

Understanding and Predicting Climate Variability and Change at Regional Scales

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1. Introduction

Better understanding and prediction of climate and its variability at regional and local scales have been a challenge for the scientific community for many decades. This challenge exists not only because of the complexity of the physical processes and interactions that occur on these scales among the different elements of the climate system, but also because of the importance such knowledge implies for society. Particularly difficult are the world's monsoon regions where more than two thirds of the Earth's populations live.

Understanding, simulating and predicting the monsoon involves multiple aspects of the physical climate system (i.e., atmosphere, ocean, land, and cryosphere), including the role of human influences. During the last decades WCRP has promoted the implementation of international research programs, modeling activities and field experiments to improve understanding and modeling of processes that shape the monsoons. The establishment of the WCRP/CLIVAR panel on the Variability of American Monsoon Systems (VAMOS, Mechoso, 2000) contributed to the organization of multinational research on the American Monsoons. VAMOS encouraged the realization of the South American Low Level Jet experiment (SALLJEX, Vera et al., 2006a), the North American Monsoon Experiment (NAME, Higgins et al., 2006), LPB Regional Hydroclimate Project (Berbery et al. 2005), and VOCALS in the southeastern Pacific (Wood and Mechoso, 2008). The West African Monsoon has also received considerable attention through the international African Monsoon Multidisciplinary Analysis (AMMA) program (Redelsperger et al., 2010). Observational campaigns, such as the GEWEX/CEOP (Coordinated Enhanced Observing Period) and the YOTC (Year of Tropical Convection) have archived both in-situ and satellite observation data, providing a continuous record of observations for studies on processes and interactions affecting monsoon variability. WCRP also sponsored the Asian Monsoon Years (AMY 2007-2012). In recent years, other regions of the world have also been the target of several WCRP sponsored programs. For example, the Mediterranean region has been the focus of several initiatives such as MedCLIVAR (Lionello et al. 2006a) and HyMeX (Drobinsky et al. 2010). The areas around the Mediterranean Sea represent a transitional region between mid-latitude and subtropical dynamics, where the northern hemisphere teleconnection patterns, ENSO and the monsoon systems play a role in shaping the regional climate (see Lionello et al 2006b, Ulbrich et al 2011, for a review). The Mediterranean is also a hotspot of climate change (Giorgi 2006, Giorgi and Lionello 2008), where warming, and water availability may have important socio-economic effects in the 21st century.

This paper highlights some of the scientific advances made under WCRP leadership

in understanding climate variability and predictability at regional scales with emphasis in the monsoon regions. The paper further discusses important challenges that the WCRP community must address in the future with respect to modeling regional climate variability and change.

2. Regional perspectives

This section presents a brief description of progress in understanding the different monsoon systems. We focus on the warm season flow over the tropics, and not only address seasonal mean monsoon features, but also monsoon variability on time scales of great societal value, such as intraseasonal, interannual, and longer including climate change. A section on issues related to Europe and the Mediterranean regions is also included for completeness in terms of reviewing regional variability and change in populated land regions, and because numerous studies have already occurred over this region that may be of guidance for future efforts in monsoon regions.

2.1 Asian-Australian Monsoons

Regional Characteristics and Variability

The Asian-Australian (AA) Monsoon exhibits the largest regional extent and strongest remote impacts on the global climate of all monsoon systems. Furthermore, the AA monsoon directly influences the livelihood of by far the greatest number of people in comparison to all others monsoon systems. The AA monsoon can be broadly viewed comprising of two subsystems: the Indian Monsoon and the East Asian Monsoon (e.g. Wang et al. 2006). These subsystems co-vary in response to external influences such as El Niño/Southern Oscillation (ENSO), but also vary independently due to the region's unique topography, land surface conditions, land-atmosphere-ocean interactions, and other internal mechanisms of variability. The Indian monsoon is associated with a cross-equatorial clockwise circulation, centered in the equatorial Indian Ocean that links the Indian monsoon trough, the Somalia jet and the southeasterly trade winds that circulate the Mascarene High south of the equator (Fig. 1). The East Asian monsoon is associated with the large-scale region of convection in the western north Pacific monsoon trough that lies at the confluence of the southeast Pacific trades and the monsoonal southwesterlies to the west. To the north of this trough is the subtropical high and associated subtropical front (the Meiyu, Baiu, or Changma), which brings heavy rainfall to East Asia at the onset of the summer monsoon.

A prominent feature of the AA monsoon is its intraseasonal variation (ISV), consisting of a series of active and break cycles, which typically originate over the western equatorial

Indian Ocean. Enhanced and suppressed convective activity associated with Boreal summer intraseasonal oscillations (BSISO) propagate both poleward over land and eastward over the ocean during the summer monsoon, exhibiting both 10-20 day and 30-50 day modes (Goswami 2005a). During the Australian summer monsoon, ISV is dominated by the Madden Julian Oscillation (MJO) with a periodicity between 30 and 50 days and propagation primarily west-east with only limited poleward influence over subtropical Australia (Wheeler et al. 2009). The amplitude of intraseasonal variations is far greater than interannual variations. The impact of these variations can be dramatic—for example the intraseasonal break in the monsoon over India in July 2002 resulted in only 50% of normal rainfall that month, causing enormous loss of crops and livestock. The ISV is an important building block of the AA monsoon. In one sense it influences predictability of the seasonal mean climate (Goswami and Ajaya Mohan 2001), while also influencing weather predictability through modulating the frequency of occurrence of synoptic events such as lows, depressions and tropical cyclones (Maloney and Hartmann 2000; Goswami et al. 2003; Bessafi and Wheeler 2006). Rainfall in the greater AA monsoon is surprisingly consistent from year to year, reflecting the robust forcing arising from the seasonal land-surface heating. However, even relatively small percentage variations, when set against large seasonal rainfall totals, can have a dramatic impact on society, particularly where agriculture remains the main source of living (Gadgil and Kumar, 2006). Floods are also common disasters in monsoon Asia. Due to the recent growth of Asian economies, flood damage is increasing, particularly in larger cities.

ENSO is the dominant forcing of monsoon interannual variability (IAV). ENSO's warm phase (El Niño) tends to be associated with reduced summer monsoon rainfall, although in the case of the Australian monsoon, the impact of El Niño is stronger in the pre-monsoon season (e.g., Hendon et al. 2011). In addition, antecedent Eurasian snow cover has been reported to contribute to monsoon IAV (e.g. Goswami 2005b) while tropical Atlantic Ocean temperatures have also been associated with variations of the Indian summer monsoon (Rajeevan and Sridhar 2008; Kucharski et al. 2008). The Southern Annular mode (SAM) also influences the Australian summer monsoon through a poleward shift of the Australian anticyclone during SAM positive phase, resulting in stronger easterly winds impinging on eastern Australia, enhancing summer rainfall (Hendon et al. 2007). A large fraction of the AA monsoon IAV is unexplained by known, slowly varying forcing and may be considered 'internal' IAV arising from interactions with extra-tropics (e.g. Krishnan et al. 2009) or scale-interactions within the tropics (e.g. Neena et al. 2011).

Long-term Trends and Projections

Lack of an increasing trend of South Asian monsoon rainfall in the backdrop of a clear increasing trend of surface temperature (Kothawale et al. 2005) has been reconciled as due to contribution from a increasing trend of extreme rainfall events being compensated by contributions from a decreasing trend of low and moderate events (Goswami et al. 2006a). It is also suggested that an increased intensity of short-lived extreme rain events may lead to a decreasing predictability of monsoon weather (Mani et al. 2009)

Future projections based on the Coupled Model Intercomparison Project-3 (CMIP3) show monsoon precipitation is projected to increase in South and East Asia during June-August and over the equatorial regions and parts of eastern Australia in December-February, though model consistency is not high locally, especially for Australia (Fig. 2). Atmospheric circulation changes impact onto projected regional precipitation (Kitoh 2010). For example, in the East Asian summer monsoon, a projected intensification of the Pacific subtropical high, defining the Meiyu-Changma-Baiu frontal zone and the associated moisture flux, may bring about increase rainfall (Kitoh 2010). Most models project an increase in the interannual variability of monthly mean precipitation (Krishna Kumar et al. 2011). The intensity of precipitation events is also projected to increase, with a shift towards an increased frequency of heavy precipitation events (e.g. >50 mm day⁻¹). Changes in extreme precipitation follow the Clausius–Clapeyron constraint and are largely determined by changes in surface temperature and water vapor content (e.g. Turner and Slingo 2009).

2.2 American Monsoon Systems

Regional characteristics and variability

The Americas form a landmass of great meridional extent, stretching from over 50°S to over 70°N. During their respective warm seasons, North and South America both show classical monsoon-type upper-level anticyclone/low-level heat low structures; large-scale convergence zones of strong precipitation with ascent to the east; and descent regions to the west over stratocumulus decks in the Pacific Ocean (e.g. Vera et al., 2006b; Marengo et al. 2010; Liebmann and Mechoso, 2011; and references therein).

The upper-level monsoon anticyclone of the North American Monsoon System (NAMS) shifts northward with season from southwestern Mexico to northwestern Mexico-southwestern United States (U.S.). Precipitation is intense over Mexico, southwest U.S., and the Gulf of Mexico. A relative minimum in precipitation along the Sierra Madre Oriental and the Caribbean is known as the mid-summer drought (MSD). Moisture transport onto the North

American continent is associated with broad-scale advection from the Gulf of Mexico and with low-level jets (LLJs) over the Gulf of California and east of the Rockies. The latter LLJ is a primarily warm season, nocturnal, feature; while less is known about the former. A prominent region for tropical cyclone development occurs just west of North America. Thus, much of the extreme precipitation in NAMS is associated with tropical cyclones occurring during the core of the monsoon season (e.g. Cavazos et al., 2008).

In the South American Monsoon System (SAMS), precipitation is intense over the Amazon region, central Brazil and Bolivia, extending northeast into the Atlantic ITCZ and southeast into the South Atlantic Convergence Zone (SACZ). The main regional circulation features are the upper-level anticyclone ("Bolivian High") and the low-level heat low in subtropical South America ("Chaco Low"). Low-level trade winds from the tropical Atlantic provide a moisture source for the SAMS. The South American LLJ (SALLJ) is present throughout the year along the eastern scarp of the Andes with strongest winds over Bolivia (e.g., Campetella and Vera, 2002, Marengo et al. 2004). Convergence of moisture, advected by the SALLJ from the Amazon Basin over the La Plata Basin (LPB) in southeastern South America, results in some of the most frequent and largest mesoscale convective systems (MCS) on earth (Laing and Fritsch, 2000).

During the warm season, the MJO modulates a number of weather phenomena affecting the NAMS (e.g. tropical cyclones, tropical easterly waves, Gulf of California surges) (Barlow and Salstein 2006; Yu et al. 2011). Intraseasonal (and even interannual and interdecadal) variations of SAMS appear to be dominated by a continental-scale eddy centered over eastern subtropical South America (e.g. Robertson and Mechoso, 2000, Zamboni et al. 2011). In the cyclonic phase of this eddy, the SACZ intensifies and precipitation weakens to the south, resembling a dipole-like structure in the precipitation anomalies; the anticyclonic phase (Fig. 3b) shows opposite characteristics (e.g. Nogues-Paegle and Mo, 1997, Nogues-Paegle and Mo 2002, Ma et al. 2007). Such an anomaly dipole pattern (Fig. 3a) is influenced on intraseasonal timescales by the MJO (e.g. Liebmann et al. 2004) and, on interannual timescales, by both ENSO (Nogues-Paegle and Mo, 2002) and surface conditions in the southwestern Atlantic (Doyle and Barros 2002).

El Niño and La Niña tend to be associated with anomalously dry and wet events, respectively, in the equatorial belt of both NAMS and SAMS. ENSO influences NAMS and SAMS activity through changes in the Walker/Hadley circulations of the eastern Pacific and through extratropical teleconnections extended across both the North and South Pacific Oceans (PNA and PSA, respectively). During austral spring, climate variability in southeastern South America

is influenced by combined activity of ENSO (Grimm et al. 2000) and SAM (Silvestri and Vera 2003).

Land surface processes and land use changes can significantly impact both NAMS and SAMS. The continental-scale pattern of NAMS IAV shows anomalously wet (dry) summers in the southwest U.S. are accompanied by dry (wet) summers in the Great Plains of North America. Stronger and weaker NAMS episodes often follow northern winters characterized by dry (wet) conditions in the southwest U.S. Moreover, spring precipitation anomalies in the SAMS region induce soil moisture and near surface temperature anomalies, which in turn alter surface pressure and wind divergence anomalies during the following summer (Grimm et al. 2007).

On longer time scales, river streamflow records exhibit an influence of the Pacific Decadal Oscillation (PDO) on precipitation in both the NAMS (Brito-Castillo et al. 2003; Englehart and Douglas 2006, 2010) and SAMS (e.g. Robertson and Mechoso, 2000; Zhou and Lau, 2001, Marengo 2009) regions. Warm PDO phase tends to have dry (wet) El Niño and wet (dry) La Niña summers in North America (southern South America) (Englehart and Douglas 2006, Kayano and Andreoli, 2007). The North Atlantic Oscillation and, the Atlantic Multi-decadal Oscillation can also influence the American Monsoons (Hu and Feng 2008; Chiessi et al., 2009), while decadal changes in the SAM influence on precipitation anomalies in southeastern South America have also been recorded (Silvestri and Vera 2009).

Long-term trends and projections

Between 1943 and 2002, NAMS onset has become increasingly later and the NAMS more erratic in its rainfall, though the absolute intensity of rainfall has been increasing (Englehart and Douglas, 2006). In the NAMS core region, daily precipitation extremes have shown significant positive trends during the second half of the twentieth century (e.g. Arriaga-Ramirez and Cavazos, 2010), while consecutive dry days with periods longer than one month have significantly increased in the U.S. southwest (Groisman and Knight, 2008). Positive trends in warm season mean and extreme rainfall have been documented in southeastern South America during the twentieth century (e.g. Marengo et al., 2009; Re and Barros, 2009).

Climate change scenarios for the 21st century show a weakening of the NAMS, through a weakening and poleward expansion of the Hadley cell (Lu et al., 2007). Projected changes in ENSO have, however, substantial uncertainty with regard to the hydrological cycle of the NAMS (Meehl et al., 2007). Changes in daily precipitation extremes in the NAMS have inconsistent or no signal of future change (e.g. Tebaldi et al. 2006). The majority of CMIP3

models project positive trends in summer precipitation for the 21st century over southeastern South America (e.g. Vera et al. 2006c). That trend has been recently related to changes in the activity of the dipolar leading pattern of precipitation IAV (Junquas et al. 2011). In addition, a weak positive trend in the frequency of daily rainfall extremes has been projected in the southeastern part of SAMS by the end of the 21st century, associated with more frequent/intense SALLJ events (e.g. Soares and Marengo, 2009).

2.3 Sub-Saharan Africa

Regional characteristics

The West African monsoon region (WAM) is characterized by summer rainfall over the continent and winter drought. In the Gulf of Guinea, equatorial cooling between May and June, associated with the development of the Atlantic cold tongue (ACT), is influential in determining both the onset and intensity of the rainy season in coastal regions (e.g. Gu and Adler, 2004, Thorncroft, et al, 2011, Nguyen et al., 2011) and the onset of the rainy season over the Sahel (e.g. Sultan and Janicot, 2003). A strong relationship exists between April-May sea surface temperature (SST) in the Angola Benguela frontal zone, to the south of the ACT, the southeasterly trades, and July-August rainfall over coastal West Africa (Reason and Rouault, 2006). From March to mid-June, the ACT results from the intensification of southeasterly trades, while during the second phase (mid-June–August), wind speeds north of the equator increase as a result of the northward progression of the intensifying trades and as a result of significant surface heat flux gradients, produced by differential cooling between the ACT and tropical waters in the Gulf of Guinea (Caniaux et al. 2011).

The Saharan heat low (SHL) is a major feature of the WAM (e.g. Thorncroft and Blackburn, 1999, Parker et al, 2005, Lavayasse et al, 2009), characterized by a deep, dry, well-mixed boundary layer (see the red dome in Fig. 4). The SHL pressure minimum on the southern flank of the Sahara is associated with the establishment of two opposing low-level flows along the InterTropical Discontinuity (ITD, dashed blue line in Fig.4): dry and hot northerly harmattan air and moist, cooler monsoon air (Lafore et al., 2010). Associated low-level baroclinicity acts to establish the midlevel African easterly jet (AEJ, yellow tube in Fig. 4). The AEJ is a key component of the WAM, providing conditions for the growth of key weather systems in the region including African easterly waves and MCSs. Variations in topography are also important for MCS generation (e.g. Laing et al, 2008). Another important WAM feature is the low-level westerly jet off the West African coast (e.g. Pu and Cook, 2010).

Equatorial East Africa receives rainfall mainly during October-December (short rains) and March-May (long rains). Some areas of the Congo basin also experience bimodal rainfall seasons, while austral summer is the main rainy season of the southern part of this basin. Due to the lack of reliable data in the Congo, far less research has been done on this area than for most other parts of Africa and its climate variability is not well understood.

Three rather unique ocean circulation features are thought to be important for southern African climate; namely, the Angola Benguela Frontal Zone (ABFZ), the Seychelles Chagos thermocline ridge (SCTR) and the Agulhas Current and its retroflexion. The circulation and properties of the Agulhas current and its significance for southern African climate variability has been addressed in several studies (e.g., Lutjeharms, 2006, Walker and Mey, 1988; Jury et al., 1993). Moreover, most of South Africa's flooding episodes are associated with cut-off lows crossing the Agulhas Current, inducing flow into the coastal mountains and flooding (e.g., Singleton and Reason, 2007). Evidence of MCS intensification in northeastern South Africa as a result of the low-level moisture inflow from the Agulhas Current has also been documented (Blamey and Reason, 2009).

The SCTR is located to the northeast of Madagascar and features a pronounced shallowing of the thermocline that is sensitive to Rossby wave signals from the tropical eastern Indian Ocean (Hermes and Reason, 2008, Schott et al., 2009). Hermes and Reason (2009) showed that small shifts in the position or strength of the South Indian Ocean anticyclone led to substantial changes in the depth and position of the SCTR. In summers with a deeper (shallower) SCTR, the number of tropical cyclone days in the South West Indian Ocean increases (decreases) (Xie et al., 2002) with impacts on rainfall over Madagascar and eastern Africa. The SCTR also influences the Indian and East African monsoons and Madden Julian Oscillation activity (Annamalai et al., 2005; Schott et al., 2009).

The ABFZ separates the cool water of the Benguela upwelling system to the south from the warm Angola Current waters further north. The ABFZ acts to some extent as an ecosystem and climate 'barrier'. South of the ABFZ, the coastal atmosphere is very stable with the adjacent Namib Desert being one of the driest regions of the world whereas, to the north, conditions are more favorable for rainfall. Thus, during the summer half of the year, warm moist air emanating from north of the ABFZ feeds into synoptic cloudbands that bring much of subtropical southern Africa's rainfall (Cook et al., 2004, Reason et al. 2006).

Climate Variability and Change

WAM rainfall varies intraseasonally with two distinct periods: 10-25 and 25-90 days (Sultan et al, 2003, Matthews, 2004, Lavender et al, 2009, and Janicot et al. 2010). Variability of rainfall in the 25-90 day range appears to have a significant MJO contribution but its impact is not straightforward, possibly arising in association with a westward propagating Rossby wave signal that can be equatorial or sub-tropical (e.g. Janicot et al, 2009, Ventrice et al, 2011) as well as eastward propagating Kelvin waves (e.g. Matthews, 2004). Variability of rainfall in the 10-25 day range has been associated with two types of evolution – a “quasi-biweekly-zonal-dipole mode” that includes a notable eastward propagating signal between Central America and West Africa, likely involving convectively coupled Kelvin waves (Mounier et al, 2008), and a “Sahelian mode” that includes a westward propagating signal in the Sahelian region (Mounier and Janicot, 2004) and shares some characteristics with equatorial Rossby waves.

Recently, the importance of SST IAV in the Atlantic, Pacific –Indian and the Mediterranean basins on the WAM has been confirmed (e.g. Losada et al, 2009, Mohino et al, 2010, 2011, Rodriguez-Fonseca et al, 2010). It has also been suggested that vegetation IAV (affected by the previous year’s rainy season) influences the early stages of the following rainy season, although models have difficulty in reproducing this behavior (Philippon et al, 2005). In equatorial East Africa, studies show the short rains over this region are strongly sensitive to ENSO (eg Ogallo, 1988; Hastenrath et al., 1993) and to the Indian Ocean Dipole (eg Saji et al, 1999; Webster et al., 1999). One of the strongest SST-rainfall correlations anywhere on the African continent exists between East African rainfall and tropical Indian Ocean SST in October-November-December. ENSO also exerts a strong influence on summer rainfall over southern Africa. Teleconnections between the North Atlantic Oscillation (NAO) and austral autumn Congo River discharge and regional rainfall have also been documented (Todd and Washington, 2004). In general, warm (cool) SST anomalies east of South Africa are associated with above (below) average summer rainfall over southeastern Africa (Reason and Mulenga, 1999). The South Indian Ocean SST dipole, which influences summer rainfall over southern Africa (Behera and Yamagata, 2001; Reason, 2001, 2002), has its southwestern pole in the greater Agulhas Current region. In addition, warm (cold) events in the ABFZ region during summer/autumn, not only disrupt fisheries but also often produce large positive (negative) rainfall anomalies along the Angolan and Namibian coasts and inland (Rouault et al., 2003). The generation mechanism for these Benguela Niños (Niñas) seems to be related to tradewind weakening (strengthening) in the western equatorial Atlantic (Florenchie et al., 2003, 2004) and by implication in the South Atlantic Anticyclone. The interior rainfall influences depend to

some extent on concurrent SST anomalies in the South Indian Ocean (Hansingo and Reason 2009), while a teleconnection between the SAM, tropical southeast Atlantic SST and central / southern African rainfall has also been identified (Grimm and Reason, 2011). On the other hand, local re-circulation of moisture (e.g., Cook et al., 2004), and land surface feedbacks (e.g., Mackellar et al, 2010) can also contribute to climate variability in southern Africa.

Multi-decadal SST variability in both the Atlantic and Pacific has been shown to be important for the WAM (e.g. Rodriguez-Fonseca et al, 2011). The partial recovery in West African rainfall over the past decade has received substantial debate over the respective roles of the Atlantic and Indian basins (e.g. Giannini et al, 2003, Knight et al, 2006, Hagos and Cook, 2008, Mohino et al, 2011). There is also evidence that the Agulhas Current itself has strengthened and warmed in recent decades (Rouault et al., 2009). Moreover, warming over the broader Indian Ocean has been linked to a tendency towards drier conditions in southern Africa over the last few decades (Hoerling et al., 2006). Regarding change signals, for Africa as a whole Kniveton et al. (2009) suggests historically there is an increasing delay in wet season onset, while in east Africa Funk et al (2008) find a historical reduction in precipitation which contradicts some of the climate projections for the region. At more local scales, in southern Nigeria, Chineke et al. (2010) note that the mid-summer break which gives rise to the bimodal rainfall season has shown significant reduction in duration from the historical 2-3 week length.

For the projections of the future, WCRP/CMIP3 models failed to show agreement on changes of West African rainfall in the 21st century projections (Biasutti and Giannini, 2006, Christensen et al., 2007, Joly et al, 2007). However, precipitation changes derived from empirical downscaling applied to GCM projection ensemble, has shown greater agreement in an increased precipitation along the southern Africa coast, widespread increase in late summer precipitation across south-east Africa, reduced precipitation in the interior, and a less spatially coherent early summer decrease (Hewitson and Crane, 2006, Tadross et al., 2009). In general across southern Africa there are indications of drying in the west and wetter condition in the east (Hewitson and Crane, 2006; Gianini et al., 2008; Batisani and Yarnal, 2010). Hewitson and Crane (2006) further note that the interplay between change in derivative aspects of rainfall (such as increasing intensity but reducing frequency) can be masked in the more common representation of seasonal averages.

2.4 Climate variability and change over Europe and the Mediterranean region

European climate is strongly influenced by large scale modes of variability over the North Atlantic and Arctic, the NAO being the most important (e.g. Hurrell 1995). Positive NAO

is associated with relatively warm and moist conditions over North West Europe and cold/dry anomalies over southern Europe (Scaife et al. 2008). Negative NAO is associated with cold winters over much of Europe (Hurrell 1995). The phase of the NAO varies on both inter-annual and decadal timescales (Hurrell and van Loon 1999, Stephenson et al. 2006), imparting corresponding variability on European climate. Modeling and observational studies indicate Atlantic Ocean variability may play an important role in NAO variability (Sutton and Allen 1997, Rodwell et al. 1999) suggesting a predictable component to the NAO and thereby European climate (Knight et al. 2005). A number of studies indicate significant changes in the frequency, intensity and geographic location of the NAO over the past decades, towards a persistent positive phase, which may explain a significant fraction of the observed warming over the Northern Hemisphere (Hurrell and van Loon 1999, Thompson and Wallace 2001), with potentially important impacts also on the frequency and intensity of extreme weather events over Europe (Scaife et al. 2008). Attribution of these changes to GHG forcing or natural variability remains unresolved (Gillett et al. 2003), partly due to the inability of GCMs to reproduce observed NAO trends (Osborn et al. 2004, Cohen et al. 2005).

The Mediterranean region is particularly sensitive to climate variability and change (Lionello et al 2006b). Beside NAO, other northern hemisphere teleconnections exert important influences on the climate of the Mediterranean (see Ulbrich et al 2011). Further, influences of ENSO, the Indian Ocean and, to a lesser degree, the WAM have also been detected (Ulbrich et al 2011, Xoplaki et al 2011). The Mediterranean region includes a large contrast in precipitation regimes, with abundant water in some areas and scarcity and irregularity over other regions (mainly in southern and eastern areas). Present trends include widespread warming and significant reduction of precipitation in some areas (Lionello et al, 2011). The observed temperature increase has been found to be consistent with the model climate change signal (Barkhordarian et al 2011). No equivalent result is available for precipitation. Models fail to reproduce the observed decrease of winter precipitation that has characterized the last decades of the 20th century over the Mediterranean (Giorgi and Lionello 2008).

A number of consistent climate change signals have been documented over Europe. These signals, which become statistically robust towards the latter half/third of the 21st century, include a clear trend to warmer, wetter winters over Northern Europe (Kjellström et al. 2011), with an early snow melt season and the near disappearance of extreme cold conditions by the end of the 21st century (Nikulin et al. 2011). Southern Europe will experience warmer and drier summers (e.g. Beniston et al. 2007), with an accompanying risk of increased

heat waves (Schär et al. 2004) and droughts (Lorenz et al. 2010), significant drying of soil in summer (van den Hurk et al. 2005) and potentially negative impacts on agriculture, water availability, natural vegetation and air quality (Fischer and Schär 2010). Precipitation variability is expected to increase across all parts of Europe, particularly in the summer season, with a consistent signal of absolute increases in extreme precipitation magnitude and a shorter return period for present-climate extreme rainfall events (Nikulin et al 2011).

Global models agree in projecting reduced precipitation for the Mediterranean region, increased temperature, and increased interannual variability in both (Giorgi and Lionello 2008), making the Mediterranean one of the most vulnerable regions to climate change globally (Giorgi 2006).

3. Regional climate simulation

3.1 Regionalization needs

Access to quality-controlled high-resolution, regional climate data is key for assessing regional climate vulnerability, impacts and the subsequent development of informed adaptation strategies. Currently, CGCMs used for seasonal to decadal prediction and climate change projection typically employ horizontal resolutions of $\sim 1\text{-}2^\circ$. This limits their ability to represent important local forcing, such as complex topography, surface heterogeneity, and coastal and regional water bodies, all of which modulate the large-scale climate on local scales. Coarse resolution also limits the ability of CGCMs to simulate extreme weather events that contribute non-linearly to the societal impact of regional climate variability. To increase the utility of GCM simulations some form of downscaling or regionalization is usually applied to increase the spatial detail of the simulated data.

Regionalization techniques currently include (i) Dynamical downscaling (DD), where a Regional Climate Model (RCM) is run at increased resolution over a limited area, forced at the boundaries by GCM data (Giorgi 1999), (ii) Global Variable Resolution Models (GVAR), that employ a telescoping procedure to locally increase model resolution over a limited area within a continuous AGCM (Deque and Piedelievre 1995) and (iii) Empirical-Statistical Downscaling (ESD), where statistical relationships, developed between observed large-scale predictors and local scale predictands, are applied to GCM output (Hewitson and Crane 1996). Most of these techniques aim to add regional detail without changing the large-scale climate derived from the GCM. All regionalization methods are, to a first-order, dependent on the quality of the large-scale climate simulated by the driving GCM.

3.2 Coordinated downscaling exercises

Several large-scale efforts have been pursued to assess regional climate change. Over Europe, MERCURE (Christensen et al. 1997, Jones et al. 1997), PRUDENCE (Christensen et al. 2007) and ENSEMBLES (van der Linden and Mitchell 2009) are all examples, where multiple GCM projections, most forced by different SRES GHG scenarios, have been used to drive a matrix of European RCMs. The ensemble of RCMs in an attempt to sample a fraction of the uncertainty space associated with projecting regional climate change. Similar efforts over North America have occurred in NARCCAP (Mearns et al. 2009, 2011) and over South America in CLARIS (Menendez et al. 2010) and other regional projects (Marengo et al 2009, 2011). A number of coordinated RCM projects have focused on specific regional phenomena, such as North American summer season precipitation (PIRCS, Takle et al. 1999, Anderson et al. 2003), WAMME and the west African monsoon (Druyan et al. 2010), R-MIP for East Asia (Fu et al. 2005) and ARCMIP (Wyser et al. 2007), for Arctic clouds, turbulence and radiation processes. A set of simulations has also been performed for the Mediterranean region (Gualdi et al 2011). The GEWEX-sponsored ICTS project (Takle et al. 2007) investigated the transferability of RCMs across a range of different regions using unmodified model formulations. Over the past ~15 years such activities, many sponsored by WCRP, have provided detailed knowledge of the RCMs' ability to simulate important regional climate processes and climate change.

In 2008 the WCRP initiated the Coordinated Regional Downscaling Experiment (CORDEX), with the intention to i) provide a coordinated framework for the development, and evaluation of accepted downscaling methodologies; (ii) generate an ensemble of high-resolution, regional climate projections for all land-regions, through downscaling of CMIP5 projections; (iii) make these projections available to climate researchers and the impact-adaptation-vulnerability (IAV) community and support the use of such data in IAV activities; and (iv) foster international collaboration in regional climate science, with an emphasis on increasing the capacity of developing nations to generate and utilize climate data local to their region. CORDEX is an unprecedented opportunity for scientists to collaborate in order to evaluate and improve downscaling methods for different regions of the world and to engage more closely with users of this data (Giorgi et al. (2009), Jones et al (2011)).

CORDEX has defined a set of target domains along with a standard resolution for regional data of 50km. The evaluation phase of CORDEX entails downscaling global reanalysis data for the past 20 years over all regions for which a group plans to generate downscaled future projections (e.g. Africa, South America, Europe, etc). For each CORDEX area, evaluation teams have been established to define key climate processes and metrics of performance

pertinent to that region, in order to make a detailed evaluation of downscaling methods for the recent past. Subsequent to this, DD and ESD methods will be applied to CMIP5 projections for the same regions. 1950-2010 will be available for evaluation while 2010-2100 constitutes the time period over which regional projections will be made. While each of the CORDEX regions will be targeted by groups local to the region, the international downscaling community has agreed to target Africa as a common domain for the coming few years, with an aim of generating an ensemble of climate projections for Africa to support the Intergovernmental Panel on Climate Change (IPCC) 5th Assessment process.

4. Challenges in monsoon simulation

Although there has been substantial progress in understanding and simulating regional climate as a result of projects promoted by WCRP, the successful prediction and simulation of the monsoon and surrounding subtropical regions remains elusive.

Regarding the AA monsoon system, seasonal prediction of land-based seasonal rainfall with the most modern dynamical coupled models such as those that contributed to APCC CliPAS (Wang et al. 2009) and ENSEMBLES (e.g. Rajeevan et al. 2011) remains too low to be of practical use, even at the shortest lead times. Poor seasonal predictions of the AA monsoon be related to the poor representation of land surface processes in the models and uncertainty of initial conditions over the land, but it also stems from local air-sea interaction in the surrounding oceans that tends to damp ocean-atmosphere variability in regions of monsoonal westerlies (Hendon et al. 2011). Regarding the American monsoons, current GCMs are able to capture their large-scale circulation features. However, models still have difficulty in producing realistic simulations of the statistics of regional precipitation and their modulation by the large-scale circulation (Wang et al., 2005; Marengo et al., 2010, 2011). Model limitations are more evident with the intensity of the MSD and SACZ, timing of monsoon onset and withdrawal, diurnal cycle, and in regions of complex terrain (e.g. Gutzler et al., 2003; Ma and Mechoso, 2007).

Limiting factors to improving simulation of the Earth's monsoon systems include the inability to adequately resolve multi-scale interactions that contribute to the maintenance of those systems (Sperber and Yasunari, 2006). A discussion of some common key processes requiring improved simulation in the different monsoon systems is presented in the following subsections.

4.1 Large to Regional Scale processes influencing monsoon variability

A prerequisite for a successful simulation of regional-scale monsoon variability is an accurate representation of large-scale modes of variability (e.g. MJO, ENSO, PDO, AMO). While CGCMs are improving in their ability to simulate such modes, capturing their remote impact on monsoon variability requires CGCMs to also simulate atmospheric and oceanic teleconnections from the mode source regions into the monsoon regions (Alexander et al., 2002). For example, there is some evidence that models that are better able to simulate the seasonal mean climate in the AA monsoon system tend to perform better in predictions of intraseasonal activity, in agreement with the notion that the seasonal mean is related to intraseasonal activity over the season, particularly in the AA monsoon region (Sperber et al. 2000, Goswami et al. 2006b, Kim et al. 2008). Moreover, accurate MJO activity forecasts could be expected to lead to significant improvements in the skill of warm season precipitation forecasts in the tropical America sector (e.g., Jones and Schemm, 2000). On seasonal to interannual time scales, CGCM skill in predicting seasonal mean precipitation in both NAMS and SAMS core domains are low and consistent with a weak ENSO impact. On the other hand, north and south of the SAMS core region, higher predictability can be attributed to increased ENSO impacts (Marengo et al., 2003). In addition, the importance of the NAO/AO and its influence on the North Atlantic storm track (Ulbrich and Christoph 1999), allied to the importance of other large scale circulation features, such as persistent blocking patterns (Trigo et al. 2004), emphasize the fundamental role large scale atmospheric circulation plays in determining European climate.

A number of phenomena that directly impact the quality of simulated monsoon climates resulting from DD can be considered both a large-scale feature (in DD prescribed as a boundary condition) as well as internal (to the DD limited-area and thus open for improved simulation through increased resolution or improved process description). An example is the MJO that can propagate into, and out of, a high-resolution limited-area, while locally influencing intra-seasonal monsoon variability. While dynamical prediction of the MJO has improved in recent years (e.g. Rashid et al. 2009, Kang and Kim 2010, Gottschalck et al. 2010), the ability of DD or GVAR techniques to locally improve simulated MJO variability has yet to be documented. A number of local phenomena may impact the large-scale monsoon circulation simulation depending on how they are locally reproduced. Examples include; Tibetan plateau snow cover and its impact on large-scale thermal gradients and thereby the monsoon circulation (Shen et al. 1998, Becker et al. 2001) or the Saharan heat low and its impact on the West African monsoon (Lavaysse et al. 2009). Interactions between regional orography and monsoon circulations have been documented for South America (Lenters and Cook 1999), South Asia (Wu et al. 2007) and East Africa (Slingo et al. 2005). Over Asia,

regionally aerosol emissions can modify both surface and atmospheric solar heating, altering thermal gradients and the monsoon-scale circulation (Meywerk and Ramanathan 1999, Meehl et al. 2008). Similar effects were found by Konare et al. (2008), related to radiative cooling due to Saharan dust and the West African monsoon. Such processes are particularly important to represent when estimating potential changes in monsoon circulations in response to future GHG and aerosol emissions (Ramanathan et al. 2001).

4.2 Key Local to Regional Processes influencing monsoon variability

A number of local- to regional-scale processes strongly influence the accuracy and utility of simulated monsoon data. All these processes are highly regional, involving complex interactions across a range of spatial and temporal scales, but are often fundamental to the specific development of each monsoon. The primary driving force of most monsoons is a thermal gradient from ocean to land that drives a low level circulation with net moisture transport onto land. The combination of low-level mass and moisture convergence, land-surface heating and interaction with topography leads to the development of organized convection. Diabatic heating from convection feeds back onto the monsoon circulation. Improvement in the understanding and simulation of such phenomena is crucial for progress in predicting monsoon variability and change.

a. Surface heterogeneity

Land surface processes and land use change play an important role in regional monsoon variability. Koster et al. (2004) identified a number of “hot spots” of land–atmosphere coupling, where sub-seasonal precipitation variability is modulated by regional soil water characteristics. Strong coupling was identified over the Great Plains of North America, northern India and West Africa–Sahel. In these regions accurate estimates of soil water, either as an initial condition or during a model integration will likely impact simulated intra-seasonal monsoon variability. Douville et al (2001) showed this to be the case in an ensemble of GCM seasonal simulations, where soil moisture was modified either over the Sudan-Sahel region or India. Soil water anomalies had a significantly larger impact on rain rates in the African monsoon than over South Asia, likely due to a stronger oceanic moisture contribution to the South Asian monsoon than over Africa. Taylor et al. (2005) further showed that a more responsive and heterogeneous surface vegetation scheme impacted both the simulated diurnal cycle of convection, as well as the frequency and intensity of convective events over West Africa. A land-surface signal appears to contribute to onset characteristics of

the North American monsoon (Small 2001). Kelly and Mapes (2010) showed that biases in land surface fluxes reduce the accuracy of seasonal precipitation in the North American monsoon. Grimm et al. (2007) highlight feedbacks between soil moisture anomalies induced by early season precipitation and subsequent intra-seasonal precipitation variability over South America. Moreover SAMS precipitation seems to be more responsive to reductions of soil moisture than to increases (Collini et al., 2008, Saulo et al. 2010). Human induced land use change and associated regional scale responses have been shown to have cooling effect on climate in summer (Zampieri and Lionello 2011), although further analysis is required for a consolidated assessment of this effect.

b. The Diurnal Cycle

Tropical land regions exhibit a marked diurnal cycle, with, in general, a precipitation minimum in the local mid-morning, convection initiating in the late morning ~4-6 hours after local sunrise, with peak precipitation intensity reached in the late afternoon/early evening, followed by a gradual reduction through local midnight into the early morning minimum (Gray and Jacobson 1977, Sui et al. 1997, Garreaud and Wallace 1997). There is, nevertheless, a large degree of regional variability, particularly over land, where the mean diurnal cycle is primarily a result of numerous MCS developing differently in variable environmental conditions (Mathon et al. 2002, Mapes et al. 2003, Nesbitt and Zipser 2003).

An accurate representation of the diurnal cycle of convection remains an unresolved problem in climate models employing convection parameterizations, with convection systematically triggered too early in the day and precipitation maxima often phased with local noon, some 6 to 8 hours earlier than observed (Yang and Slingo 2001, Guichard et al. 2004). Figure 5 present the representation of the mean diurnal cycle of rainfall for July-August-September, averaged over of West Africa from 10 RCMs that downscaled ERA-interim using the CORDEX-Africa domain. TRMM is used as an observational reference, with a clear peak in precipitation from ~18.00 local time to 03.00 in the night and a minimum at local noon. ERA-interim 24-hour forecast precipitation is completely out of phase with TRMM, exhibiting a maximum at local noon and minimum from early evening to early morning. Most RCMs show the same out of phase shape for the diurnal cycle. Two models exhibit an evening/nocturnal precipitation maximum (UQAM-CRCM and SMHI-RCA). Both these employ variants of the Kain-Fritsch convection scheme (Kain and Fritsch 1990, Bechtold et al. 2001) with relatively advanced convective trigger functions and entrainment /detrainment schemes that are

responsive to large-scale conditions. (Kain and Fritsch 1990) Clearly, much work remains to fully simulate all components of the precipitation diurnal cycle over tropical land regions.

Excessive triggering of convection over land contributes to models precipitating too frequently and at too low intensities (Dai et al. 2006), while an incorrect phase to the diurnal cycle of convection and associated precipitation and clouds can induce systematic biases in the diurnal cycle of surface temperature and surface evaporation (Betts and Jakob 2002). Such errors may have a cumulative impact on soil moisture through the rainy season. Recent studies have increased our basic understanding of the diurnal cycle of convection (Grabowski et al. 2006, Khairoutdinov and Randall 2006, Hohenegger et al. 2008) and suggest a number of extensions to convection parameterizations that may improve the diurnal cycle. These include; advanced convective trigger functions, that account for heterogeneous surface and atmospheric forcing (Rio et al. 2009, Rogers and Fritsch 1996), convective entrainment that is sensitive both to the size of developing convective systems and the surrounding environment (Grabowski et al. 2006), the inclusion of evaporatively driven downdrafts and the impact of cold pools on vertical stability (Khairoutdinov and Randall 2006, Rio et al. 2009).

c. Low Level Jets

As discussed earlier, LLJs are an integral part of many monsoon systems. Statistically significant relationships have been found between nocturnally-peaking LLJs and nocturnal precipitation extremes in numerous disparate regions of the world (Monaghan et al. 2010). Widespread changes in the amplitude of near-surface diurnal heating cycles have been recorded as an important component of LLJ maintenance and that careful assessment of the impact of these changes on future LLJ activity is required.

Modeling studies indicate that land-atmosphere coupling is particularly strong over India, East Asia, West Africa, and the Great Plains of North America (Koster et al. 2004, 2006) and along the eastern slope of the subtropical Andes (Saulo et al. 2010). The complicated interactions involved in their formation and maintenance provides an excellent testbed for understanding interactions of a multitude of physical parameterizations. Improvement in the simulation of LLJs should lead to a better representation of the phase and amplitude of the diurnal cycle of precipitation and thus warm season rain. This is a severe test for models given the unique land-sea distributions, surface types, and orographic influences of the disparate monsoon regions (Sperber and Yasunari, 2006).

d. Regional ocean-atmosphere coupling

The primary source of water for monsoon rainfall is evaporation from the ocean. Processes influencing sea surface temperatures (SSTs) and ocean thermocline depth are therefore likely important for a good representation of monsoon precipitation. While coupled regional climate models are still in their infancy, a number of efforts have occurred, with indications that detailed representation of coastal ocean processes may lead to improvements in model simulations of monsoon intra-seasonal variability in some regions (Annamalai et al. 2005). Furthermore, it is well established (Schade and Emanuel 1999, Cione and Uhlhorn 2003, Pasquero and Emanuel 2008) that surface cooling from enhanced ocean mixing during the passage of a tropical cyclone reduces the likelihood of a subsequent cyclone in the same region over the coming weeks. In that sense, cyclone variability in the Bay of Bengal seems sensitive to a detailed representation of ocean mixed layer processes. On seasonal time scales, coupled ocean-atmosphere models are required to simulate the observed negative correlation between precipitation and SST over the warm waters of the AA monsoon region (Wang et al. 2005). Xie et al. (2007) present a regional coupled model of the tropical East Pacific and highlight the ability of the model to simulate tropical ocean instability waves, Central American gap winds and their impact on coastal SSTs. They further use the coupled system to investigate links between stratocumulus clouds, surface radiation, regional SST gradients and the seasonal evolution of the ITCZ and the Pacific equatorial cold tongue, indicating that such studies with high resolution coupled models can help understand and reduce systematic biases in coupled GCMs.

Better monitoring and understanding of air-sea interaction processes in subtropical anticyclones / subtropical and tropical gyres in the South Atlantic and South Indian Oceans will likely not only lead to improvements in the understanding and modeling of the ABFZ, SCTR and southern African climate but also in climate variability of the landmasses on the other side of the basin in each case; namely, South America and Australia.

Recently, Somot et al. (2010) compared a GVAR coupled to a high-resolution regional model of the Mediterranean Sea, to an equivalent uncoupled GVAR. Projected future climate changes differed significantly over Europe between the 2 configurations, due to a more detailed representation of the Mediterranean, stressing the importance of regional coupled processes in European climate. The inclusion of an interactive Mediterranean Sea in regional climate model simulations seems relevant because the Mediterranean is a major source of moisture for nearby region and water mass changes are a basic factor determining future sea level evolution (Gualdi et al 2011).

5. Concluding remarks

Considerable challenges remain before predictions of regional monsoon variability can be achieved at a level of accuracy required by society. These challenges relate both to our basic understanding of physical processes, as well as to their successful representation in numerical models. In this paper we highlighted a number of challenges that we consider crucial to improve our ability to simulate and thereby predict regional climate variability, particularly in monsoon regions. Central to many of these is the representation of moist convection and its interaction with regional dynamics and surface processes. For all aspects of monsoon simulations (seasonal to decadal prediction, to climate change) the representation of multi-scale convection and its interaction with coupled modes of tropical variability (where coupling refers both to ocean-atmosphere and/or land-atmosphere coupling) remains the leading problem to be addressed.

Other processes, however, can also play an important role in climate simulation at regional levels. The influence of land cover change requires better quantification. Likewise, aerosols loading resulting from biomass burning, urban activities or, land use changes due to agriculture are potentially important climate forcings requiring better understanding and representation in models.

Besides the progress already made, more work is required to elucidate mechanisms that give rise to intraseasonal variability. This timescale is key for users of climate forecasts and so there is a high societal need to exploit any potential predictability present using current dynamical and/or statistical models. It is expected that new observational and modeling campaigns, such as DYNAMO (Dynamics of Madden-Julian Oscillation) and YOTC (Year of Tropical Convection) will contribute to improving the understanding and numerical representation of active and break monsoon cycles. Alongside this, it is important to consider how the time-varying, large-scale environment interacts with variability in regional weather systems including MCS, easterly waves and tropical cyclones.

On longer timescales an improved understanding of monsoon variability on interannual to decadal is required to better understand, attribute and simulate near-term climate change and to assess the potential for inter-annual and longer monsoon prediction. As of 2011, the next phase of climate change model simulations (WCRP/WGCM/CMIP5) is ongoing in support of the IPCC 5th Assessment Report (Taylor et al. 2009). CMIP5 simulations will provide improved regional-scale information compared to earlier GCM intercomparison projects, through the use of higher resolution models. Careful analysis of these simulations will provide new indications of how climate change may affect monsoon systems particularly in the

coming decades. Community analysis of simulated monsoon processes in these runs are expected, with some activities having already started (e.g. by CLIVAR-AAMP). CORDEX will also downscale CMIP5 runs over monsoon land regions, allowing the benefits of increased model resolution in simulating e.g. intraseasonal variability of the various monsoons to be assessed.

Intense work is currently dedicated in many WCRP programs and projects to improve models, data-assimilation and data-gathering components of numerical climate prediction systems in order to increase forecast skill. However, further advances are needed: accelerating the improvement of overall model performance, and strengthening the links between model evaluation at the level of the application and the process-oriented refinement of the model formulation. These, in turn, require closer collaboration among the data, model user, and model development communities on the one hand and the academic and “operational” model development communities on the other (Jakob, 2010). An important area of common research is the design of metrics to quantify the ability of models to simulate key features of regional climate systems. It should also be noted that models largely developed in the Northern Hemisphere may not perform optimally over parts of South America or sub-Saharan Africa and attention needs to be given to regionally sensitive parameterizations.

The community must also exploit the rapidly evolving computing opportunities afforded by advances in computer hardware and software engineering. Priority must be given to developing multi-model, multi-member prediction systems, running models with sufficient resolution to resolve key topographic features and mesoscale factors that mold regional climate. Complementing this effort is the need to expand climate models into earth system models that more thoroughly represent the climate system (Shukla et al. 2009).

A challenge will still remain to connect predictions of regional climate variability and projections of change to practical outcomes. More research and investment is needed to translate climate data into actionable information at the regional and local scales required for decisions (Vera et al. 2009). The expansion of activities must include: i) Better determination and availability of agreed and reliable sets of data/variables required to address specific socio-economic sector vulnerability; ii) ways of securing climate observing systems, particularly in less developed regions; iii) ways to assemble, quality-check, reprocess and reanalyze datasets relevant to climate prediction at regional and local scales; iv) characterization of the uncertainties associated with climate predictions including properly accounting for those aspects that are and are not predictable; v) tailoring climate information to local scales and sector needs, and vi) supporting long-term training of climate scientists in developing nations, coupled with an effort to ensure suitable infrastructures by which scientists in these regions

can access, analyze, and ultimately develop prediction data and subsequently distribute this data to users in the region.

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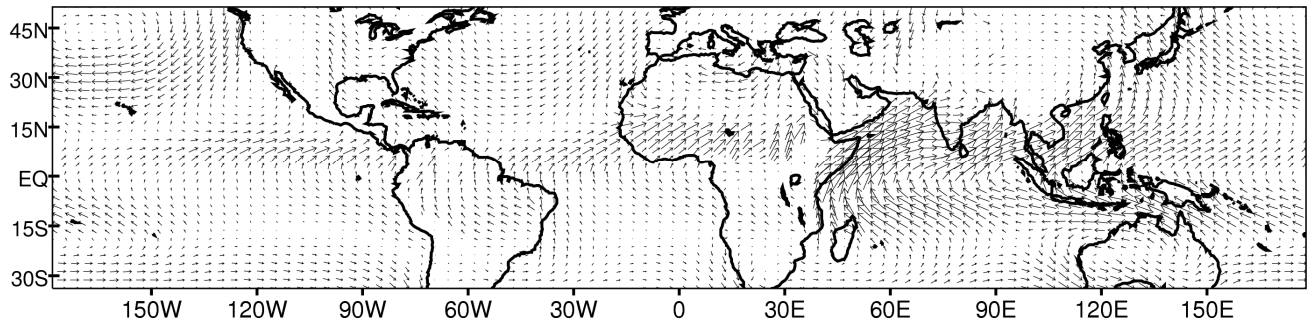


Figure 1: Seasonal change in observed lower tropospheric wind (925hPa) over the tropical monsoon regions (JJA minus DJF). Note the obvious reversal from north-easterly to south-westerly winds near West Africa and India and from anticlockwise to clockwise circulation in tropical South America, from northern hemisphere winter to summer. (Courtesy of A. Turner).

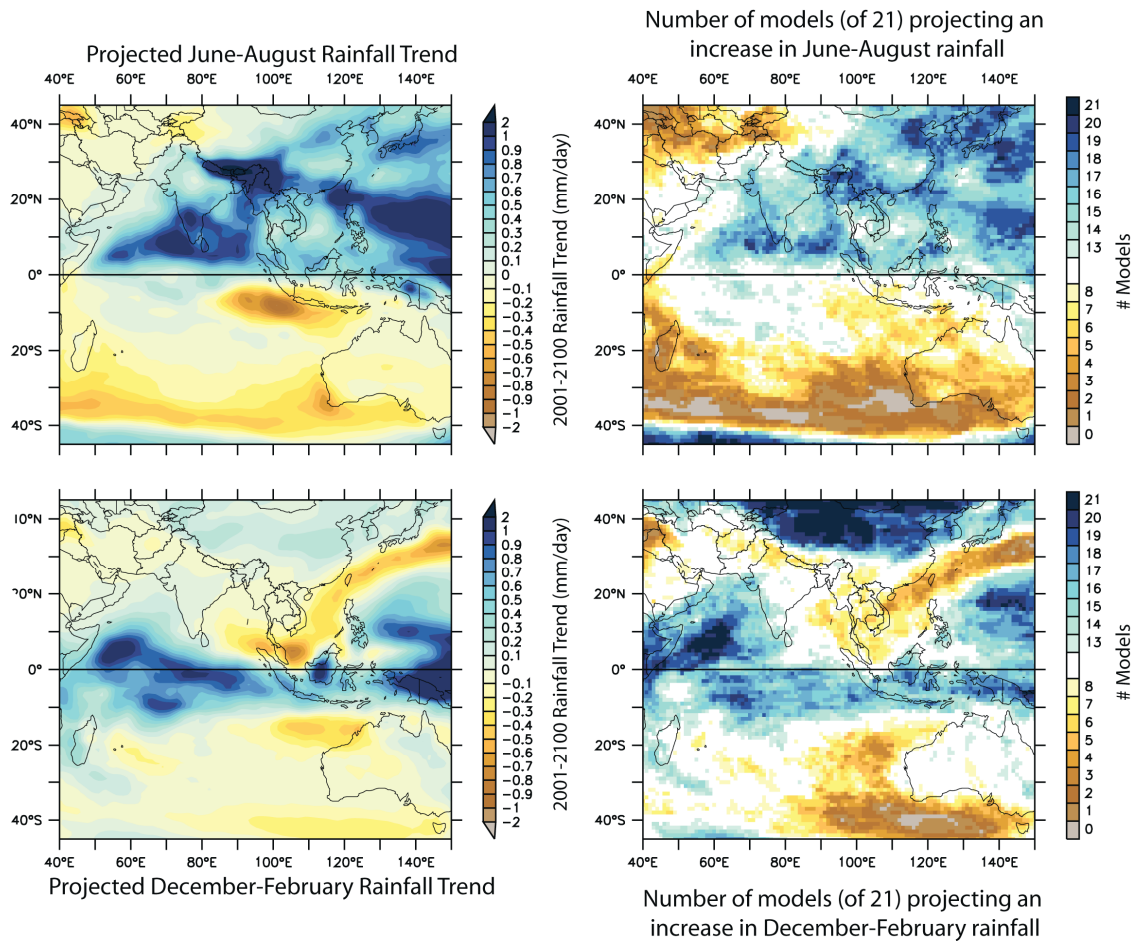


Figure 2: Projected change in precipitation amount over the Asian-Australian monsoon region in June-August (top row) and December-February (bottom row) due to anthropogenic climate change using the CMIP-3 models. The left panels show the 2001-2100 trend in mm/day (21-model average), and the right panels show the number of models (of 21) that have an increasing trend. (Adapted from Christensen et al., 2007).

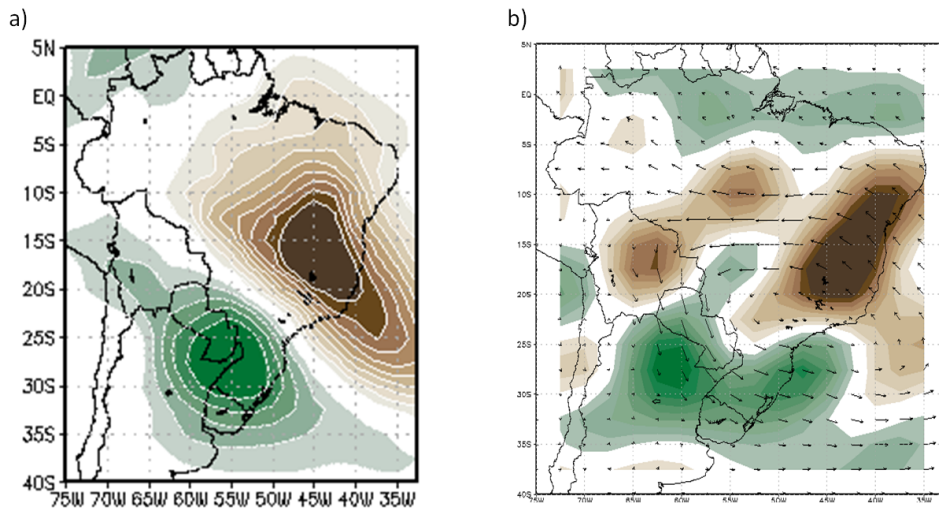


Figure 3: a) EOF1 pattern for 10-90 day filtered OLR anomalies during austral summer. b) Regression map between EOF1 principal component and 850-hPa wind anomalies (vectors) and the associated divergence (shading) (Courtesy of Paula Gonzalez, IRI)

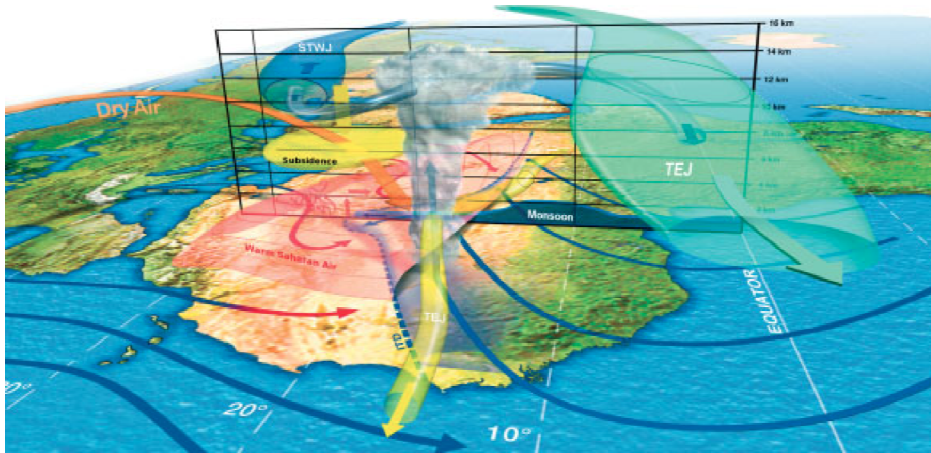


Figure 4: (Three-dimensional schematic view of the West African monsoon (see text for details). From Lafore et al, 2011.

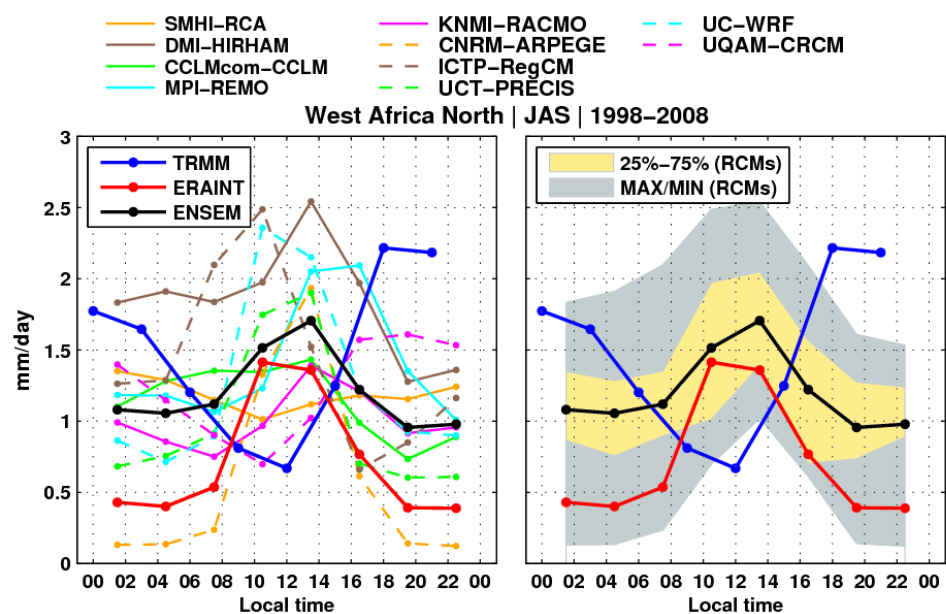


Figure 5: Mean diurnal cycle of precipitation averaged over West Africa and for the period 1998-2008. Left panel shows TRMM342B, ERA-interim, the 10 RCM ensemble mean and results from each RCM. Right panel plots in yellow shading the spread of the 50% most accurate RCMs and the full spread of the RCM results.