Seasonal Drought in the Greater Horn of Africa and Its Recent Increase during the March–May Long Rains

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ABSTRACT

This paper provides a review of atmospheric circulation and sea surface temperature (SST) conditions that are associated with meteorological drought on the seasonal time scale in the Greater Horn of Africa (the region 10°S–15°N, 30°–52°E). New findings regarding a post-1998 increase in drought frequency during the March–May (MAM) “long rains” are also reported. The period 1950–2010 is emphasized, although rainfall and SST data from 1901–2010 are used to place the recent long rains decline in a multidecadal context. For the latter case, climate model simulations and isolated basin SST experiments are also utilized.

Climatologically, rainfall exhibits a unimodal June–August (JJA) maximum in west-central Ethiopia with a generally bimodal [MAM and October–December (OND) maxima] distribution in locations to the south and east. Emphasis will be on these three seasons. SST anomalies in the tropical Pacific and Indian Oceans show the strongest association with drought during OND in locations having a bimodal annual cycle, with weaker associations during MAM. The influence of the El Niño–Southern Oscillation (ENSO) phenomenon critically depends on its ability to affect SSTs outside the Pacific. Salient features of the anomalous atmospheric circulation during drought events in different locations and seasons are discussed. The post-1998 decline in the long rains is found to be driven strongly (although not necessarily exclusively) by natural multidecadal variability in the tropical Pacific rather than anthropogenic climate change. This conclusion is supported by observational analyses and climate model experiments, which are presented.

1. Introduction

Drought is an inescapable feature of the Greater Horn of Africa’s generally semiarid climate and with roughly 75% of the labor force in the region involved in smallholder, rain-fed agriculture (Salami et al. 2010), its impacts on crops and human populations can be acute. The severe drought of 2010/11 in Kenya, Somalia, and southeastern Ethiopia, for example, affected some ten million people and was a contributing factor to more than 250,000 fatalities in Somalia alone (UN OCHA 2011; Checchi and Robinson 2013). The combination of drought and civil conflict in 1983/84 triggered a famine that led to an estimated one million fatalities in northern Ethiopia and Sudan (Africa Watch 1991; Devereux 2000), with the scale of the tragedy leading the United States Agency for International Development (USAID) to establish the Famine Early Warning System (FEWS) in 1985 in an attempt to avert such calamities in the future.

Given the inevitability of drought and its deleterious consequences it is not surprising that across the Greater Horn, where rain-fed agriculture is the mainstay, where food security is often threatened (Funk et al. 2008; Funk and Brown 2009), and where the largest contribution to electricity generation is hydropower (Kaunda et al. 2012), that drought information is especially valued. Nor is the interest in drought information limited to assessments of current conditions or the provision of seasonal forecasts. Seemingly contradictory signals are emerging regarding drought on longer time scales, particularly in eastern Ethiopia, Kenya, Somalia, and north-central Tanzania. Since 1999 there has been a substantial decline in the March–May (MAM) “long rains,” inevitably leading to speculation over the possible role of anthropogenic climate change (Lyon and DeWitt 2012). Some studies (e.g., Funk et al. 2008) have suggested that the long rains decline has been occurring over a longer period, and link it to an upward trend of SSTs in the central Indian Ocean that has been observed over the past
several decades. Meanwhile, there is a strong consensus among climate model projections that the long rains will increase during the current century in response to anthropogenic greenhouse gas forcing (Christensen et al. 2007; Shongwe et al. 2011). Is the post-1998 increase in drought frequency during MAM associated with long-term climate change or something else? What is the role of decadal variability in modulating drought occurrence during the long rains? How does seasonality fit into the drought picture more generally, including interannual variations in its occurrence?

This paper reviews some of the main characteristics of meteorological drought in the Greater Horn, including the seasonality of its occurrence [the MAM, June–August (JJA), and October–December (OND) seasons are examined] and its relationship to large-scale patterns of sea surface temperature (SST) variability in the global oceans. The study domain considered here extends from 30° to 52°E and 10°S to 15°N (box in Fig. 1). Emphasis is first placed on seasonal droughts associated with interannual variability of rainfall, with decadal-scale variability subsequently considered, focused on the MAM season only. Before considering drought across the region some general background on the mean climate state and the variability about the mean is presented.

While the paper emphasizes linkages between seasonal drought in the Greater Horn and anomalous regional and large-scale SST and atmospheric circulation patterns, there certainly remain considerable gaps in our understanding of the region’s climate. As reviewed by Nicholson (1996), there appears to be no single cause for its semiarid mean state, for example, and it is challenging to identify coherent synoptic circulation systems associated with the generation of daily rainfall. While some process studies have been undertaken (e.g., Vizy and Cook 2003), the underlying physical mechanisms associated with regional and local rainfall and its variability have yet to be fully elucidated. Indeed, gaining a better physical understanding of what regulates the annual cycle of rainfall itself across the Greater Horn requires additional work. For example, why is the region semiarid despite its location in the deep tropics? Why do the long rains in MAM generate more rainfall (climatologically) than the “short rains” of OND? New insights provided by this paper relate to the decadal modulation of drought during the MAM long rains season, linking it to decadal-scale SST variations in the tropical Pacific. Understanding the associated physical mechanisms responsible for the identified behavior on these time scales, however, remains an area of active research.

The paper is outlined as follows. Section 2 provides details on the data used in the study while section 3 provides a general overview of the region’s climate. Section 4 focuses on seasonal-to-interannual variations in drought and its relationship to SST variations in the global tropics. Section 5 explores the post-1998 increase in drought frequency during MAM by placing this behavior in a longer historical context on decadal time scales. Section 6 provides a brief summary and the main conclusions of the study.

2. Data and methodology

a. Observational data

Digital elevation data from the 5-minute spatial resolution gridded elevations/bathymetry for the world (ETOPO5) dataset were obtained from the data library at the International Research Institute (IRI) at http://iridl.ldeo.columbia.edu/SOURCES/NOAA/NGDC/ETOPO5/ele/.. Four precipitation datasets were used. The primary dataset was the gridded, gauge-based monthly analysis from the Global Precipitation Climatology Centre version 6 (GPCC; Rudolf and Rubel 2005) at 0.5° latitude × 0.5° longitude spatial resolution and covering the period 1901–2010, with some results compared with the similar,
time series version 3.2 (TS3.2) gridded product (0.5° latitude × 0.5° longitude resolution) from the University of East Anglia (UEA; Harris et al. 2014). Merged gauge-satellite rainfall estimates from the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin 1997) and the Global Precipitation Climatology Project (GPCP, version 2.2; Huffman et al. 2009) were also utilized. The spatial resolution of both the CMAP and GPCP data is 2.5° latitude × 2.5° longitude, with data available for the periods 1979–2011 and 1979–2009, respectively. The precipitation data were used to construct area-average rainfall anomaly time series for the MAM, JJA, and OND seasons. When constructing the time series for a given season, grid points in the study domain (box in Fig. 1) where the associated climatological 3-month rainfall (e.g., OND) was less than 25% of the climatological total annual rainfall (indicating a climatological dry season) were first masked before computing an area-average anomaly. Thus, throughout the paper the terms long rains, short rains, and JJA rainfall will correspond to specific locations within the study domain (i.e., the
areas shaded blue/green in Figs. 2b–d). Using standardized anomalies to compute the area-average rainfall indices had very little influence on the overall results.

Monthly averaged values of various atmospheric variables contained in the National Centers for Environmental Prediction (NCEP) Reanalysis 1 product (R1; Kalnay et al. 1996) were employed. The gridded R1 data cover the period 1948–2013 and are at 2.5° latitude–longitude resolution. Vertically integrated (850–500 hPa) moisture flux anomalies were computed using R1, the anomalous flux being separated into three terms:

$$\langle qV \rangle_a = \langle q_m V_a \rangle + \langle q_a V_m \rangle + \langle q_a V_a \rangle,$$

where the subscripts $m$ and $a$ refer to monthly mean and anomalous values, respectively, $q$ is the specific humidity, $V$ is the horizontal wind vector, and the angle brackets represent a seasonal average. The first term on the right-hand side of (1) was dominant, indicating the important role of circulation anomalies, and is the quantity plotted in subsequent figures. Gridded, monthly SST data were obtained from the extended reconstructed SST version 3b dataset (ERSST; Smith and Reynolds 2003) and the Hadley Center Sea Ice and Sea Surface Temperature dataset (HadISST; Rayner et al. 2003). The ERSST data have a spatial resolution of 2.0° latitude $\times$ 2.0° longitude; HadISST is at 1.0° resolution. Unless otherwise noted, the climatological base period used for computing anomalies in all fields was 1971–2000. A $t$ test was used to evaluate the statistical significance of temporal correlations and composite anomaly fields.

b. Model data

The use of climate model experiments was limited to the examination of the post-1998 decline in the long rains during MAM. Two sets of climate model experiments were evaluated: “GOGA” simulations, where the model was forced with observed global SSTs and runs forced only with observed SSTs in the tropical Pacific (“POGA” runs). The climate model utilized is the ECHAM4.5 ( Roeckner et al. 1996) atmospheric model having a horizontal resolution of roughly 2.8° latitude/longitude (a T42 spectral truncation was used). GOGA runs consisted of an ensemble of 24 members, using slightly different initial atmospheric conditions and forcing only from the observed global SSTs. Previous work (Lyon et al. 2014) found that this model performs well in simulating the seasonal climate of the Greater Horn and it is one of models included in the multimodel ensemble used by the IRI in the generation of its seasonal climate forecasts for the globe. The POGA runs use ECHAM4.5 forced with observed SSTs only in the tropical Pacific (25°S–25°N, 120°E–80°W). Using slightly different initial atmospheric conditions, an ensemble of 30 members was generated for the POGA runs, with a slab ocean model run outside the tropical Pacific domain so that SSTs could respond thermodynamically to the overlying atmosphere. Although the slab model contains no ocean dynamics it does include a climatological surface heat flux term ($q$ flux) to keep the SST annual cycle close to that found in observations (Lyon et al. 2014).

3. Climatological background

The climate of the Greater Horn shows considerable spatial heterogeneity owing in large part to substantial topographic variations across the region (e.g., Hession and Moore 2011). Despite its location within the deep tropics much of the region is semiarid, with the driest conditions located over northern Somalia (Fig. 2a). There are likely several factors contributing to the general aridity of the region (Nicholson 1996), one being a generally divergent large-scale flow, particularly over the low-lying, eastern regions. Average annual rainfall nonetheless varies by an order of magnitude between the northeastern coastal regions of Somalia and the western highlands of Ethiopia, where orographic rainfall is substantial. Even in highland areas spatial variations in rainfall can be substantial over distances of tens of kilometers given orographic variations and the seasonally varying direction of the low-level wind flow (Nicholson 1996). Orography also controls important features of the atmospheric circulation, an example being the gap between the Ethiopian and Kenyan highlands leading to the formation of the Turkana low-level jet, which channels southeasterly flow from over the Indian Ocean inland (Kinuthia and Asnani 1982; Kinuthia 1992; Viste and Sorteberg 2013).

Given such complexities, it is not surprising that the annual cycle of precipitation also shows considerable spatial heterogeneity (Figs. 2b–d). Across much of the region the annual cycle is generally bimodal, with greatest amounts occurring during the long rains in MAM and the short rains, typically during OND. In west-central Ethiopia rainfall shows more of a unimodal, boreal summer maximum (JJA considered here) as low-level winds carry moisture into the region from the continental interior, influenced by the migration of the intertropical convergence zone (ITCZ) across the eastern Sahel into the western parts of the study domain. In Tanzania, the rainfall annual cycle becomes more unimodal in southern areas of the country, with an austral summer maximum in southernmost areas (Ogallo 1988). This is largely associated with the southern
migration of the ITCZ across the western Indian Ocean into that region.

While precipitation during the MAM and OND seasons depends fundamentally on the meridional translation of the ITCZ across the equator, topography also plays a role in affecting the annual cycle of precipitation. The long rains, for example, typically undergo a substantial meridional “jump” of several degrees latitude during MAM as low-level southerly flow develops off the equatorial east coast (Riddle and Cook 2008), concurrent with the development of the Findlater jet. The jet is itself tied to the north–south orientation of orography (Findlater 1969; Hart 1977; Vizy and Cook 2003).

Interannual variability of the short rains is typically greater than that for the long rains as seen in Fig. 3a, based on the GPCC data (the geographical regions considered are roughly the same; see shading in Figs. 2b and 2d). The standard deviation of rainfall for the two seasons differs by roughly a factor of 2 (19.7 mm for OND and only 10.4 mm for MAM in GPCC). In addition, almost 50% of the variation in annual rainfall is accounted for by the variability of the short rains alone (only 23% is accounted for by the long rains), consistent with previous studies (Beltrando 1990; Camberlin and Wairoto 1997; Mutai and Ward 2000; Nicholson 2000). This has important implications for interpreting rainfall variations based on annual average data (or paleoclimate data with coarse temporal resolution) as the variability of the short rains dominates the annual signal. For locations like west-central Ethiopia having a JJA rainfall maximum (see Fig. 2c) Fig. 3b indicates large interannual variations but also characteristics that are somewhat reminiscent of the Sahel, with comparatively wet conditions during the 1950s and 1960s and then a downward trend with increased drought in the 1980s and 1990s and a recent recovery (although 2009 was quite dry). While the mechanisms responsible for generating rainfall across the Sahel vis-à-vis the western regions of the Greater Horn domain can vary substantially (e.g., Nicholson 1996), the temporal correlation between the JJA rainfall index in Fig. 3b and a JJA Sahel rainfall index for a nonoverlapping area (8°–18°N, 20°W–25°E) is $r = 0.73$, which is statistically significant at $p < 0.01$.

4. Drivers of drought on the seasonal time scale

a. Relationships to ENSO and other SST anomalies

Despite considerable spatial heterogeneity of the climate across the Greater Horn, variations from the mean state typically exhibit much greater spatial coherence on the seasonal time scale (e.g., Nicholson 1996; Mutai et al. 1998; Bergonzini et al. 2004). The short rains of OND show the strongest teleconnection to large-scale climate modes such as ENSO (Beltrando 1990; Beltrando and Camberlin 1993; Goddard and Graham 1999; Indeje et al. 2000; Schreck and Semazzi 2004). While drought in tropical land areas is generally more closely linked with El Niño events (Lyon 2004; Lyon and Barnston 2005) as tropics-wide tropospheric warming leads to a more stable atmosphere (e.g., Chiang and Sobel 2002), drought during the short rains season frequently develops in association with La Niña. The primary reason is that ENSO’s influence on the short rains is determined in large part by its influence on tropical Indian Ocean SSTs, with the western portion of the basin typically warming (cooling) as a lagged response to anomalous low-level winds and associated heat fluxes during El Niño (La Niña) events (Goddard and Graham 1999). The resulting zonal SST gradient anomalies favor decreased low-level moist static energy and rainfall over the western Indian Ocean and adjacent equatorial land areas during La Niña, with generally opposite conditions occurring during El Niño episodes. Of course anomalous zonal SST gradients are observed in the absence of ENSO and an Indian Ocean dipole (IOD) pattern has been reported (Saji et al. 1999) as a distinct and important mode of oceanic variability associated with short rains variability (e.g., Black et al. 2003; Behera et al. 2005; Molg et al. 2006). SSTs in other ocean basins appear to play
a secondary role in influencing the short rains, although some studies suggest the possible role of tropical Atlantic SSTs (Nicholson 1996; Camberlin et al. 2001; Anyah and Semazzi 2006), for example. It is noted, however, that even ENSO typically accounts for less than 50% of the variance in the short rains at a given location.

Unlike the short rains, the long rains of MAM show only a modest correlation with ENSO or any other large or regional-scale climate anomalies on the seasonal time scale (Ogallo et al. 1988; Hastenrath et al. 1993; Okoola 1999; Camberlin and Okoola 2003; Pohl and Camberlin 2011). That rainfall in OND is more strongly influenced by ENSO than in MAM suggests that successive rainfall seasons are largely uncorrelated. Indeed, the time series of the area-averaged short and long rains plotted in Fig. 3a correlate (over 1950–2010) at $r = -0.05$. Thus, the development of drought during the short rains season yields no statistical predictive information regarding its possible continuation during the subsequent long rains (although some climate model simulations suggest there may be increasing coherence between these seasons in recent years; Hoell and Funk 2014). In terms of the MAM season itself, there is some evidence of weak teleconnections on subseasonal time scales when the long rains are divided into the subseasons of March–April and, separately, May (Ininda 1999; Mutai and Ward 2000; Zorita and Tilya 2002; Camberlin and Okoola 2003; Moron et al. 2013). For example, Mutai and Ward (2000) suggest that ENSO shows some influence on the long rains in May, but not consistently in the prior two months, while Moron et al. (2013) find greater spatial coherence to rainfall in March than in April or May.

For locations such as west-central Ethiopia that exhibit a boreal summer rainfall maximum, seasonal drought occurrence shows some association to El Niño (Seleshi and Zanke 2004; Korecha and Barnston 2007; Segele et al. 2009), similar to the ENSO rainfall signal across the Sahel (Janicot et al. 1996; Rowell 2001; Camberlin et al. 2010, and many others). While a cross-equatorial dipole pattern of SST anomalies in the tropical Atlantic is also associated with rainfall variability across the Sahel on both interannual and longer time scales (Folland et al. 1986; Barnston et al. 1996; Ward 1998; Janicot et al. 2001; Hoerling et al. 2006), it generally does not have as much influence on JJA rainfall in west-central Ethiopia (Korecha and Barnston 2007). The relationship with Sahel rainfall has also been shown to be nonstationary in time [e.g., see Losada et al. (2012) and references therein]. A recent study by Williams et al. (2012) suggests that SSTs in the southern tropical Indian Ocean may also influence trends in JJA rainfall in portions of the Greater Horn. Overall, when considering all seasons and locations, SST variability associated with ENSO is the largest source of forcing for seasonal drought in the Greater Horn. However, even during the short rains of OND when its influence is strongest, statistically ENSO accounts for less than half of rainfall variance. El Niño favors drought during JJA in west-central Ethiopia whereas it is La Niña that is associated with drought in eastern Ethiopia, Kenya, Somalia, and northern Tanzania during the short rains. Neither phase of ENSO is a determinant factor in the development of drought during the MAM season. While Indian Ocean SSTs (often linked to the behavior of ENSO) are important during OND, they show little association with MAM or JJA rainfall.

To illustrate the SST relationships just discussed, Fig. 4 shows the temporal correlation (1950–2010) between the area-average rainfall time series for MAM, JJA, and OND and the contemporaneous, detrended seasonal SST anomalies in the ERSST dataset. Detrending was accomplished by subtracting a linear trend from each grid point (SST) or time series (rainfall). Based on a two-tailed $t$ test and assuming only 30 degrees of freedom (given that some autocorrelation exists in SSTs from one year to the next), correlations with absolute magnitudes exceeding roughly 0.35 are statistically significant at the 95% confidence interval. As anticipated, for the MAM long rains season (Fig. 4a) the correlations in all ocean basin locations are weak, with very few grid points anywhere reaching the level of statistical significance. For JJA rainfall (Fig. 4b) statistically significant correlations are found mainly in the tropical Pacific, with the negative correlations in east-central regions indicating that El Niño events are associated with drought conditions. The magnitude of the correlations there ($\approx -0.5$), however, indicates that El Niño accounts statistically for only about one-quarter of JJA rainfall variance. As indicated earlier, the JJA rainfall index (predominantly focused on west-central Ethiopia) is significantly correlated with a (spatially nonoverlapping) Sahel rainfall index. However, the former index shows no statistically significant relationship to tropical Atlantic SSTs in Fig. 4b. During OND, the short rains index shows statistically significant correlations with SSTs associated with ENSO in the tropical Pacific and positive correlations in the western tropical Indian Ocean. The simultaneous negative correlations in the eastern Indian Ocean in OND indicate the Indian Ocean dipole pattern (Saji et al. 1999). Thus, during OND drought in eastern Ethiopia, Kenya, and Somalia (i.e., locations experiencing the short rains) is associated with La Niña conditions in the Pacific and a negative IOD in the near-equatorial Indian Ocean, featuring below-average temperatures in the western part of the basin off the east
coast of Africa. However, the response of the Indian Ocean to ENSO forcing is critical during this season (e.g., Goddard and Graham 1999; Behera et al. 2005) if La Niña conditions are to translate into the development of drought. Maximum correlations with western Indian Ocean SSTs of $\approx 0.7$ indicate that less than half the variance in the short rains index is statistically accounted for by linear regression.

Since drought may certainly develop in the absence of ENSO, the 10 driest MAM, JJA, and OND rainfall seasons (1950–2010) were identified in the area-average rainfall time series based on detrended GPCC data (see Table 1). The composite standardized SST anomalies (also based on detrended data) for the three seasons are shown in Fig. 5 where only statistically significant values ($p < 0.10$) are plotted. The composite SST anomaly patterns in Fig. 5 are consistent with the correlation analysis of Fig. 4. For MAM (Fig. 5a), tropical ocean regions show virtually no coherent and statistically significant anomaly structures, with only a small region of

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below-average SSTs seen north of the equator in the central Pacific. During JJA, positive SST anomalies in the east-central tropical Pacific with a surrounding horseshoe pattern of negative anomalies suggest the role of El Niño in generating drought in the northwestern part of the study domain. For OND, the spatial extent of statistically significant SST anomalies increases dramatically (ENSO events are typically more developed during this season), with a clear La Niña pattern in the tropical Pacific and below-average SSTs in the western Indian Ocean (again, often forming in response to La Niña). For the OND short rains, 7 of the 10 driest OND seasons occurred during La Niña events, defined as a seasonal average SST anomaly of less than \(-0.5^\circ\text{C}\) in the Niño-3.4 region. For JJA rainfall, 6 of the 10 driest years were associated with El Niño events (Niño-3.4 SST anomalies exceeding 0.5°C).

b. The protracted drought of 1982–84

While the above relationships provide some insight into seasonal drought occurrence in the Greater Horn (and suggest some potential predictability given its relationship to ENSO and other SST patterns), it is protracted drought events that lead to the most devastating impacts (e.g., Little et al. 2001; Verdin et al. 2005). In addition, while ENSO and the IOD can play a role, they are certainly not the only factors involved. To give an example of the complexity of drought in this context a very brief overview of some of the climate characteristics of the severe drought of 1982–84 during the JJA season is presented here. This event was associated with the failure of multiple rainy seasons, which showed some similarities to, as well as important differences from, the composite view. Table 1 indicates that JJA rainfall during 1982 and 1984 was among the 10 driest seasons over the period 1950–2010. As mentioned earlier, this protracted drought helped trigger a famine that resulted in substantial loss of life, particularly in western Ethiopia and south-central Sudan.

Some climate features of the 1983/84 drought event are summarized in Fig. 6, which shows standardized SST anomalies and concurrent seasonal rainfall (GPCC) and 850–500-hPa vertically integrated moisture flux anomalies [the first term on the right-hand side of (1)] from RI.
for the JJA season of 1982, 1983, and 1984. During JJA of 1982 a major El Niño event was developing in the Pacific while SSTs were below average (above average) in the tropical South Atlantic (western Indian) Ocean (Fig. 6a). Rainfall was below average across much of northern Ethiopia westward into Sudan and Chad. The anomalous integrated moisture flux field for JJA 1982 (Fig. 6b) indicates an anomalous anticyclonic circulation over much of the Sahel, with an anomalous flux of dry air from the north into northern Ethiopia. In a statistical sense, the developing El Niño was consistent with seasonal drought occurring in this region during JJA of 1982.

The subsequent summer of JJA 1983 was during the tail end of the strong El Niño, with SSTs in the far eastern equatorial Pacific remaining well above average (Fig. 6c). The below- (above-) average SSTs in the southern tropical Atlantic (western Indian Ocean) weakened (strengthened) relative to the previous year. Rainfall across northwestern Ethiopia and much of Sudan and South Sudan remained below average (though not as dry as in 1982). The anomalous integrated moisture flux pattern in JJA 1983 (Fig. 6d) again indicates an anomalous anticyclonic circulation over much of the Sahel, with an anomalous northeasterly flux extending from south of the Arabian Peninsula into northern Ethiopia. Previous work on the JJA 1983 drought event (focusing on conditions in the Sahel) has suggested that the very warm conditions in the western Indian Ocean...
were also an important contributor to the drying (Bader and Latif 2011). However, as shown in Fig. 5, no robust relationship exists between western Indian Ocean SSTs and the area-average JJA rainfall index for the Greater Horn. Indeed, the model results of Bader and Latif (2011) show wetter than average conditions across drought-stricken areas of Ethiopia in JAS 1983 in response to SST forcing from the Indian Ocean. Overall, it is not clear the role Indian Ocean SST anomalies played in association with the drought in northwestern Ethiopia in JJA 1983.

JJA 1984 was the driest of the three years in northern Ethiopia and was accompanied by an impressive band of below-average rainfall that extended from Ethiopia westward across the Sahel (Fig. 6e). This rainfall anomaly feature was part of a pronounced dipole pattern, with enhanced rainfall extending from over the Gulf of Guinea (not shown) eastward across much of the Congo basin. The moisture flux anomaly pattern in JJA 1984 (Fig. 6f) indicates an eastward expansion of the anomalous anticyclonic circulation seen in previous seasons, extending from across the central Sahel to Yemen. An anomalous northeasterly flux is again seen across northern Ethiopia. The anomalous SST pattern of JJA 1984 was markedly different from 1982 and 1983, however. Much of the eastern tropical Pacific was now cooler than average, a condition not statistically linked to drought in northern Ethiopia during JJA. SST anomalies in the Atlantic, however, underwent a more dramatic shift, reversing sign and becoming much warmer (cooler) than average in the South (North) Atlantic. This pattern of Atlantic SST anomalies favors drought in the Sahel owing to a southward shift of the ITCZ from its climatological position (Palmer 1986; Folland et al. 1986; Lamb and Peppler 1992; Vizy and Cook 2001, among others) and likely played a key role in generating drought across that region and possibly into Ethiopia in 1984. It is less clear what, if any, role the very cool conditions in the western Indian Ocean may have played in 1984.

Overall, this brief overview of the 1982–84 drought serves to show that multiple factors, some likely tied to SSTs and others not, contributed to the severity and duration of this particular multyear event.

c. General atmospheric circulation features associated with Greater Horn drought

Recurrent anomalies in rainfall and atmospheric circulation features accompanying drought in the Greater Horn were identified by examining composites of the 10 driest seasons, computed separately for MAM, JJA, and OND and listed in Table 1. Composite anomalies of the 850–500-hPa vertically integrated vector moisture flux

![Figure 7](image-url)
the Greater Horn region are weak, with some suggestion of an anomalous southerly flux over the western equatorial Indian Ocean and an anomalous westerly flux located over the central Indian Ocean. There is an anomalous anticyclonic circulation in the flux field centered over the Persian Gulf area. On the subseasonal time scale, previous work (Camberlin and Philippon 2002; Camberlin et al. 2010; Camberlin 1997) has found an anomalous southerly low-level flow off the African coast (and westerly flow across the western portion of the north Indian Ocean) is associated with anomalously dry conditions in the Greater Horn. In addition, Camberlin et al. (2010) find that the Indian monsoon onset dates are significantly correlated with the cessation date of the long rains in the Greater Horn. An early onset of the Indian monsoon is thus associated with an early demise of the long rains. A study of the onset and cessation dates of the long rains finds a shorter rainy season is linked to reduced total seasonal rainfall (Camberlin and Okoola 2003).

JJA rainfall anomalies in the Greater Horn region (Fig. 7b) are confined to the northern and western areas of the study domain, primarily over western Ethiopia into South Sudan (locations with a JJA seasonal rainfall maximum). Rainfall is noticeably below average across much of India and in portions of the Maritime Continent. This overall pattern is generally expected during El Niño conditions. The anomalous moisture flux pattern for JJA exhibits an anomalous easterly flux over the central Indian Ocean north of the equator, extending from India toward the Greater Horn. There is a suggestion of an anomalous anticyclonic circulation over the Arabian Peninsula and India. These moisture flux and precipitation patterns are generally consistent with previous studies (Bhatt 1989; Camberlin 1997; Whetton and Rutherford 1994; Vizy and Cook 2003), which were found to extend to the subseasonal time scale as well (e.g., Camberlin 1997) during JJA. At upper levels (not shown) this includes a weakened tropical easterly jet over the Indian Ocean (Osman and Hastenrath 1969; Hulme and Tosdevin 1989; Camberlin 1997; Vizy and Cook 2003), consistent with a weaker upper-level anticyclone over South Asia during seasons with weaker JJA rainfall there.

For the OND short rains season the composite precipitation anomalies (Fig. 7c) indicate that much of the Greater Horn study domain has below-average rainfall. Similar to MAM, it is also anomalously dry over land areas to the north and east, while above-average rainfall is seen over portions of southern India and the Maritime Continent into the Philippines. This rainfall anomaly pattern is generally consistent with La Niña conditions during this season. In contrast to MAM, the anomalous moisture flux is considerably stronger over the Indian Ocean in OND, with an anomalous westerly flux extending eastward to the Maritime Continent. This zonally oriented feature over the equatorial Indian Ocean has been shown in previous studies (Goddard and Graham 1999; Hastenrath 2000, 2001; Bergonzini et al. 2004; Schreck and Semazzi 2004; Hastenrath et al. 2007) to be closely linked to rainfall variability in the Greater Horn during OND. Combined with the SST anomalies, the situation is physically consistent: a cooler (warmer) than average western (eastern) Indian Ocean is associated locally with higher (lower) sea level pressure with the associated pressure gradient resulting in the anomalous westerly flow near the equator (Hastenrath et al. 1993). In the Greater Horn there is a resulting divergence of the low-level moisture flux, favoring decreased rainfall (Goddard and Graham 1999) in long rains locations.

The relationship between short rains variability (1950–2010) and the low-level zonal wind component over the central Indian Ocean during OND is shown in Fig. 8a. The figure shows the anomalous zonal component of the 850-hPa wind (U850) in R1 averaged over the region 5°S–5°N, 60°–90°E [similar to Hastenrath (2000), but multiplied by −1] along with the OND short rains index. The two time series are highly correlated \( r = 0.83 \) and statistically significant at \( p < 0.01 \) over this 61-yr period. Notice that for the MAM long rains (Fig. 8b) the correlation is only \( r = 0.23 \), which is not statistically significant. Also note that the magnitude of
the zonal wind anomalies is typically much weaker in MAM compared to OND, with the standard deviation of the index during OND (1.98 m s$^{-1}$) being nearly twice that of MAM (0.98 m s$^{-1}$).

Since seasonal drought in the Greater Horn during the OND short rains is typically associated with La Niña events and simultaneously with the U850 index over the central Indian Ocean, the latter is likely to be at least partially related to ENSO. Indeed, the correlation between U850 and a Niño-3.4 SST anomaly index is $r = -0.69$ for OND. The temporal correlation between global SSTs (40°S–40°N) and the U850 index (here, not multiplied by $-1$) is shown in Fig. 9 based on detrended ERSST and R1 data, respectively. Again, absolute correlation values exceeding roughly $r = 0.35$ are statistically significant at the 95% level. The figure indicates that during OND, U850 is indeed significantly correlated with ENSO as seen by the correlation pattern in the Pacific (La Niña being associated with anomalous westerlies for the U850 index in the Indian Ocean and drought in the eastern Greater Horn). The U850 index is also significantly correlated with an east–west dipole pattern of SSTs over the Indian Ocean.

Changes in the equatorial Walker circulation have also been linked to drought in portions of the Greater Horn. For the seasons considered here, Walker circulation changes are most pronounced in OND given the typical life cycle of ENSO events and associated changes in Indian Ocean SSTs (Ogallo 1988; Hastenrath et al. 1993; Hastenrath 2000; Black et al. 2003; Pohl and Camberlin 2011). Goddard and Graham (1999) conducted isolated-basin experiments with an atmospheric general circulation model (AGCM) and concluded Indian Ocean SSTs dominate the contribution from the Pacific in affecting an anomalous Walker cell over the Indian Ocean and thereby short rains variability in the Greater Horn. During La Niña, for example, enhanced rainfall over the west tropical Pacific and Maritime Continent constitutes an anomalous ascending branch of the circulation, while associated cooling of the western Indian Ocean favors decreased rainfall (the descending branch). Lower- and upper-level equatorial zonal wind anomalies over the Indian Ocean are viewed as completing a closed circulation.

During MAM, the picture is more nuanced. Of course, drought during MAM in the Greater Horn it is attended by anomalous local descent, but other circulation features such as the zonal wind anomalies along the equatorial Indian Ocean (e.g., the U850 index in Fig. 8) show little in the way of consistent variation during such cases. Hastenrath (2000) argues that a closed Walker circulation is not observed across the Indian Ocean during the MAM season while other studies, such as that of Pohl and Camberlin (2011), infer its existence from a vertical shear index of the zonal wind between 850 and 150 hPa for the same region. For the period 1979–2007 Pohl and Camberlin (2011) find a modest but statistically significant relationship between their shear index and eastern Greater Horn rainfall. They also show that temporal variations in the shear index are correlated with SST variations in the tropical Pacific and Indian Oceans resembling ENSO (and rather similar to Fig. 9 here). Interestingly, after removing the association with ENSO, Pohl and Camberlin (2011) find statistically significant correlations between their shear index and SST anomalies in the central equatorial Pacific, a point returned to in section 5.

To visualize some aspects of the above discussion, Fig. 10 shows composite anomalies of the 200-hPa velocity potential and associated divergent wind in R1 for the 10 driest MAM and OND seasons. Given issues concerning the quality of the R1 data including possible spurious trends, particularly in the tropics (Santer et al. 1999; Kistler et al. 2001; Trenberth et al. 2001; Chan and Nigam 2009), anomalies for specific events were computed from base periods before the introduction of
satellite information (1950–79) and after its assimilation (1980–2009). With this caveat in mind, the composite for the 10 driest years during MAM (Fig. 10a) shows anomalous upper-level divergence over the eastern Indian Ocean and western Maritime Continent region with anomalous convergence and inferred descent over the Greater Horn to the west. An anomalous easterly component of the divergent wind over the equatorial Indian Ocean is suggestive of an upper-level branch of a zonal overturning circulation. The overall pattern is somewhat similar for OND (Fig. 10b), but with anomalies of greater magnitude and with the three centers of action in the velocity potential field all shifted eastward relative to their MAM locations. A more robust, anomalous Walker circulation across the equatorial Indian Ocean is again inferred in the divergent wind field and there is a clear ENSO signature to the overall pattern.

Finally, in the Greater Horn, a region that is highly reliant on rain-fed agriculture, dry spells within any given rainy season can be as detrimental to crops as seasonal drought (e.g., Barron et al. 2003). Here it is simply noted that on subseasonal time scales, rainfall variations associated with the Madden–Julian oscillation (MJO) have been reported for both the short and long rains (e.g., Mutai and Ward 2000; Pohl and Camberlin 2006). While the spatial patterns of the rainfall response to the MJO over the Greater Horn show some differences from the typical ENSO response, Pohl and Camberlin (2011) indicate that the effects of ENSO and the MJO are typically additive. For example, during OND La Niña is typically associated with reduced (enhanced) rainfall over the Greater Horn (the Maritime Continent). When the enhanced convection phase of the MJO reaches the maritime region during a La Niña event the remote drying response in the Greater Horn may thus be further enhanced.

5. Decadal variability of the MAM long rains

a. Background

As mentioned in the introduction, since roughly 1999 there has been an increased frequency of drought in long rains locations during MAM. While the largest rainfall declines during this period have occurred during the months of April and May, a similar decline in the MAM seasonal total is also observed (Lyon and DeWitt 2012). This rainfall decline has led some to speculate that an anthropogenic climate change signal has begun to emerge in the region. Yet, paradoxically, the consensus of climate model projections is that the region will become wetter during the current century (Christensen 2007; Shongwe et al. 2011).

Figure 11a shows time series of anomalies (1979–2010 base period) in MAM long rains area-average indices based on three datasets described in section 2a. The figure suggests an overall precipitation decline over roughly the past 30 years, but also that rainfall in the post-1998 period has largely not recovered to earlier values following an abrupt downward shift. In addition,
FIG. 11. (a) Time series of area-average MAM rainfall anomalies (1979–2010 base period) for the Greater Horn from three datasets (mm month$$^{-1}$$) for the period 1979–2010 computed in the same manner as described for GPCC in section 2a. (b) Linear trend multiplied by segment length over which it was computed (shading, mm) for the MAM long rains index from GPCC. Vertical axis is segment length, horizontal axis the end date of fitted trend. Top inset in (b) shows statistically significant values ($$p < 0.05$$) based on a two-tailed t test; bottom inset in (b) shows the absolute value of the maximum positive or linear trend (multiplied by segment length, mm) as a function of segment length.
while there appears to be an overall downward trend in the time series, decadal variability could very well be a major contributor to a perceived 30-yr decline. To show this, the GPCC MAM long rains time series from 1901 to 2010 was used. The time series was not detrended. Rather, linear trends were fit to segments of the time series increasing in length from 3 to 109 yr at 2-yr intervals following Liebmann et al. (2010). The shading in Fig. 11b shows the magnitude of the trend for each segment (plotted at the end time of each segment), which has been multiplied by the length of the segment to yield units of millimeters. The plot indicates that both increasing and decreasing rainfall trends have been observed on different time scales over the 110-yr period analyzed. The largest-amplitude trends are for the shortest segment lengths, indicating the major influence of interannual variability. Given the sensitivity of linear trends to the start and end points of the segments, the recent drought period starting around 1999 results in negative trends for all segment lengths greater than about 15 yr. However, based on a two-tailed t test, no statistically significant (p < 0.05) downward trends are identified prior to 1999 for segment lengths greater than roughly 20 yr (shaded insert of Fig. 11b). In addition, the maximum absolute values of positive and negative trends for different segment lengths are roughly the same (line insert of Fig. 11b), indicating no preference for drying across all time scales. The same analysis applied to the UEA data (not shown) provided very similar results but with much weaker drying trends for the post-1999 period than seen in GPCC.

To date the post-1998 decline in the Greater Horn long rains has received fairly little attention from researchers. Funk et al. (2005) reported a downward trend in the long rains has been occurring since the 1980s, with an “abrupt decline since 1996” in their smoothed MAM rainfall time series for Ethiopia. In the report they presented some preliminary analyses showing statistical associations between the decrease in long rains precipitation and other regional wind and precipitation fields. They speculated that the decline in the long rains was largely attributable to increasing SSTs in the southern equatorial Indian Ocean. Funk et al. (2008) later employed statistical techniques and an idealized climate model experiment to further argue that a post-1980 rainfall decline was dynamically linked to increasing SSTs in the south-central Indian Ocean. In this view, rising SSTs there result in enhanced rainfall, disrupting onshore moisture transports into the Greater Horn as a result of a classic Gill–Matsuno-type response to the diabatic heating associated with the enhanced convection. They reported that the Indian Ocean warming likely has an anthropogenic component such that, despite the consensus of climate projections for a wetter Greater Horn due to anthropogenic forcing, the rainfall response to that forcing may in fact be of the opposite sign. A subsequent paper by Williams and Funk (2011) argued that the increasing tropical Indian Ocean SSTs are indicative of a westward expansion of the tropical Pacific warm pool, with an associated westward-shifted Walker circulation resulting in increased subsidence and reduction of the long rains in the eastern Greater Horn.

An alternative view of the post-1998 MAM long rains decline has recently been put forward (Lyon and DeWitt 2012). That study suggested that the key aspect of a perceived downward trend in rainfall over the past 30 years was not a continuous decline, but rather an abrupt downward shift that occurred around 1999. They demonstrated that an abrupt shift in tropical Pacific SSTs occurred in 1998/99, characterized by warming in the western part of the basin and cooling further east. Rather than the Indian Ocean, Lyon and DeWitt (2012) demonstrated the fundamental role of the tropical Pacific Ocean in forcing the recent rainfall decline by undertaking isolated basin experiments with an AGCM. When the model was forced with observed SSTs in the tropical Pacific while holding them to climatological (seasonally varying) values elsewhere, they obtained robust drying in eastern Ethiopia, Kenya, Somalia, and northern Tanzania, a regional-scale rainfall anomaly pattern consistent with observations during the post-1998 period (and similar to conditions during the severe drought of MAM 2011). While the Pacific SST pattern identified by Lyon and DeWitt (2012) was suggestive of decadal variability in the basin, they did not examine that aspect in detail.

b. Multidecadal Pacific SST variability and the recent increase in MAM drought

There have only been a few studies conducted on decadal rainfall variability in the Greater Horn and its relationship to global SSTs during the modern instrumental period, particularly for the MAM season. Nicholson (2000) provides a general overview of rainfall variability for the continent but does not make connections to specific SST patterns in the global oceans. Jury (2010) found a modest association between areas of the Greater Horn having a bimodal seasonal rainfall distribution and SST variability associated with the Pacific decadal oscillation (PDO) with little influence from its counterpart, the Atlantic multidecadal oscillation (AMO). Beltrando (1990) found some suggestion of decadal variations in April rainfall in a limited region of the Greater Horn but did not attribute these variations to any particular pattern of SST variability. Clark et al.
(2003) found interdecadal variations in the association between Greater Horn rainfall and the Indian Ocean zonal mode, but only consider the OND short rains.

Motivated by earlier findings, Lyon et al. (2014) recently examined SST variations in the global oceans over roughly the last 110 years, emphasizing the MAM season. In the study the global average SST anomaly and (simultaneous) ENSO signal were first removed from the SST data using linear regression before applying an empirical orthogonal function (EOF) analysis on the residual field. Results for the Pacific domain (40°S–60°N,120°E–80°W) are shown here in Fig. 12, which displays both the loading pattern and associated principal component (PC) time series of the leading EOF (accounting for 18% of the total residual variance) for the ERSST dataset. The loading pattern (Fig. 12a) is reminiscent of the PDO, with the associated PC time series (Fig. 12b) showing substantial multidecadal variability and a recent shift around 1999 (seen most clearly in the 9-yr running average). The upward shift in the PC time series in 1999 indicates SSTs becoming cooler in the eastern Pacific and warmer in the western portion of the basin with the timing of earlier shifts consistent with previous studies of multidecadal variations of the PDO (e.g., Mantua et al. 1997; Deser et al. 2004). The seasonal (9-yr moving average) values of the PC time series in Fig. 12 correlate at \( r = -0.67 \) (\( r = -0.86 \)) with associated values of the PDO index of Mantua et al. (1997), the latter based on SSTs limited to the North Pacific (north of 20°N). Regressions of the PDO index onto global SST anomalies (e.g., Deser et al. 2010) reveals a spatial pattern across the Pacific similar to the loading pattern in Fig. 12a, including a reversal in sign between the east-central and western equatorial Pacific. For the post-1998 period, however, the PC time series does exhibit differences from the behavior of the PDO. The 9-yr moving average PC time series has remained positive while the PDO index has shown greater variability over this period. One possibility is that decadal-scale changes in the character of ENSO (e.g., McPhaden et al. 2011) and the associated SST anomalies project more strongly onto the EOF loadings in Fig. 12 than those associated with the PDO in the North Pacific. This requires further investigation. Meanwhile, previous studies have indicated the challenge of using a single index to capture multidecadal Pacific variability (Deser et al. 2004), even in the North Pacific (e.g., Bond et al. 2003). The index used here is thus one indicator of multidecadal Pacific variability that is related to, but not identical to, the PDO index.

To see the relationship between the identified multidecadal SST variability in the Pacific and near-global precipitation variations, Fig. 13 shows the difference between MAM seasonal precipitation averaged across three individual cool phases (1914–25, 1947–76, and 1999–2010) and the most recent warm phase (1977–98). The pattern of recent drying in the Greater Horn during the post-1998 period (Fig. 13a) is remarkably similar to that seen in the two earlier periods. In addition, the rainfall anomaly patterns are very similar on a near-global scale across the cool phases. Using a slightly different methodology, Yang et al. (2014) obtain similar results. The post-1998 long rains decrease is thus seen to be a regional feature of a near-global precipitation anomaly pattern associated with multidecadal SST variability in the Pacific (Lyon et al. 2014).

The relationship between Pacific multidecadal variability and the MAM long rains is further supported by the time series in Fig. 14, which show the leading PC from the residual SST EOF analysis (i.e., Fig. 12b) and the detrended long rains index from GPCC. The time series cover the period 1901–2010 and both have been smoothed using a 9-yr running average. There is a clear
and robust relationship between the two series over the full period. The temporal correlation between the series is $r = 0.73$, which is statistically significant at $p < 0.01$ based on a two-tailed $t$ test and assuming 12 degrees of freedom. Similar results (not shown) are obtained if the GPCC rainfall data are not detrended; for that case $r = 0.69$. These results are also in agreement with Yang et al. (2014). It is emphasized that the focus here is on decadal-scale variations in rainfall. There is still considerable interannual variability in the MAM long rains, which is not necessarily linked to SST variations in the Pacific or elsewhere. For example, 1984 was the third driest MAM season between 1979 and 2010, while 1981 was the wettest year in that same period (see Fig. 11a).

The key role of tropical Pacific SSTs in forcing the post-1998 MAM long rains decline is seen in Fig. 15, which compares the precipitation response of the ECHAM4.5 AGCM (described in section 2b) for the GOGA runs (Fig. 15a) with the POGA runs (Fig. 15b). The plots show the difference in the ensemble-mean rainfall averages for 1999–2011 minus 1977–98. Both the GOGA and POGA runs capture the drying over the Greater Horn. Notably, both sets of runs are also able to capture many of the observed features in the near-global MAM precipitation anomaly pattern associated with the post-1998 shift in Pacific SSTs (i.e., as shown in Fig. 13a). Lyon et al. (2014) show that both GOGA and POGA runs are also capable of capturing some general characteristics of the 1998/99 shift itself, as revealed in various atmospheric fields.

While the observed SSTs used to force the GOGA and POGA runs include any anthropogenic signal, Fig. 16 provides a simple estimate of the contribution of natural variations to the observed difference in MAM SSTs between the two 14-yr periods 1999–2012 and 1985–98. Figure 16a shows this difference in the full SST
field, while Fig. 16b shows the difference after first removing, by linear regression, the near-global (55°S–55°N) average SST anomaly (a proxy for the externally forced signal; Trenberth and Shea 2006) from each grid point on a seasonal basis. The SST anomaly patterns and anomaly magnitudes are very similar, indicating that the vast majority of the observed change in SSTs since 1998 is associated with decadal variations, which includes any changes in the behavior of ENSO. In addition, changes in the zonal gradient of equatorial Pacific SST, measured as the difference between the western (5°S–5°N, 130°–170°E) and eastern (5°S–5°N, 80°–120°W) Pacific, are essentially identical when the global average SST is first removed (a gradient increase of 0.38°C) and when it is retained (increase of 0.33°C). Very similar results (not shown) were obtained when using the HadISST dataset.

Overall, the observational analyses indicate that the post-1998 decline in the MAM Greater Horn long rains appears to be driven strongly by multidecadal SST variability in the Pacific. The rainfall anomaly patterns across the Greater Horn during the periods 1914–25 and 1947–76 (Fig. 13c) are remarkably similar to that of 1999–2010 (Fig. 13a), for example. The POGA runs are able to capture the MAM drying during the long rains for the post-1998 period and many of the observed features of the near-global rainfall anomaly pattern. This result points to the key role of the tropical Pacific in forcing the post-1998 increase in long rains drought.

As a final view of recent atmospheric changes during the MAM season, Fig. 17 shows the difference in CMAP rainfall, 200-hPa velocity potential, and associated divergent wind (from R1) between the averaging periods 1999–2011 and 1979–98. Drying during the long rains season is associated with a La Niña–like pattern, with enhanced rainfall in the western tropical Pacific and drying in the east-central portion of the basin. Over much of the Indian Ocean rainfall is below average, the exception being extreme northern locations where there is a zonally elongated band of increased rainfall. The velocity potential anomalies are in very good agreement with the precipitation field, indicating strong convergence (divergence) over the Greater Horn and the south-central Indian Ocean (west Pacific). In comparison with Fig. 10a (which shows composite anomalies of velocity potential for the 10 driest MAM seasons since
1950), Fig. 17 indicates that the centers of action over the Indian and Pacific Oceans have shifted eastward.

In terms of the Greater Horn climate paradox, whether anthropogenic climate change will bring a wetter or drier future, the evidence provided here and in recent studies (Lyon and DeWitt 2012; Lyon et al. 2014; Yang et al. 2014) all indicate that decadal-scale variability in Pacific SSTs is a strong contributor to the post-1998 rainfall decline, which can easily be misinterpreted as long-term climate change. The extent to which anthropogenic forcing may or may not have modified the recent decadal climate signal remains an open question. Funk (2012) argues that exceptional warming in the western Pacific–Indian Ocean warm pool has contributed to more frequent droughts in the Greater Horn while noting that a specific attribution of the warming has not yet been made. The modeling experiment of Lott et al. (2013) concludes there was an anthropogenic contribution to the forcing of the MAM 2011 season drought (but not OND 2010), but they were not able to assess the statistical significance of their results.

The recent study by Yang et al. (2014) provides a cautionary note regarding the use of phase 5 of the Coupled Model Intercomparison Project (CMIP5) climate models to directly assess the variability of Greater Horn rainfall on different time scales. They found that the coupled models do a poor job replicating the rainfall climatology of the region and in capturing observed relationships between decadal rainfall variability and global SSTs. This has clear reliability implications when using climate projections. Taken at face value, the CMIP5 models generally indicate wettening in the Greater Horn as a result of anthropogenic forcing by midcentury (Yang et al. 2014). Alternatively, Cook and Vizy (2012, 2013) use a regional model and conclude that the regional response to anthropogenic forcing will result in drying in the Greater Horn during MAM over the same time period, using a subset of phase 3 of CMIP (CMIP3) climate projections for their boundary conditions. A more general and fundamental question is how tropical Pacific SSTs will be affected by climate change. The La Niña–like pattern of tropical Pacific SSTs associated with recent decadal variability is similar to longer-term trends in SSTs. Several studies report a strengthening of the Walker circulation over the Pacific (Chen et al. 2002; Compo and Sardeshmukh 2010; Solomon and Newman 2012; L’Heureux et al. 2013), not a weakening. Whether these trends continue or not has important implications for rainfall in the Greater Horn as well as for many other locations around the globe.

6. Summary and conclusions

The climate of the Greater Horn shows great spatial heterogeneity owing in large part to its complex terrain. Local rainfall regimes within the region generally consist of unimodal (JJA maximum) and bimodal (MAM and OND maxima) annual cycles, with annual rainfall totals varying by as much as an order of magnitude between the driest stretches of the coastal plains of Somalia and highland locations in Ethiopia. Given the region’s diverse climate, it is not surprising that the development of seasonal drought has different drivers. ENSO is seen to play the largest role, although even during OND when its influence is largest it accounts for less than half of interannual rainfall variability. El Niño events are associated with drought during JJA in the northwestern parts of the study domain, while La Niña is related to droughts in areas having an OND seasonal rainfall maxima to the south and east. Drought development during OND, however, also depends critically on the behavior of SSTs in the tropical Indian
Ocean, which may develop in response to, or independent of, ENSO. On interannual time scales, the long rains of MAM shows the weakest association with global SSTs of the three seasons considered.

In terms of anomalous atmospheric circulation features associated with seasonal drought, composites of the 10-driest seasons from 1950–2010 provide some suggestion of an anomalous Walker circulation cell across the Indian Ocean during MAM whereas a much more robust overturning circulation is identified during OND. The largest contributor to the anomalous vertically integrated moisture flux is the anomalous horizontal wind operating on the climatological specific humidity field. Anomalous fluxes have their largest magnitude in OND, show the least spatial coherence in MAM, and in JJA suggest a linkage between rainfall in the eastern Greater Horn and India, consistent with other studies. Overall, however, the physical mechanisms associated with rainfall variations in the Greater Horn on seasonal and subseasonal time scales are not very well understood and require more systematic study.

A current concern in the Greater Horn is the post-1998 increase in drought frequency during the MAM long rains season, underscored by the severe impacts associated with the 2010/11 drought. While some studies have tied this increase in drought frequency to longer-term upward trends in Indian Ocean SSTs with an anthropogenic contribution (leading to expectations for continued drought), here evidence is provided that the post-1998 decline in MAM is strongly (though not necessarily exclusively) a manifestation of natural multi-decadal variability of SSTs in the tropical Pacific basin, rather than anthropogenic climate change. In addition, while CMIP5 model projections generally show that the Greater Horn will become wetter during this century, they are also found to do a poor job capturing the annual cycle of rainfall and simulating historical decadal variability of the Greater Horn long rains and their relationship to global SST patterns. This result provides an important cautionary note regarding the interpretation of CMIP5 model projections for the region. Overall, whether anthropogenic forcing will lead to a wetter or drier Greater Horn during the current century is seen to remain an open question.

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