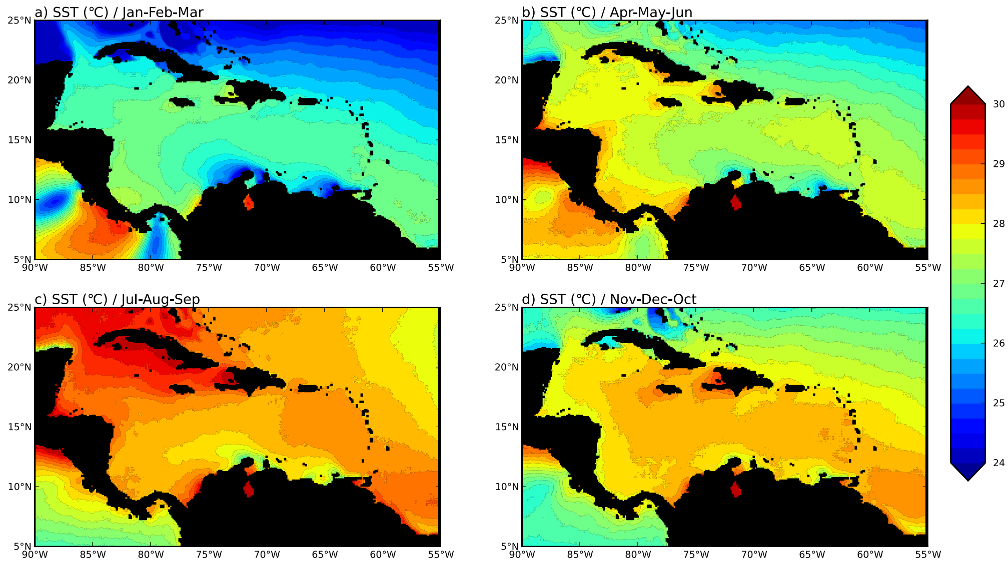

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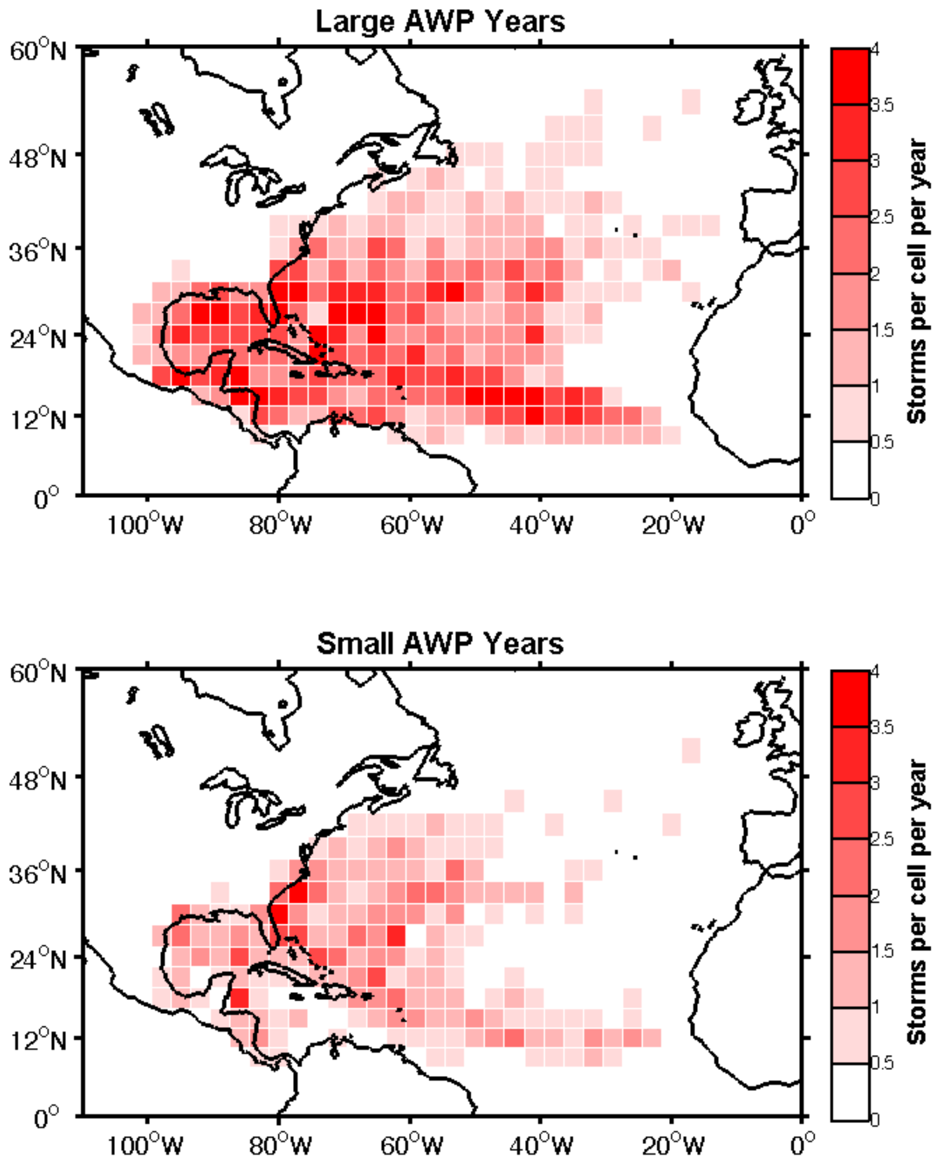

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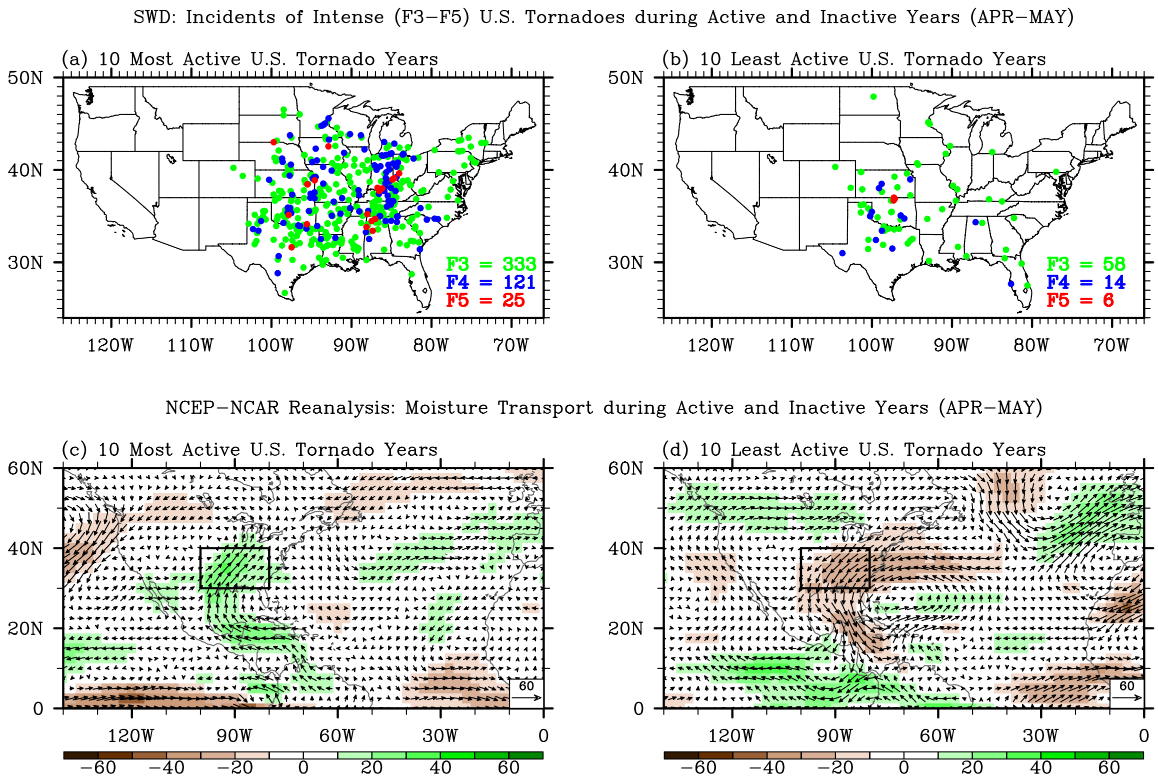

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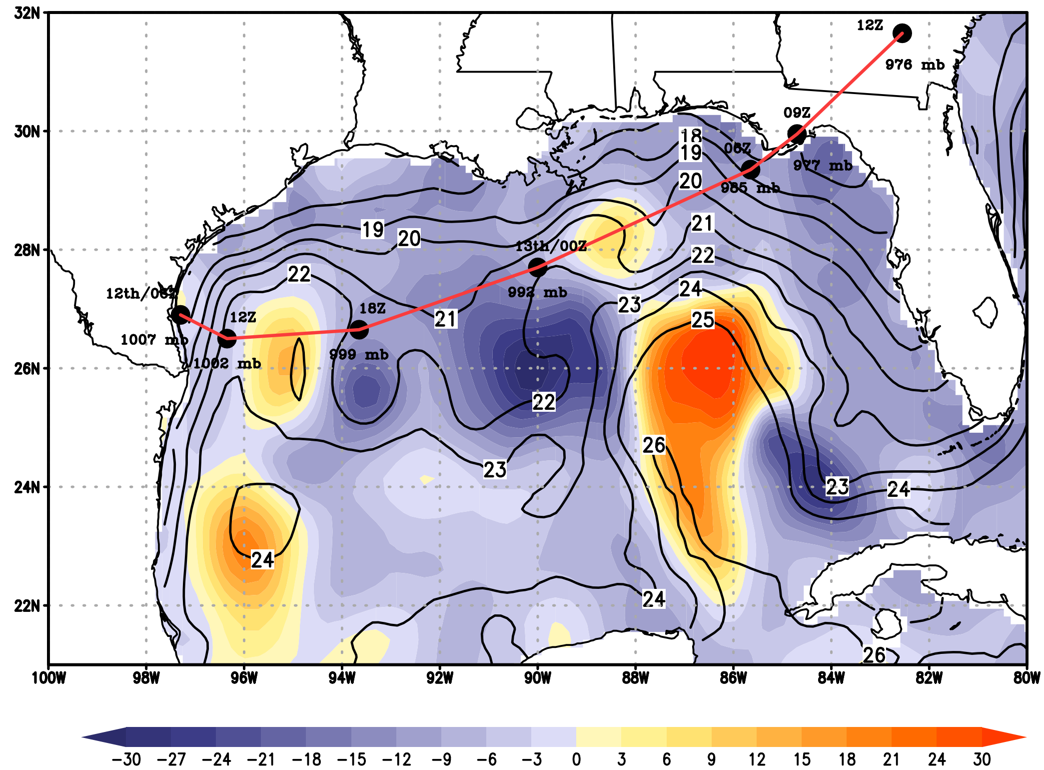

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Seasonal Cycle of AWP Area

Months	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
ERSST v3	0.00	0.00	0.00	0.01	0.05	0.40	2.49	3.71	3.99	1.63	0.07	0.00
BCC-CSM1-1	0.00	0.00	0.00	0.00	0.00	0.01	0.29	0.68	0.38	0.02	0.00	0.00
CanESM2	0.00	0.00	0.00	0.01	0.00	0.03	0.29	0.83	0.78	0.15	0.01	0.00
CCSM4	0.00	0.00	0.00	0.02	0.02	0.04	0.51	1.31	1.12	0.12	0.00	0.00
CNRM-CM5	0.00	0.00	0.00	0.01	0.02	0.07	0.27	0.79	0.85	0.28	0.03	0.01
CSIRO-Mk3.6	0.00	0.00	0.00	0.00	0.04	0.43	2.06	2.99	3.09	1.42	0.26	0.00
GFDL-CM3	0.00	0.00	0.00	0.00	0.00	0.06	0.54	1.18	0.76	0.11	0.00	0.00
GFDL-ESM2G	0.00	0.00	0.00	0.00	0.00	0.01	0.18	0.81	0.81	0.22	0.01	0.00
GFDL-ESM2M	0.00	0.00	0.00	0.00	0.00	0.01	0.14	0.70	0.64	0.26	0.02	0.00
GISS-E2-H	0.00	0.00	0.02	0.09	0.19	0.26	0.31	0.70	1.02	0.90	0.24	0.01
GISS-E2-R	0.00	0.00	0.02	0.16	0.40	0.68	1.73	3.55	3.83	2.85	1.04	0.09
HadGEM2-ES	0.05	0.04	0.06	0.10	0.21	0.33	0.57	1.37	1.48	0.57	0.21	0.10
INM-CM4	0.01	0.00	0.00	0.01	0.02	0.00	0.01	0.06	0.02	0.00	0.00	0.01
IPSL-CM5A-LR	0.00	0.00	0.00	0.00	0.00	0.03	0.07	0.21	0.07	0.01	0.00	0.00
MIROC5	0.00	0.00	0.00	0.02	0.04	0.09	0.11	0.23	0.26	0.12	0.02	0.00
MPI-ESM-LR	0.00	0.00	0.00	0.00	0.00	0.08	0.71	2.05	2.45	1.01	0.14	0.01
MRI-CGCM3	0.00	0.00	0.00	0.00	0.00	0.02	0.12	0.39	0.38	0.18	0.01	0.00
NorESM1-M	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.03	0.02	0.00	0.00	0.00

0.00 0.05 0.10 0.15 0.20 0.30 0.40 0.60 0.80 1.00 1.25 1.50 1.75 2.00 2.50 3.00 3.50 4.00

AREA OF 28.5°C MONTHLY AVERAGED ISOTHERM (10^6 km^2)

Figure 11: The average monthly Atlantic Warm pool areas (in x10^6 km^2) from 1909–2005 based on various CMIP5 20th century simulations and ERSSTv3 observations. Each cell in the table is color coded (cool colors indicate a small AWP; warm colors indicate a large AWP) in order to show the average seasonal evolution of the Atlantic Warm Pool's areal extent. Adapted from Kozar and Misra (2012).

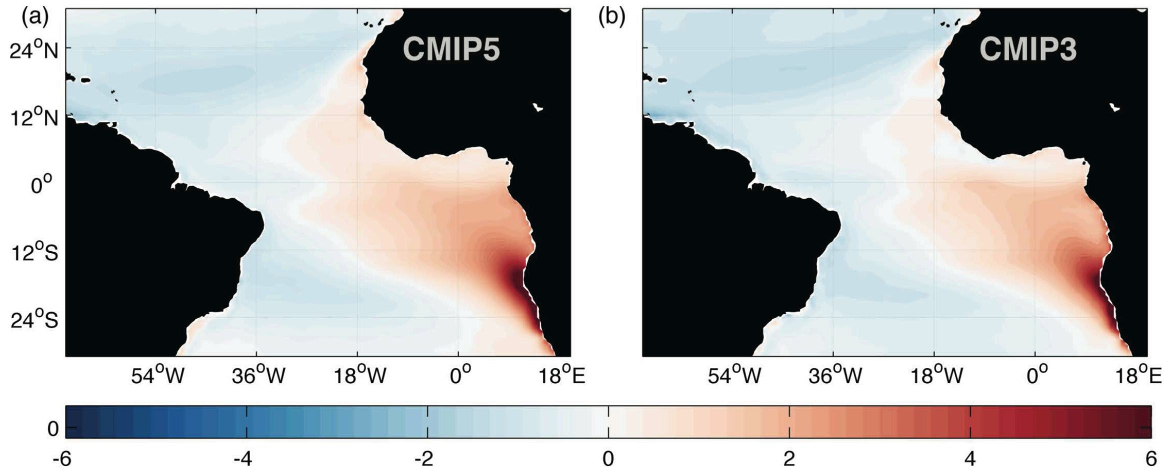

Figure 12: Annual mean SST biases in CMIP5 (a) and CMIP3 (b) model ensembles. The biases are referenced to observed Reynolds SST. After Xu et al. 2014.

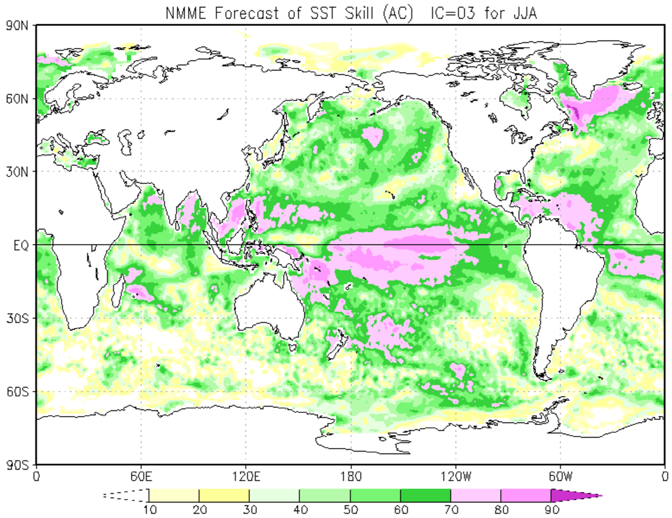

Fig. 13: Anomaly correlation (AC) skill of SST prediction for seasonal means of June-July-August (JJA). AC is computed based on the NMME hindcasts over the 1981-2010 period. The NMME ensemble initialized in March was used.

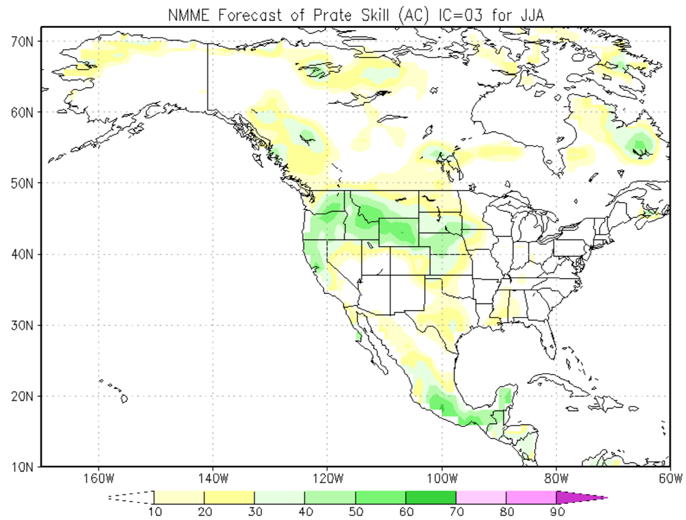

Figure 14: Same as Fig. 13 but for seasonal mean June-July-August (JJA) precipitation anomaly.