AGU PUBLICATIONS

Reviews of Geophysics

REVIEW ARTICLE

10.1002/2016RG000544

Climate and climatic variability of rainfall over eastern Africa

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Key Points:

- The factors leading to drought in eastern Africa, especially during the short rains, are inadequately understood
- Factors in interannual variability of the short rains are nonstationary; the most direct is the zonal circulation in the equatorial Indian Ocean
- Better seasonal partitioning could improve our understanding of rainfall in eastern Africa, its global teleconnections, and seasonal forecasts

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Citation:
Nicholson, S. E. (2017), Climate and climatic variability of rainfall over eastern Africa, *Rev. Geophys.*, *55*, doi:10.1002/2016RG000544.

Received 31 OCT 2016
Accepted 25 APR 2017
Accepted article online 28 APR 2017

Abstract This review examines several aspects of the climate of eastern Africa. The climatic commonality throughout the region is the frequent occurrence of drought severe enough to incapacitate the population. Because of recent droughts and evidence of disastrous, long-term climatic change, the region has become a major focus of meteorological research. This review covers six relevant topics: climatic regionalization, seasonal cycle, intraseasonal variability, interannual variability, recent trends, and seasonal forecasting. What emerges is a markedly different view of the factors modulating rainfall, the dynamics associated with the seasons, and the character of teleconnections within the region and the interrelationships between the various rainy seasons. Some of the most important points are the following. (1) The paradigm of two rainy seasons resulting from the biannual equatorial passage of the Intertropical Convergence Zone is inadequate. (2) The "long rains" should not be treated as a single season, as character, causal factors, and teleconnections are markedly different in each month. (3) The long rains have been declining continuously in recent decades. (4) The Madden-Julian Oscillation has emerged as a factor in interannual and intraseasonal variability, but the relative strength of Pacific and Indian Ocean anomalies plays a major role in the downward trend. (5) Factors governing the short rains are nonstationary. (6) Droughts have become longer and more intense and tend to continue across rainy seasons, and their causes are not adequately understood. (7) Atmospheric variables provide more reliable seasonal forecasts than the factors traditionally considered in forecast models, such as sea surface temperatures and El Niño–Southern Oscillation.

1. Introduction

Eastern Africa includes primarily the countries Kenya, Uganda, Tanzania, Ethiopia, Eritrea, Djibouti, Somalia, South Sudan, Rwanda, and Burundi (Figure 1). The commonalities of these countries include the frequent occurrence of severe drought [*Nicholson*, 2016a] and populations heavily dependent on the rainfall that is meager in most of the region. The region is equally plagued by floods. Of the seven most flood-prone countries in Africa, five are in eastern Africa [*Li et al.*, 2016]. These extreme events have devastating effects on the population, especially when occurring in the same year [*Nicholson*, 2014a]. There is serious concern for the region's future because of the strong consensus that climate change will have a major effect on rainfall, although the projections are quite diverse [*Shongwe et al.*, 2011; *Cook and Vizy*, 2013; *Otieno and Anyah*, 2013; *Kent et al.*, 2015; *Rowell et al.*, 2015; *Tierney and Ummenhofer*, 2015].

This paper provides a review of our current state of knowledge of the region's rainfall regime and its temporal and spatial variability. Six topics are considered: regionalization of the rainfall regime, the rainy seasons, intraseasonal variability, interannual variability, recent trends and extreme events, and seasonal forecasting. The newest findings concerning the region's rainfall regime are summarized, and factors that need further research are identified.

To facilitate the discussion, terminology needs to be clarified. Some decades ago, the term East Africa formally referred to the three countries of Uganda, Tanzania, and Kenya. Because much of the research on the region's climate was conducted in these countries, especially Kenya, many authors retain the term "East Africa" and refer specifically to these countries. Other authors utilize the term Greater Horn of Africa or GHA, indicating a broader analysis sector. The distinction is important because most of East Africa has a bimodal rainfall regime, with peaks in both the boreal spring and autumn. For such areas, the term "equatorial rainfall region" is frequently utilized in this article. In much of the additional area comprising GHA rainfall peaks in the boreal summer, so that term utilized here is the "summer rainfall region". Figure 1 delineates these two regions.

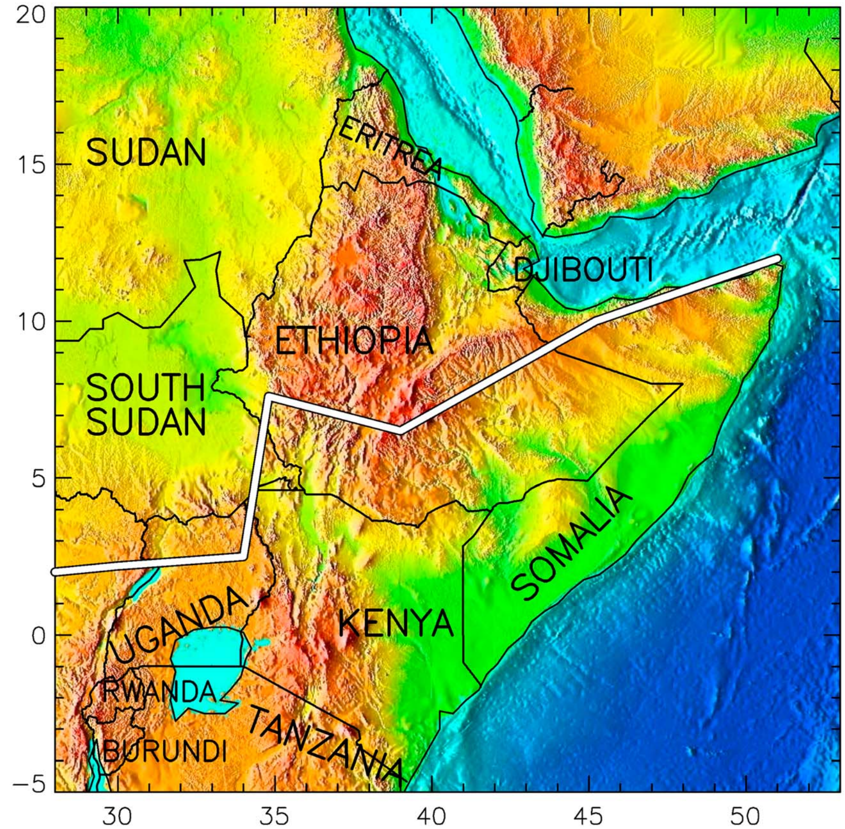

Figure 1. Location of equatorial (southern sector) and summer rainfall (northern sector) regions superimposed upon schematic of East African topography.

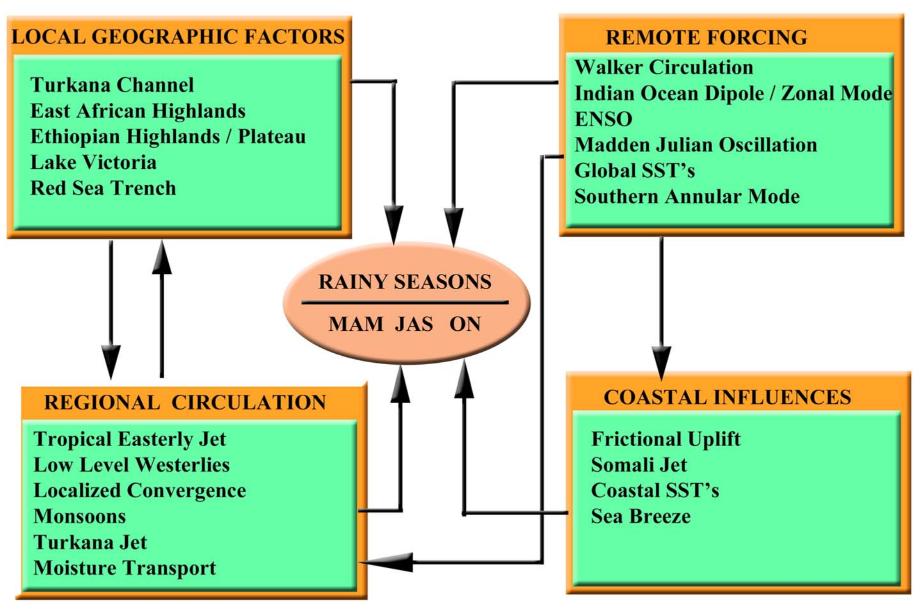

Figure 2. Schematic showing the factors influencing the three rainy seasons (MAM, JAS, and ON) of eastern Africa and the interrelationships among the factors.

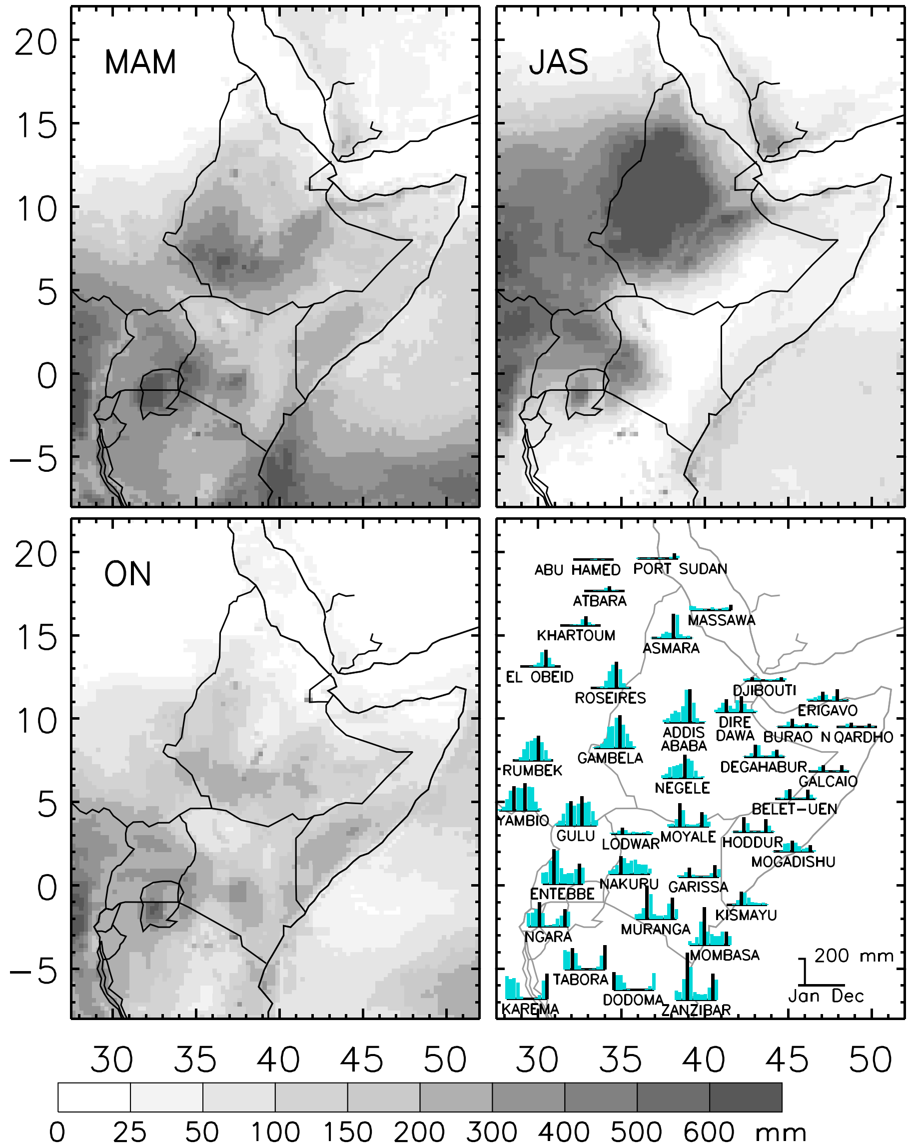

Figure 3. Mean rainfall in the three rainy seasons (March to May, July to September, and October–November) and schematic of the seasonal cycle at select stations. Seasonal rainfall is the total for the season, in millimeters. Black vertical lines/lines at each station indicate the month or months of maximum rainfall at the indicated stations.

2. Climatic Controls and Regionalization

The general precipitation climatology of eastern Africa is extremely heterogeneous, due to the complexity of large-scale controls. These include topography, lakes, and the maritime influence, in addition to the seasonal dynamics of tropical circulation (Figure 2). The seasonality and amount of rainfall vary immensely over short distances (Figure 3), but there is a relatively strong coherence in the patterns of interannual variability. Over most of the region, mean annual rainfall is between 800 and 1200 mm. It is much higher over the highlands and much lower over northeastern Kenya and Somalia. Rainfall is concentrated in the boreal summer in the north and northwest but in the boreal winter in the southernmost sectors. Most elsewhere in eastern Africa an equatorial rainfall regime prevails, with rainfall peaks in the boreal spring and autumn. The rains of the boreal spring are concentrated in March to May, while the autumn rains are concentrated in October–November. Mean monthly rainfall in these seasons and in the boreal summer is shown in Figure 3. A peculiar exception to the prevaility seasonality occurs in the area of the Red Sea Trench, where a winter precipitation

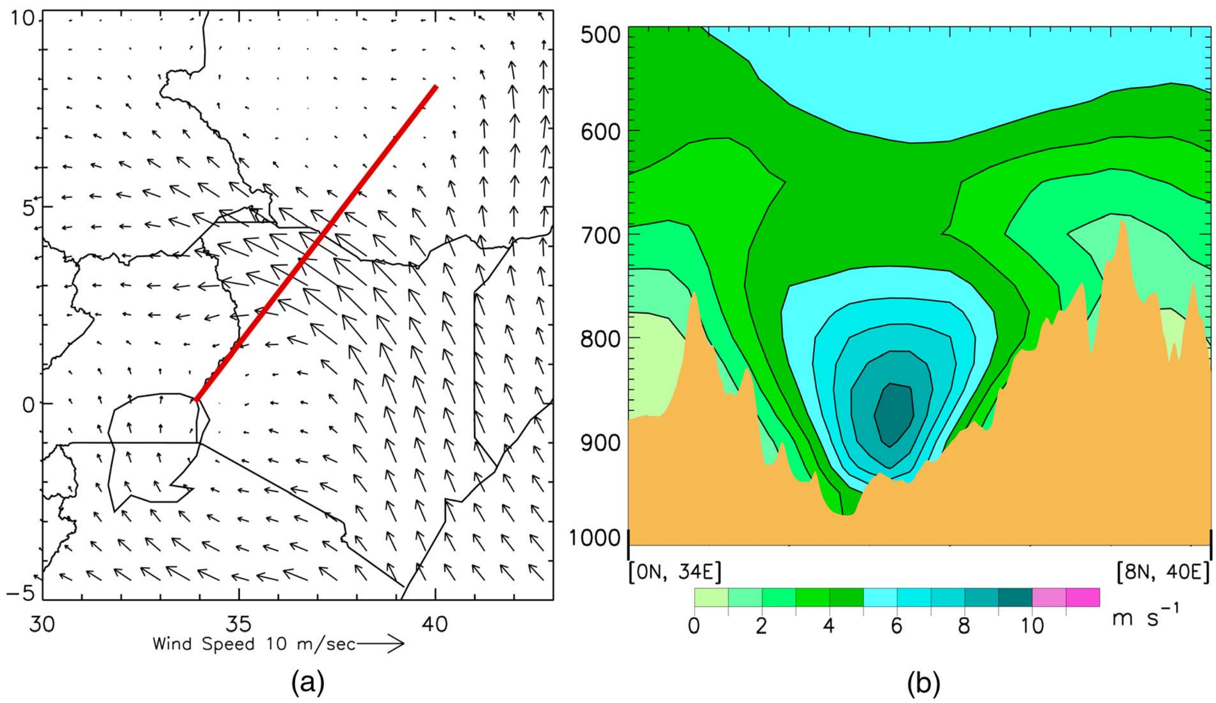

Figure 4. (a) Mean vector winds over eastern Africa at 850 hPa. (b) Profile of annual mean wind speed through the core of the Turkana jet (along the transect shown in Figure 4a) [from *Nicholson*, 2016a].

maximum prevails in the lower elevations and a summer precipitation maximum prevails some 2000 m higher up [*Flohn*, 1965].

Eastern Africa lies on the fringe of both the West African and Indian monsoons. The summer rainfall regime, especially over Ethiopia and South Sudan, is loosely linked to the West African monsoon. The Indian monsoon creates low-level easterly/northeasterly flow from November to March and predominantly southerly flow from May to October (southwesterly in Northern Hemisphere and southeasterly in Southern Hemisphere). However, the peak monsoon months (e.g., July to September and December to February) correspond to the dry seasons in the regions affected.

Classical literature on East Africa [e.g., *Trewartha*, 1961; *Griffiths*, 1972] pointed out that its dry climate is anomalous for equatorial regions and suggested that the factors in aridity are perplexing. Several factors have been suggested in the literature, including the thermal stability of the monsoon flows, the fact that they parallel the coast during the dry seasons of the boreal summer and boreal winter, and the presence of dry stable air aloft [*Nicholson*, 1996]. However, topographic influences on the flow appear to play a decisive role [*Hession and Moore*, 2011].

Oettli and Camberlin [2005] compared monthly rainfall distribution with several topographic parameters, such as slope and the mean and standard deviation of elevation. Relationships varied seasonally. For example, east facing stations were wetter during the boreal autumn but were drier during the boreal summer and winter seasons. Mean elevation appears to have little effect on rainfall amount, but it has strong control on the frequency of rainfall occurrence [*Camberlin et al*., 2014].

The impact of topography on rainfall is exerted to a large extent by its influence on the low-level flow. In the boreal summer the highlands of eastern Africa channel the flow into a strong southerly current that becomes southwesterly when it crosses the equator. This is the source of the southwest monsoon flow and Somali Jet over the Arabian Sea [*Paegle and Geisler*, 1986; *Slingo et al*., 2005]. The topography also controls the diurnal variation of the Somali Jet, as the flow tends to go *over* the high topography during the day and *around* or *along* it at night. The Turkana Channel (Figure 4) also concentrates the flow into an intense, low-level jet [*Kinuthia and Asnani*, 1982; *Kinuthia*, 1992].

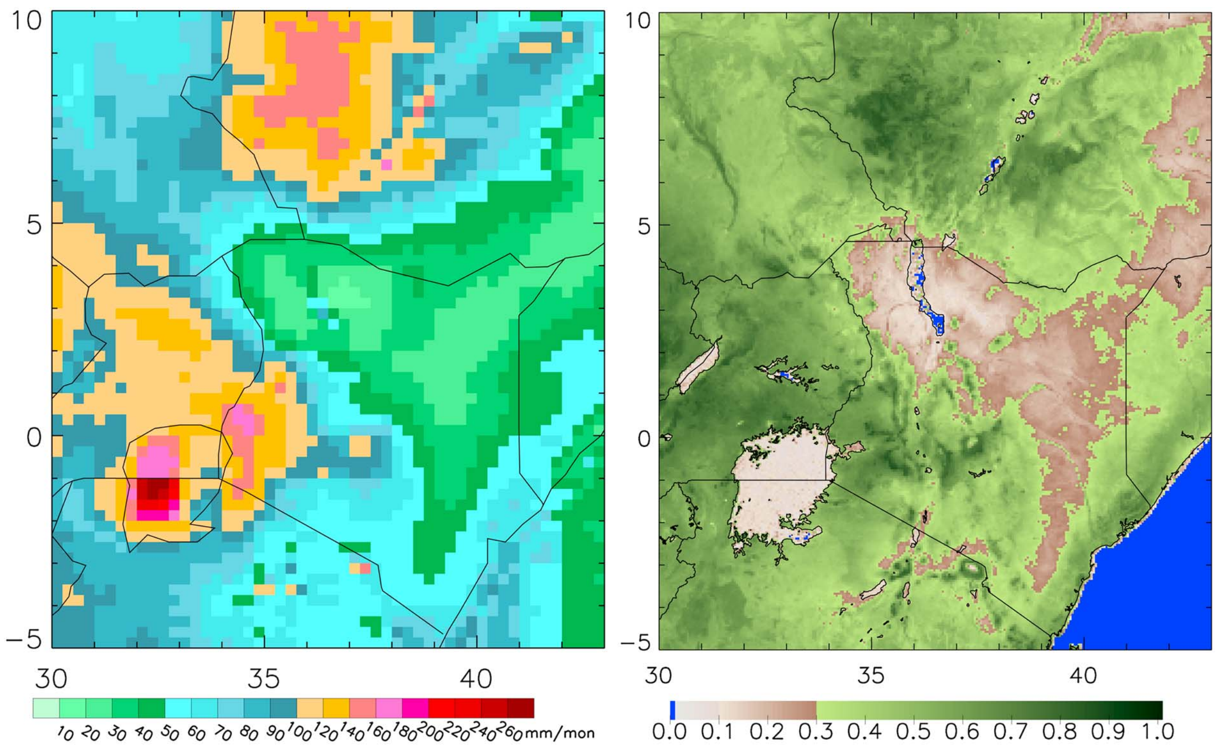

Figure 5. (left) Annual mean divergence at 850 hPa over eastern Africa. (right) Normalized difference vegetation index (NDVI, annual mean) over eastern Africa [from *Nicholson*, 2016a].

The result is large-scale divergence at low levels (Figure 5) that is associated with the monsoon flows affecting the region [*Flohn*, 1965] and with the presence of the Turkana Jet [*Nicholson*, 2016b]. The latter impacts the low-level divergence throughout the region and modulates the diurnal cycle of rainfall; the jet's intensity appears to modulate rainfall amount. The highlands also block the flow of moist air from the Congo Basin and may create lee rain shadows.

Another important aspect is the zonal circulation over the Indian Ocean [*Hastenrath et al*., 2011; *Pohl and Camberlin*, 2011; *Nicholson*, 2015a]. During most of the year the prevailing equatorial circulation is westerlies at low levels, easterlies aloft, ascent in the eastern Indian Ocean and subsidence over the western Indian Ocean and eastern equatorial Africa (Figure 6). The subsidence generally overlies a shallow region of ascent forced by the highlands. The intensity and vertical extent of the subsidence are clearly coupled with year-to-year variations in rainfall during the boreal autumn [*Smoleroff*, 2015].

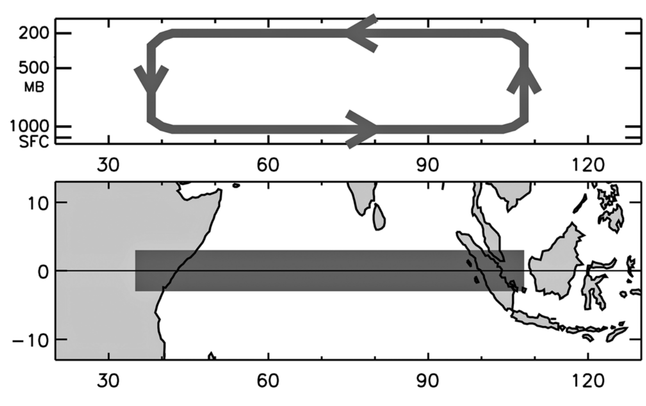

Figure 6. Schematic of the zonal circulation cell over the Indian Ocean during October–November. The shaded box in the bottom diagram shows the approximate geographical location of this cell.

Coastal influences and the effects of lakes are also factors in the regional distribution of rainfall, as well its diurnal variations and seasonality. Lake Victoria creates a mesoscale circulation system that results in a strong maximum of rainfall over the lake, a nocturnal rainfall regime in its western half, and an afternoon rainfall regime in the east. While the mean rainfall over its catchment is 1354 mm/yr, the mean rainfall over

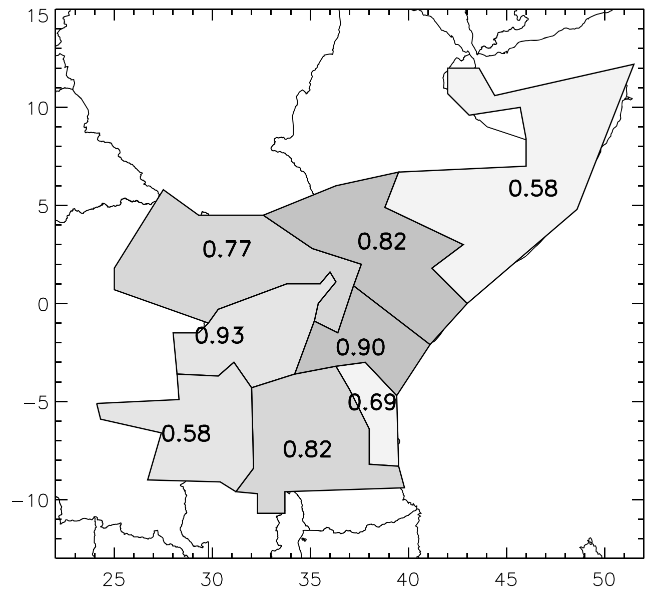

Figure 7. Correlation between annual rainfall in individual regions of eastern Africa and annual rainfall averaged over the region as a whole, based on the years 1922 to 2012.

the lake itself is roughly 1800 mm/yr [*Yin and Nicholson*, 1998; *Yin et al.*, 2000]. The mean annual rainfall on Nabuyongo Island climbs to over 3000 mm [*Flohn*, 1983]. Moreover, the lake's surface temperature has a strong influence on rainfall over the lake's basin [*Sun et al.*, 2015] and the lake also influences convective activity over the Kenyan highlands [*Okeyo*, 1987].

The mesoscale sea breeze circulation along the coast and its interaction with the monsoon flow [*Camberlin and Planchon*, 1997] produces a coastal precipitation regime that contrasts sharply with the regime farther inland. Annual rainfall is higher, and it occurs mainly at night or early morning, in contrast to the afternoon or early evening rains inland. The seasonality of rainfall is also different from inland. Rainfall occurs throughout the boreal summer months, which are dry inland, and rainfall peaks in May instead of April.

In view of the complexity in the precipitation regime, attempts have been made to regionalize the precipitation regimes of eastern Africa for purposes of prediction, agriculture, or analysis of interannual variability. Most studies have utilized principal component (PC) analysis, sometimes in combination with clustering techniques or linear correlation. These various studies differ greatly in terms of the geographical regions and seasons considered, the number of stations utilized in the analysis, and the time period evaluated.

Consequently, the number of homogeneous regions determined varies from 3 or 4 [*Camberlin and Planchon*, 1997] to 26 [*Ogallo*, 1989], with the determined number of regions often influenced by arbitrary decisions in the approach. Also, depending on the treatment of data, the result might reflect the seasonal cycle [e.g., *Basalirwa*, 1995, 1997], interannual variability [e.g., *Ng'ongolo and Smyshlyaev*, 2010], or decadal variability [*Omondi et al.*, 2013b]. The areas that frequently stand out in the regionalization schemes include the Lake Victoria region, the highlands, and the coastal plain [*Camberlin and Planchon*, 1997; *Indeje and Semazzi*, 2000; *Indeje et al.*, 2000]. Separate regions within Ethiopia should also be considered [*Elsanabary et al.*, 2014; *Zhang et al.*, 2016].

It can be argued, however, that overall, many of the regionalizations confirm a fairly strong homogeneity with respect to interannual variability, especially for the short rains season [*Moron et al.*, 2007]. In the analysis of *Indeje et al.* [2000], the first rotated empirical orthogonal function (EOF) for October-November-December (OND) showed positive anomalies throughout eastern Africa and explained 53% of the variance. *Nyenzi* [1988] showed a similar result for annual rainfall, as did *Pohl and Camberlin* [2006a, 2006b] for intraseasonal time scales. The regionalization of *Indeje and Semazzi* [2000] also showed strong coherence throughout most of Tanzania and Kenya east of the highlands.

Nicholson et al. [2012] delineated seven rainfall regions within the equatorial rainfall regime (Figure 7), but these are highly intercorrelated. The correlation of each region with that of the whole area of the equatorial rainfall regime ranges from 0.55 to 0.89 over a period of roughly 100 years. This homogeneity is also illustrated by comparing three widely spread individual stations with the time series for the region as a whole (Figure 8). In this figure rainfall is expressed as a standardized seasonal departure (i.e., the standard deviation from the long-term seasonal mean at each station, averaged over all stations in the region). This representation of rainfall is typically used in publications on African rainfall. The similarity of the interannual variability confirms that the region can be relatively well represented from a small number of stations.

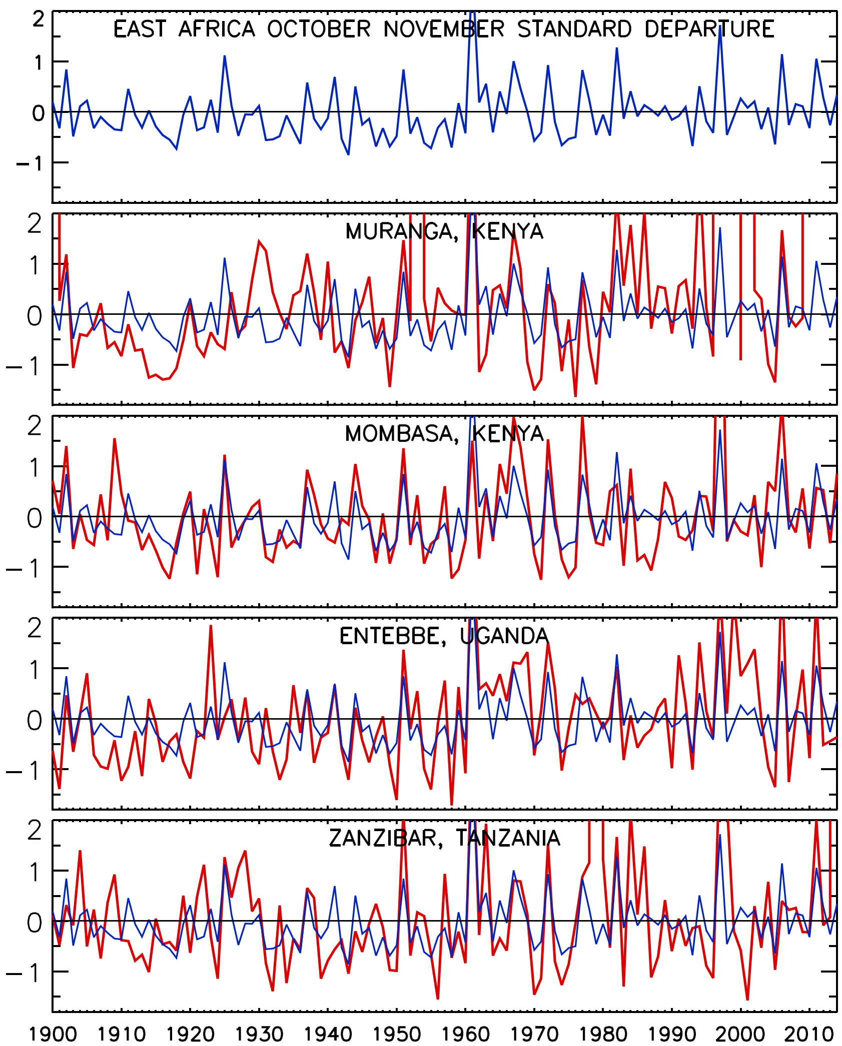

Figure 8. Standardized October–November rainfall at four individual stations (red lines) compared to the regional average (blue lines) over a network of roughly some 250 stations. Values represent a regionally averaged standard departure from the seasonal mean.

The above results thus suggest that on interannual times scales variability is relatively coherent throughout at least the equatorial rainfall region of East Africa. This coherence extends to some extent to the summer rainfall region as well. Nevertheless, the results of several studies suggest that the equatorial and summer rainfall regions should be treated separately [*Indeje et al*., 2000; *Camberlin and Philippon*, 2002; *Bowden and Semazzi*, 2007; *Nicholson*, 2014c, 2016a]. In cases where increased regional detail is desirable, the coastal, western highlands and Lake Victoria regions should also be considered separately, at least during the long rains season.

3. The Rainy Seasons

Most of East Africa experiences two rainy seasons during the course of the year. These occur in the transition seasons and are traditionally defined as March to May (MAM) and October–November (ON). This is the case in Kenya, northern Tanzania, Somalia and southern Ethiopia, Rwanda, Burundi, and most of Uganda (Figure 1). However, the western highlands and coastal regions also receive significant rainfall in the boreal summer [*Davies et al*., 1985; *Camberlin*, 1996]. Areas to the north and west, primarily Eritrea, Djibouti, northern Ethiopia, northern Uganda, and South Sudan, receive most of their rainfall in the boreal summer (July to September or JAS). Areas farther south, inland regions of central and

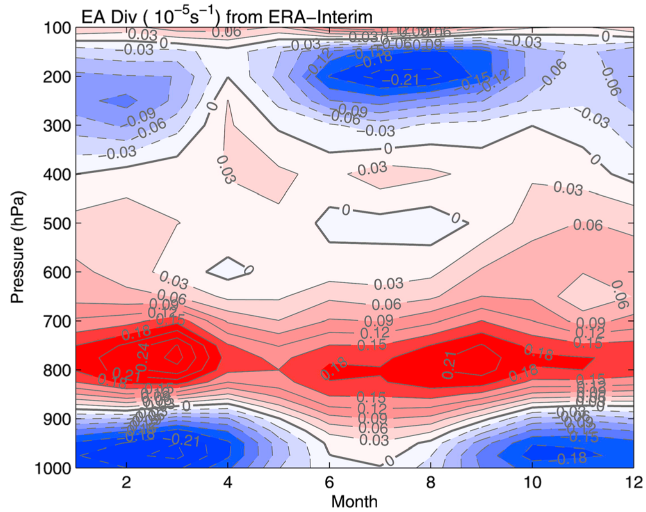

Figure 9. Annual cycles of wind divergence averaged over eastern Africa [from *Yang et al.*, 2015].

southern Tanzania, receive peak rainfall in the boreal winter. Unlike most tropical regions, organized disturbances are seldom encountered in any of these seasons [*Moron et al.*, 2007].

The MAM rains are generally termed the "long rains," and they are the heavier and longer of the two seasons. Local names for this season are *masika* in Kenya and Tanzania and *gu* in Somalia. These rains protrude into northeast Ethiopia, where they are known as the *belg* (small) rains. The "short rains" of ON are locally termed *vuli* in Kenya and Tanzania and *der* in Somalia. The long rains are more reliable, so that most of the interannual variability is associated with the short rains, although that season receives less rainfall on average [*Nicholson*, 1996; *Camberlin and Wairoto*, 1997; *Camberlin and Philippon*, 2002; *Hastenrath et al.*, 2011].

The partitioning of the rainy seasons differs among various authors, causing some confusion. The atmospheric circulation over the continent is generally fairly stable from June to September (JJAS) and from December to March (DJFM), with rapid changes during the remaining months. Consequently, some studies [e.g., *Nicholson*, 1986; *Ogallo et al.*, 1988] considered two 4 month seasons (JJAS and DJFM) and two 2 month seasons (AM and ON) [*Hastenrath et al.*, 2011]. *Camberlin and Philippon* [2002] further examined this question for eastern Africa using PC analysis. Spatial patterns shown by the PCs were similar in March and April, in July through September, in October and November, and in January and February. May and December were very different. Thus, they strongly questioned the homogeneity of the long rains season. This topic is considered later in further detail.

3.1. The Seasonal Cycle

The classic explanation of the seasonal cycle of precipitation over East Africa is the seasonal migration of the Intertropical Convergence Zone (ITCZ). A close examination of the wind regime in the region suggests that this explanation is not tenable [*Nicholson*, 2017] because of contrasts in the seasonal cycle, large-scale teleconnections, and spatial patterns of variability. The most comprehensive study of the seasonal cycle in East Africa is that of *Yang et al.* [2015]. They demonstrate that the region is convectively stable year round, with this factor playing a role in the region's general aridity. Divergence in the lower troposphere appears to play a role as well. Although regionally integrated convergence is positive during the rainy season, it is associated with coastal uplift and topographic uplift in the western highlands; surface divergence prevails over much of the region.

The changes that enhance precipitation during the two seasons appear to be increases in surface moist static energy and vertically integrated moisture flux, which is generally negative during the year but becomes weakly positive during the rainy seasons. These changes are associated with sea surface temperature (SST) changes in the western Indian Ocean, just off the coast [*Yang et al.*, 2015]. Other factors that appear to play a role in creating the seasonal cycle include the low-level Turkana Jet Stream, which appears to suppress rainfall during the boreal summer, and the intensity of the upper level subsidence [*Nicholson*, 2016b].

3.2. Contrast Between the "Long" and "Short" Rains

Several dynamical differences between the long and short rains seasons are summarized by *Camberlin and Wairoto* [1997] and *Hills* [1978]. In both seasons the general flow pattern is easterly or southeasterly at low levels, northeasterly around 700 mbar, and easterlies in the upper troposphere. However, there is stronger southerly/northerly flow at lower levels during the long/short rains, a result of different pressure

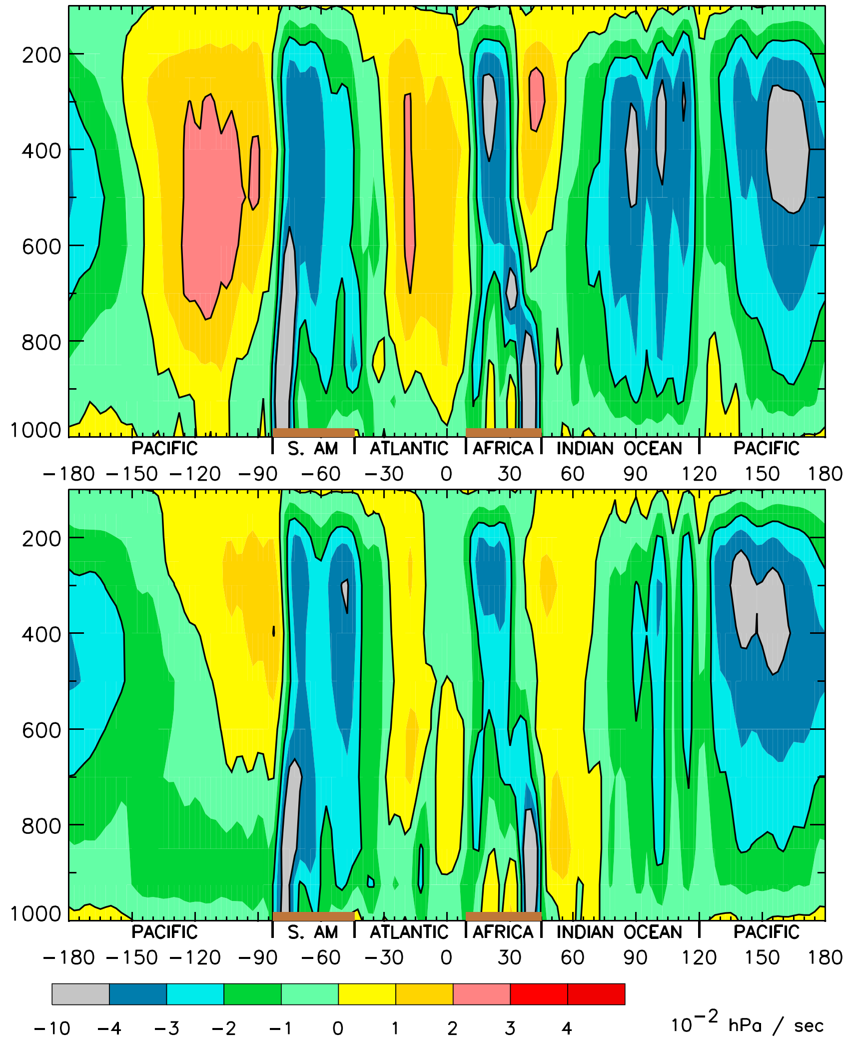

Figure 10. Vertical profile of omega (10^{-2} hPa s^{-1}) as a function of latitude averaged for the area between 10°N and 10°S. This is vertical motion, with positive values denoted subsidence and negative values denoting ascent. The top diagram is for October–November (i.e., the short rains), and the bottom diagram is for March–April–May (i.e., the long rains).

conditions over the Arabian Sea. The transition between seasons is quicker in October–November than during March–April. Examining a limited number of pibal and radiosonde stations, these authors suggested that at upper levels (300 to 200 mbar) divergence/convergence prevails during the short/long rains. An analysis based on ERA Interim data, however, suggests that during the long rains this divergence is generally limited to April (Figure 9) [*Yang et al*., 2015].

During both seasons zonal circulations prevail over the Indian Ocean, with westerly winds near the surface and easterly winds in the upper troposphere. This circulation appears to be particularly important during the short rains (Figure 10), when ascent prevails in the east and descent prevails in the west, over eastern Africa [*Ogallo et al*., 1988; *Beltrando*, 1990; *Hastenrath et al*., 1993]. During all seasons except the boreal summer there is a strong inverse correlation between the zonal winds at 850 mbar and 150 mbar on both interannual and intraseasonal time scales [*Pohl and Camberlin*, 2011].

Hastenrath [2000] suggested that a coherent zonal cell (Figure 6) exists on an interannual basis only during the boreal autumn, when the changes in zonal winds are accompanied by vertical motion anomalies at the eastern and western boundaries. Accordingly, he notes the absence of this cell during the boreal spring, while *Pohl and Camberlin* [2011] suggest that it is merely weaker than during the boreal autumn. Consistent with both views, the subsidence at upper levels over eastern Africa (Figure 10) is notably weaker during April and May than during October and November [*Nicholson*, 2017]. The contrasts in the importance of the zonal cell and in the magnitude of subsidence may help to explain why the Madden-Julian Oscillation (MJO, see section 4.3) has a much stronger influence in the long rains than in the short rains [*Berhane and Zaitchik*, 2014].

The long and short rains also exhibit strong contrasts in rainfall characteristics. For example, the spatial coherence of rainfall over eastern Africa is much greater during the short rains than during the long rains [*Nicholson*, 1996; *Moron et al*., 2007; *Hastenrath et al*., 2011]. Averaged over 26 Kenyan stations, the standard deviation of mean rainfall is 41% during the short rains versus 55% during the long rains [*Camberlin and Wairoto*, 1997]. Spatial coherence is highest at the peak of the short rains [*Camberlin et al*., 2009]. Station-to-station correlations of interannual time series are weaker in April and in May than for any other rainy months [*Camberlin et al*., 2010].

The two seasons also contrast sharply in terms of interannual variability. It is also much greater during the short rains [*Nicholson*, 1996; *Camberlin and Wairoto*, 1997]. The zonal winds at 850 mbar are more variable in OND than any other time of year [*Pohl and Camberlin*, 2011]. Also, time series of interannual variability for the months of March, April, and May are mutually uncorrelated, while those for October and November are strongly correlated throughout eastern Africa [*Nicholson and Entekhabi*, 1986]. These contrasts are largely related to the impact of El Niño–Southern Oscillation (ENSO), which is large and of the same sign throughout the OND season but weak and inconsistent during the long rains [*Beltrando*, 1990; *Rowell et al*., 1995; *Camberlin and Wairoto*, 1997; *Nicholson and Kim*, 1997; *Indeje et al*., 2000].

3.3. Contrasts Among the Months of the Long Rains

Camberlin and Philippon [2002] applied PC analysis to determine the appropriate seasonal partitioning over eastern Africa. Spatial rainfall anomaly patterns shown by the PCs were similar in March and April but quite different in May. However, spatial coherence was very different in March and April, being highest in March and weakest in April. They concluded overall that May should be considered separately from March and April when investigating atmospheric dynamics. Consistent with this is *Ogallo*'s [1988] finding that correlations between cumulative rainfall during the season and global SST anomalies are extremely low. The exception to this appears to be the Kenya coast. May rainfall is particularly high along the coast, probably a result of SSTs being higher than air temperatures [*Camberlin and Planchon*, 1997]. This suggests a strong sensitivity of coastal rainfall to coastal SSTs.

Several other studies provide additional confirmation of the inhomogeneity of the long rains season. In northern Tanzania March and April rainfall is linked to the zonal contrast between the Indian Ocean and the East African land mass and to the surface zonal winds. The factors are different in May and include the meridional temperature contrast between the Asian continent and the Indian Ocean, as well as the meridional surface winds [*Zorita and Tilya*, 2002]. Thus, May is strongly linked to the Asian monsoon. Spatial coherence and potential predictability via SSTs peak during March. Although more rainfall occurs in April and May, it is much less spatially coherent during those months [*Camberlin et al*., 2009, 2010; *Moron et al*., 2013]. The zonal circulation and subsidence aloft become progressively weaker from March to May [*Nicholson*, 2017], and the factors in predictability are notably different in each month [*Nicholson*, 2015b].

Intraseasonal differences in the teleconnections to rainfall have been shown by numerous studies [*Camberlin and Okoola*, 2003]. The patterns of rainfall variability as well as its teleconnections are dissimilar between the early (March–April) and late (May) parts of season [*Ininda*, 1999; *Camberlin and Philippon*, 2002; *Zorita and Tilya*, 2002]. *Rowell et al*. [1994] found that a late start of the long rains (reflecting March rainfall) is associated with a cool tropical Atlantic. That study also found rainfall teleconnections with the extratropical southwest Atlantic and the equatorial Indian Ocean during May. No teleconnections with SSTs were apparent in April, which the authors interpreted as evidence that the peak of the long rains is dominated by internal, chaotic atmospheric variability. The impact of El Niño is also different in each month of the long rains. There appears to be little relationship between the Southern Oscillation Index in March or April, but some relationship is evident in May [*Rowell et al*., 1994; *Mutai and Ward*, 2000]. During the post–El Niño year the impact tends to be slightly positive in March but negative in May [*Nicholson and Kim*, 1997]. The opposite is true for the El Niño year [*Indeje et al*., 2000].

Collectively, the intraseasonal contrasts described above and the lack of atmospheric or oceanic teleconnections for the season as a whole [*Ogallo et al*., 1988; *Hastenrath et al*., 1993; *Camberlin and Philippon*, 2002] strongly argue for the consideration of each month of the long rains separately in analysis and prediction [*Rowell et al*., 1994]. Notably, *Berhane et al*. [2014] found that the rainfall teleconnections in the JJAS season over East Africa are also very different in each month. They similarly argue that subseasonal analysis is required to advance the understanding and prediction of precipitation variability in the Blue Nile Basin of East Africa.

3.4. The Summer Rainy Season

During this season the bulk of the rainfall occurs over northern and northwestern parts of eastern Africa: northern Ethiopia, Eritrea, Djibouti, northern Uganda, and South Sudan. This season accounts for some 50% to 80% of the rainfall over Ethiopia's agricultural regions [*Korecha and Barnston*, 2007]. Coastal regions and the western highlands also experience summer rainfall, but the rainfall is associated with different factors.

In this region shallow lower tropospheric westerlies appear in June and persist through August [*Camberlin*, 1995; *Segele and Lamb*, 2005]. This flow is overlain by easterlies with maxima at two levels. The uppermost, the Tropical Easterly Jet, is at roughly 150 mbar and the lower, weaker maximum is attained at around 550 mbar. The low-level flow is predominantly southwesterly and increases in speed eastward, culminating in the low-level Somali Jet off shore. This low-level wind pattern is relatively steady from June to September, with southeast trades south of the equator becoming southwesterly upon transit into the Northern Hemisphere [*Nicholson*, 2014c].

The onset of the westerlies ushers in the boreal rainy season, which peaks in July and August [*Segele and Lamb*, 2005]. Rainfall rapidly diminishes in September. Low-level cyclonic flow over the Arabian Peninsula provides ascent over the Yemen Highlands and northeastern Ethiopia and is critical in the development of westward propagating convective systems that characterize the peak of the rainy season. Two confluence zones over Ethiopia favor the development of disturbances [*Segele et al*., 2009b]. A northern one over Eritrea and northernmost Ethiopia is associated with the Intertropical Convergence Zone, the southern boundary of which lies near 15°N. The one farther south lies above the Rift Valley and Djibouti. Near the surface it is associated with the trough over the Arabian Peninsula, but confluence from the monsoon trough prevails around 850 mbar.

It was long assumed that moisture transport into the region was mostly from the Atlantic, specifically the Gulf of Guinea. Very comprehensive studies of *Viste and Sorteberg* [2013a, 2013b] altered that picture. They documented three sources of moisture transport into the region: the Gulf of Guinea, the Indian Ocean, and the Mediterranean. Moisture from the Gulf of Guinea appeared to be the least important but the most variable from year to year.

3.5. The Seasonal Progression

The timing of the various rainy seasons in eastern Africa is difficult to establish [*Camberlin and Okoola*, 2003] because the definition of seasons is to some extent arbitrary. Decisions have to be made as to the definition of a rain event, as well as the thresholds in event size, cumulative rainfall, length of period without dry days, etc., which define the onset [*Segele and Lamb*, 2005]. It is also necessary to synthesize observations made from individual stations. Objective approaches to this step have been taken by *Riddle and Wilks* [2013], using principal component analysis and cluster analysis, and by *Boyard-Micheau et al*. [2013], using multivariate analysis.

For Kenya and northern Tanzania *Camberlin and Okoola* [2003] give an average start and cessation of the long rains as 25 March and 21 May, respectively, but the year-to-year variability is large. *Camberlin et al*. [2009] indicate a standard deviation of 14.5 days for the start and 10.3 days for the season end. A delayed onset is associated with cold/warm SST anomalies in the South Atlantic/Indian Ocean. This is conducive to enhanced surface divergence over East Africa. *Okoola* [1999a, 1999b] suggested that the onset of the season is triggered by extratropical southerly surges through the Mozambique Channel. An abrupt northward shift of the rains occurs in late March/early April [*Riddle and Cook*, 2008].

Toward the end of the long rains a second abrupt shift occurs. It generally coincides with a decrease in rainfall in equatorial Africa and eastern Ethiopia and an increase in rainfall in western Ethiopia [*Riddle and Cook*, 2008]. The end of the long rains is strongly correlated with the onset of the Indian monsoon (r = 0.55 over 44 years) [*Camberlin et al*., 2010]. The cessation of the rains is associated with the strengthening of the southerly winds (the Somali Jet), which enhances wind divergence and wind shear [*Camberlin et al*., 2010], and with the onset of active convection over the Arabian Sea [*Okoola*, 1998].

The boreal summer rains, termed the *kiremt* season in Ethiopia, provide 65% to 95% of the rainfall over northern Ethiopia [*Segele and Lamb*, 2005]. The rains start in early to middle March over SW Ethiopia and progress northeastward. However, the rains of March to May are generally associated with the equatorial rainy season that prevails farther south in Kenya, Somalia, and Tanzania. In central Ethiopia, shallow westerly monsoon flow typically appears 5–10 days before the onset. At the same time a cooling and depression of dew point occur. The northeasternmost regions of Ethiopia generally do not receive rainfall until middle to late July. The cessation of the season progresses from NE to SW, with the end of the season generally ranging from early September to middle or late November. The standard deviations of the onset and cessation are generally on the order of 8 to 20 days, being greatest where season is longest.

Camberlin et al. [2009] give average start and end dates for the short rains in Kenya and northern Tanzania as 23 October and 26 December, respectively. The standard deviations are 24.7 and 20.4 days for the start and end. This is roughly twice the variability they calculated for the long rains. In the region evaluated the spatial mean of rainfall during the short rains was 226 mm, compared to 357 mm in the long rains. Interannual variability is 77% and 38% for the two seasons, respectively. The amount of rainfall received during the short rains is correlated most strongly with the onset date (−0.83 over 30 years) and the length of the season (0.91 over 30 years). For the short rains the intensity of rainfall per day and the percent of rain days during the season are almost as important as the onset date in determining the amount of rainfall received during the season. However, the most important variable is the length of the season. The correlation with the onset date is −0.78, compared with 0.89 for the season length.

4. Intraseasonal Variability

Intraseasonal variability essentially refers to the wet and dry spells that occur within the rainy season, particularly the temporal rhythm of their occurrence. The time scale is intermediate to synoptic and seasonal [*Kijazi and Reason*, 2005]. In East Africa, generally 3 to 4 wet spells of roughly 5 to10 day duration comprise the rainy season [*Kabanda and Jury*, 2000]. Extensive research has been done on the time scales of variability and the dynamic factors associated with the intraseasonal variability. Fewer papers have considered such characteristics as intensity, duration, and frequency.

This section summarizes research on several aspects of intraseasonal variability. These include its temporal character and propagation of convection, links to zonal wind anomalies, and links to the Madden-Julian Oscillation (Figure 11). Specifically, the relationship of the MJO to intraseasonal and interannual variability, to ENSO, and to a dipole relationship between eastern and western regions is examined.

4.1. Seasonal Versus Intraseasonal Rainfall Characteristics

Several papers have examined such characteristics as mean daily rainfall intensity, frequency of dry spells, and number or frequency of rain days and compared them with seasonal characteristics such as onset, cessation, and total seasonal rainfall amount [*Moron et al*., 2007; *Camberlin et al*., 2009; *Gitau et al*., 2013]. Results have been very consistent. The most spatially coherent variable is number or frequency of rain days, followed by seasonal total. There is little coherence in daily rainfall intensity or the frequency of dry spells longer than 5 days. The spatial coherence of all variables is distinctly greater in the short rains than in the long rains. The greatest spatial coherence is evident at the peak of the short rains. Spatial coherence during the long rains is greatest early in the season.

The seasonal total is strongly correlated with season onset for both the long and short rains [*Camberlin and Okoola*, 2003]. It is also correlated with the date of cessation for the short rains but not for the long [*Camberlin et al*., 2009; *Philippon et al*., 2015]. The onset is strongly negatively correlated with both the length of the season and the total amount of rainfall, but the onset and cessation appear to be uncorrelated. For the long rains, the onset and cessation are also independent of the rainfall intensity or number of days of rainfall.

The strongest correlation among seasonal statistics is between the total amount of rainfall and the season length: 0.91 for the long rains for the period 1958 to 1987 and 0.89 for the short rains. The intensity and

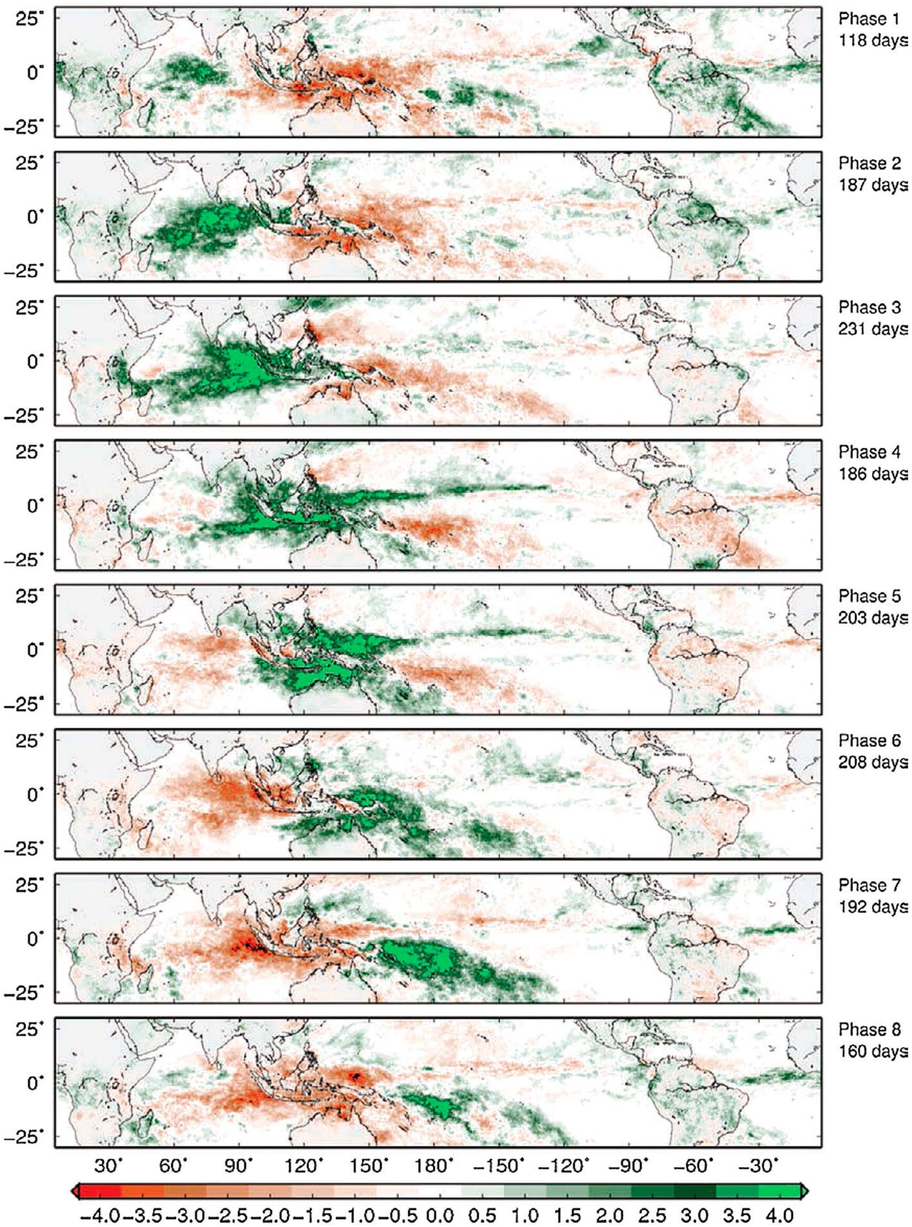

Figure 11. Composites of intraseasonal (30–90 day) anomalies in Tropical Rainfall Measuring Mission precipitation (mm d^{-1}) during November–April of 1998–2012 [from *Zhang*, 2013]. Each diagram portrays an individual phase of the MJO. The eastward progression between phases 1 and 8 is clearly seen.

number of rain days are strongly correlated with total seasonal rainfall during the short rains. During the long rains seasonal rainfall is only moderately correlated with those statistics but is strongly correlated with the number of extreme wet events [*Pohl and Camberlin*, 2006a]. Farther south, in Tanzania, rainfall during the long rains is correlated with both the onset and end and thus with season length [*Kijazi and Reason*, 2012].

Camberlin and Okoola [2003] found that a late onset of the long rains is correlated with positive sea level pressure over the Indian Ocean and negative sea level pressure over the South Atlantic. This pressure pattern is associated with abnormally warm temperatures over the African continent and South Atlantic, weakened trade winds, a weakened St. Helena High, and easterly zonal wind anomalies in the midtroposphere. These factors are evident as early as January or February. In the late onset years, the abnormally high pressure over the Indian Ocean is associated with midlevel easterly wind anomalies over equatorial Africa.

A zonal wind index for 20–35°E, 5°N–5°S at 500 mbar is correlated at −0.71 with the onset date of the long rains. The correlation between 700 mbar zonal winds and the onset date is −0.75. Thus, westerly anomalies promote an early onset date, probably by promoting convergence and uplift over East Africa. The westerly anomalies are also associated with stationary or transient troughs extending from the subtropics to the equator. These are particularly effective when the troughs extend on both sides of the equator, creating the "equatorial bridge" circulation pattern described by *Johnson and Mörth* [1960]. Evidence of interaction with such troughs was also put forward by *Pedgley* [1966], *Habtemichael and Pedgley* [1974], *Shukla and Mo* [1983], and *Okoola* [1999a, 1999b].

For the short rains the leading mode of intraseasonal variability is associated with the covariability of ENSO and the Indian Ocean Dipole [*Bowden and Semazzi*, 2007]. During warm/positive events, pentad rainfall is consistently above normal during the entire season, despite strong intraseasonal fluctuations. However, the cold/negative cases are not mirror images. While overall rainfall is below normal during the season, pronounced wet and dry spells occur within it. Looking specifically at rainfall over coastal Tanzania, *Kijazi and Reason* [2005] found that the increased October to December rainfall during El Niño events is associated with an early onset, fewer dry spells, and more intense rainfall compared to long-term averages. However, the impact is not uniform along the coast [*Gamoyo et al.*, 2015]. During La Niña, the season starts late, rainfall is less intense, and the number of dry spells is above average.

4.2. Zonal Winds

Many early papers demonstrated a link between midlevel westerly/easterly wind events and wet/dry spells [*Nakamura*, 1969; *Kiangi and Temu*, 1984; *Davies et al.*, 1985; *Camberlin and Wairoto*, 1997; *Okoola,* 1999a, 1999b]. However, there is some geographic variation to these associations [e.g., *Camberlin*, 1996]. This generalization is valid for coastal regions, but the opposite relationship exists in the western highlands of the Kenya/Tanzania region. This appears to be true for all three of the rainy seasons of eastern Africa.

Camberlin [1996] demonstrated this for the boreal summer season, when the Kenya coast and the western highlands receive significant amounts of rainfall. In the highlands wet/dry spells are linked to westerly/easterly 700 mbar wind anomalies in western Kenya. The opposite anomaly pattern prevails near the coast. There rainy spells are generally common to the whole coast and are well defined in time and space. The heaviest rain events involve interaction between the sea breeze and monsoon circulations and depend on three major features: (1) the intensity of the land/sea breeze, (2) strong easterly to southeasterly wind anomalies within the 850 mbar to 700 mbar layer (hence an acceleration of the Somali Jet), and (3) sea surface temperatures that are higher than air temperatures [*Camberlin and Planchon*, 1997].

Pohl and Camberlin [2006b] showed that an inverse relationship between the coast and the western highlands also prevails in the long and short rains. In both seasons wet/dry spells over the highlands/coast tend to occur synchronously with westerly anomalies over equatorial East Africa at 850 mbar. At upper levels (200 mbar) easterly anomalies prevail instead, indicating a zonal circulation cell over East Africa and the western Indian Ocean. In the highlands the westerly wind anomalies produce moisture advection from the Congo Basin and are associated with high pressure over the Atlantic and low pressure over the Indian Ocean. The opposite pressure pattern prevails during wet spells in the east, when easterly anomalies advect moisture from the Indian Ocean. *Kijazi and Reason* [2012] found that farther south, over Tanzania, the region received moisture from both the Congo Basin and the Indian Ocean during wet spells.

In the central highlands near Nairobi a somewhat different picture emerges. Easterly anomalies are associated with dry spells in the long rains, but with wet spells during the short rains. *Camberlin and Wairoto* [1997] based this conclusion on an analysis limited to the months of April and November and suggested that orographic uplift is the cause of the link between easterlies and wet spells during the short rains. They also noted that the northeast trades are also accelerated during wet spells and suggested this resulted from an "equatorial duct" situation, with high-pressure cells on either side of the equator [*Johnson and Mörth*, 1960]. That mechanism is consistent with the one described by *Hastenrath* [2007], with a weakening of the Indian Ocean zonal cell reducing subsidence over East Africa.

4.3. Relationship to the MJO

Several studies have shown that the intraseasonal variability over eastern Africa does not appear to be random but instead is concentrated on specific time scales. The time scale varies from 20 to 80 days but is strongest around 40 to 50 days [e.g., *Sandjon et al*., 2012], i.e., the time scales of the MJO [*DeMott et al*., 2015]. Rainfall here is strongly coherent with MJO amplitude on this time scale. Overall, the MJO explains some 40% of the intraseasonal variance in convective activity in central equatorial Africa, including parts of eastern Africa [*Pohl and Camberlin*, 2006b; *Sandjon et al*., 2014a, 2014b]. It appears to be the dominant factor controlling intraseasonal variability over East Africa during both the long and short rains [*Pohl et al*., 2005]. *Anyamba* [1984] traced similar temporal patterns over the western Indian Ocean, with 40 to 50 day activity likewise being dominant there.

Wet conditions tend to occur in the first half of the MJO cycle along the coast and in the second half of the MJO cycle in the highlands [*Hogan et al*., 2015]. The contrasting rainfall conditions found in the two regions for the opposite MJO phases are strongly correlated with the pressure gradient between the Indian and Atlantic Oceans [*Pohl and Camberlin*, 2006b]. This pattern appears to reflect different rain-producing mechanisms. MJO phases leading to wet spells in the western (highland) region are those associated with the development of large-scale, deep convection in the Africa-Indian Ocean region. These events are fueled by low-level westerly moisture advection. MJO phases leading to wet spells in the eastern (coastal) region are often those associated with overall suppressed deep convection in the Africa-Indian Ocean region. In these phases moisture advection is from the Indian Ocean and stratiform rainfall or shallow convection is likely responsible for the wet spells.

Three major centers of particularly intense convection are evident over equatorial Africa and are centered over the northern Congo, southwestern Tanzania, and southern Ethiopia [*Sandjon et al*., 2014b]. The convective anomalies that generate rainy spells are coupled with anomalies in the zonal wind, particularly low-level westerly anomalies. The convective cells appear to commence in the equatorial Atlantic, especially the Gulf of Guinea [*Mutai and Ward*, 2000; *Tazalika and Jury*, 2008]. Eastward propagating features are dominant (roughly 62%), which is consistent with the eastward propagation of the MJO signal. The propagation is apparent in zonal winds, outgoing longwave radiation, and rainfall [*Mpeta and Jury*, 2001; *Kijazi and Reason*, 2005; *Sandjon et al*., 2012]. Consistent with the aforementioned studies, *Chan et al*. [2008] found that snowfall on Mount Kilimanjaro is associated with waves of convective activity propagating from west to east.

Pohl and Camberlin [2006a] found that the MJO also modulates the interannual variability of the long rains season and extreme daily events within the season. However, its contribution varied greatly from year to year. The MJO amplitude was found to explain some 44% of the interannual variability for the region as a whole and for the western highlands during the period 1979 to 1995, as well as 64% of the variability of the onset date of the season. However, its influence was minimal in the eastern coastal area, where rainfall is strongly correlated with the cessation date of the rainy season but not with the onset. For extreme events, the MJO explained overall 33% of the variance in the western highlands but had a small negative correlation with extreme events at the coast.

Overall, when the MJO is very strong the long rains season commences earlier, the number of wet spells is greater, and daily rainfall is enhanced during March and April, compared to years when the MJO is very weak [*Pohl and Camberlin*, 2006a]. A dramatic shift occurs in middle to late April. The MJO's influence in the western highlands becomes negligible, while at the coast its influence is reversed. Thus, at the coast its influence over the entire long rains season is close to zero. *Berhane and Zaitchik* [2014] found that its overall influence is strongest early and late in the long rains but in midseason for the short rains. Their results suggested that the mechanism of influence is zonal shifts in the MJO convective center, with both dynamic factors (convergence) and thermodynamic factors (advection and thermal stability) playing a role in the modulating of rainfall.

Several studies have shown evidence that the MJO's influence on intraseasonal variability is modulated by ENSO but results are somewhat contradictory. Examining the year as a whole, *Sandjon et al*. [2012] found that the MJO is strong during La Niña or ENSO-neutral years but weak or absent during strong El Niños. *Kijazi and Reason* [2005] reached a similar conclusion. Consistent with this, *Sandjon et al*. [2014a] found that the intraseasonal oscillations in East Africa were strongest and most variable during the 1981 to 1990 decade, when ENSO was also extremely variable. Looking specifically at the long rains, *Pohl and Camberlin* [2006a] found that MJO activity is weak during the El Niño phase but did not find evidence of strong MJO activity in the La Niña phase.

The association between the two factors appears to be their common impact on zonal circulation over the Indian Ocean. *Pohl and Camberlin* [2011] used the zonal wind shear between 150 mbar and 850 mbar over the central equatorial Indian Ocean as an index of this circulation. They found that the shear and the MJO have up to 60–65% common variance for periods between 30 and 60 days. ENSO and the wind shear together explain almost 55% of the variance on interannual time scales. The shear is particularly strong when the rainfall-enhancing phase of the MJO over Africa occurs during an El Niño, amplifying the seasonal departures generally associated with El Niño. This most often occurs during the long and short rains but seldom during the boreal summer.

5. Interannual Variability

Much of the early work on interannual variability over East Africa focused on spectral analysis. *Ogallo* [1979], *Rodhe and Virji* [1976], and *Nicholson and Entekhabi* [1986] all found three very prominent spectral peaks at roughly 2.3, 3 to 4, and 5 to 6 years. These same three spectral peaks characterize the Southern Oscillation Index (SOI) and coherence between the SOI and East African rainfall is strong on all three time scales. This suggests strong forcing by ENSO. Further evidence of the link to ENSO was provided by *Ropelewski and Halpert* [1987], *Ogallo et al*. [1988], *Kiladis and Diaz* [1989], *Hutchinson* [1992], *Beltrando and Camberlin* [1993], *Nicholson and Kim* [1997], *Indeje et al*. [2000], and *Nicholson and Selato* [2000]. Only a handful of papers suggested links to the Atlantic or Indian Oceans [e.g., *Nicholson and Nyenzi*, 1990; *Ogallo et al*., 1994], but these links appear to be strongest on interdecadal time scales [*Omondi et al*., 2012, 2013a, 2013b]. Thus, the prevailing view became one in which ENSO was the primary driver of interannual rainfall variability in eastern Africa, with El Niño/La Niña resulting in anomalously wet/dry conditions and affecting primarily the short rains.

Papers by *Goddard and Graham* [1999], *Saji et al*. [1999], *Webster et al*. [1999], and *Hastenrath* [2000] were instrumental in broadening that view. The first of these studies showed that atmospheric response to Indian Ocean variability was required to simulate ENSO's impact on rainfall in eastern, as well as southern Africa. *Nicholson et al*. [2001] used observations to demonstrate this for southern Africa. *Hastenrath* [2000], in agreement with *Goddard and Graham* [1999], showed the critical role of the zonal circulation over the Indian Ocean in transmitting ENSO's signal.

Webster et al. [1999] and *Saji et al*. [1999] demonstrated the existence of an SST mode in the Indian Ocean that is closely linked to rainfall variability in eastern Africa. Termed the Indian Ocean Zonal Mode (IOZM) or Indian Ocean Dipole (IOD), this mode consists of opposite SST anomalies in the eastern and western tropical Indian Ocean. The positive phase of the mode is characterized by warm/cold anomalies in the western/eastern pole and above average boreal autumn rains over eastern Africa. The importance of the IOZM in modulating these rains is further evidenced by numerous subsequent papers [e.g., *Black et al*., 2003; *Clark et al*., 2003; *Saji and Yamagata*, 2003; *Black*, 2005; *Slingo et al*., 2005; *Behera et al*., 2006; *Ummenhofer et al*., 2009; *Nakamura et al*., 2011; *Nicholson*, 2015a].

Pohl and Camberlin [2011] showed that both interannual variability and intraseasonal variability over equatorial East Africa are strongly controlled by zonal wind shear (difference between 850 mbar and 150 mbar winds) over the central equatorial Indian Ocean. This conclusion held for all seasons except the boreal summer, when the winds at both levels are relatively constant. Most of the variability is associated with the lower level winds. They also found that the shear is more closely correlated with ENSO than with the IOZM in all seasons but the boreal summer.

Much of the recent literature has dealt with the relative importance of the Pacific versus the Indian Ocean in driving interannual variability, the relationship between ENSO, and the Indian Ocean Zonal Mode and changes over time in these roles and relationships. These issues are all reviewed in this section in the context of the interannual variability of the short rains. The interannual variability of the boreal summer and boreal spring or long rains is also reviewed.

5.1. The Long Rains of March to May

Although the long rains produce more rainfall than the short rains over most of East Africa, they are generally less variable. Perhaps because of that, few studies examined this season until it became apparent that the long rains have been declining dramatically [*Funk et al*., 2008; *Williams and Funk*, 2011; *Lyon and Dewitt*,

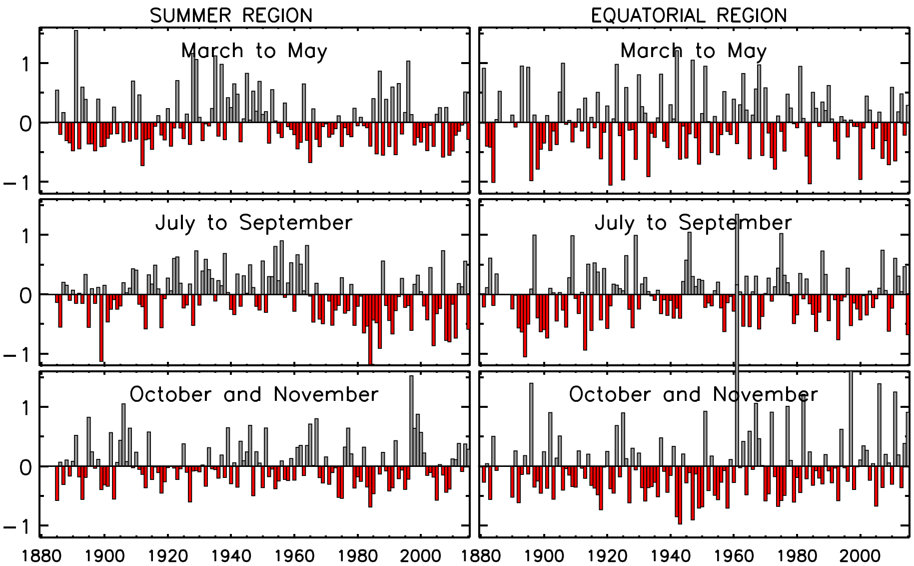

Figure 12. Time series of March to May, July to September, and October–November rainfall for the summer and equatorial rainfall regions shown in Figure 1. Values represent a regionally averaged standardized departure from the seasonal mean, as in Figure 6. Data are from the Centennial Trends data set [*Funk et al.*, 2015]. The value for October–November 1961 is 3.1.

2012; *Yang et al*., 2014]. The decline is apparent in both the summer and equatorial rainfall regions (Figure 12). In the summer region an abrupt shift occurred in the late 1990s from predominantly very wet years to an almost continuous period with well below average rainfall. In the equatorial region rainfall steadily declined in the three decades following the early 1980s and a series of very dry years prevailed around 2010.

Possible factors in this decline are discussed in sections 6.2 and 8.5. Before this decline spawned interest in this season, most information on factors governing the long rains was derived in the context of continental-scale studies of ENSO. *Liebmann et al*. [2014] stressed that the sources of interannual variability of the long rains have been difficult to pin down, in part because MAM rainfall appears to be only weakly constrained by SST anomalies on interannual time scales. Moreover, seasonal rainfall reflects the net influence of factors operative on intraseasonal, interannual, and decadal time scales [e.g., *Pohl and Camberlin*, 2011; *Omondi et al*., 2012].

Nicholson and Kim [1997] found that the net impact of El Niño on the MAM season tends to be insignificant because anomalies switch sign in middle of season, being positive in March of the post–El Niño year but negative in May and close to zero in April. *Nicholson and Selato* [2000] similarly found little impact on MAM rainfall during the La Niña year but some reduction of rainfall in the post–La Niña year.

The results of *Hoell et al*. [2014] further help to explain the weak ENSO signal. They showed that La Niña can either increase or decrease MAM rainfall in East Africa, depending on the characteristics of the episode. Two types of La Niña, both with cold anomalies in the central and eastern Pacific and neutral conditions in the western Pacific, tend to produce weak positive rainfall anomalies in MAM. Two types with cold but weaker anomalies in the eastern and central Pacific but warm anomalies in the western Pacific tend to produce strong negative MAM rainfall anomalies. The various types produce different patterns of subsidence and ascent over East Africa. Both the extent and magnitude of the dry anomaly is closely linked to the degree of subsidence, but divergence of moisture flux also plays a role.

The most comprehensive study of the interannual variability of the long rains was that of *Pohl and Camberlin* [2006a]. They found that interannual variability and seasonal extremes are strongly influenced by the Madden-Julian Oscillation (MJO). They compared its influence in the western highlands and farther east in the coastal regions of Kenya and Tanzania. Wet anomalies were associated with westerly low-level wind anomalies in the west but easterly anomalies in the east. This configuration in wind anomalies was preferentially seen in distinct phases of the MJO. The influence of the MJO varied from year to year but overall explained 44% of the year-to-year variance in the long rains between 1979 and 1995.

Zorita and Tilya [2002] examined the long rains in northern Tanzania. They showed, in agreement with *Camberlin and Philippon* [2002], that interannual variability in March and April is linked to factors different from those associated with May rainfall. March and April appear to be controlled by the zonal temperature contrast between the Indian Ocean and adjacent East African land mass, thus by surface zonal winds. May rainfall is linked to a meridional temperature contrast between the Indian Ocean and the Asian land mass and thus to meridional surface winds. This provides a link to the Indian monsoon. High May rainfall in northern Tanzania is associated with positive rainfall anomalies in Indian rainfall in June to August.

Hoell and Funk [2013] pointed out the need to consider the interaction between factors operative on interannual and decadal time scales to fully understand the variability of the long rains. *Omondi et al*. [2012] underscored the 10 year periodicity in the low-frequency variance of the long rains and showed strong but regionally dependent correlations with SSTs in the Indian Ocean. *Omondi et al*. [2013a, 2013b] showed that each ocean exerts some control on the decadal time scale, but the links are weaker than with the short rains.

Liebmann et al. [2014] likewise noted that different factors appear to be operative on interannual and decadal time scales and also pointed to a disparity between model results and observations. Models suggest that the role of SSTs may be larger on decadal time scales, especially SSTs in west to central Pacific and western Indian Oceans. Within an ensemble of model runs SSTs explained up to 40% of the variance of the long rains. Observed correlations with SSTs were found to be weak but within the range of the individual ensemble runs. Correlation with the east-west SST gradient across the equatorial western Pacific was −0.22 for observations but −0.74 for the model ensemble average. *Williams and Funk* [2011] and *Hoell and Funk* [2013] suggested that this increased gradient induces an intensification of convection over the Indo-Pacific warm pool, which then leads to increased subsidence over eastern Africa. Sensitivity to this gradient appears to be increasing in recent decades [*Hoell and Funk*, 2013], possibly contributing to the disparity between observations and model results.

5.2. June Through September: The Boreal Summer

There have been few papers on the boreal summer rainy season (JJAS) over eastern Africa. This is the dominant rainy season in roughly the northern half of Ethiopia and in parts of South Sudan, western Kenya, and most of Uganda. Traditionally, it is assumed that the summer rains are related to airflow from the Congo Basin and the Atlantic. However, most research indicates that the Pacific and Indian Oceans are also important in the development of this season. Time series of the rainfall in this season are shown in Figure 12. In the summer rainfall region a decline occurred in the 1960s and rainfall has remained relatively low, while interannual variability has increased since the late 1980s. Such a change is not apparent in the equatorial rainfall region.

The first intensive studies of interannual variability in this season are those of *Camberlin* [1995, 1997], who examined only regions with a summer rainfall peak, such as Ethiopia, western Kenya, and much of Uganda. These studies documented a strong relationship between summer rainfall and the Indian monsoon, a relationship that was relatively stable throughout the twentieth century, as well as a strong relationship with ENSO and the Pacific Walker cell. El Niño/La Niña episodes were shown to be associated with drought/wet seasons, a finding confirmed by *Segele et al*. [2009a, 2009b] and *Nicholson and Selato* [2000]. Even though ENSO influences both eastern Africa and the Indian monsoon during this season, the teleconnection between the two regions appears to be independent of ENSO. A particularly strong inverse relationship was shown between boreal summer rainfall in eastern Africa and surface pressure at Bombay. During the period 1953–1988, the common variance was 79%.

In central Ethiopia the summer rains have a significant, positive correlation with the SOI not only simultaneously but also at a 3 month lag. In this region summer rainfall, and especially September rainfall, is inversely correlated with pressure and SST over Indian Ocean. Correlations are also very strong correlation with SSTs in the Arabian Sea. Teleconnections between the summer rains in northern Ethiopia-Eritrea and either ENSO or the Indian Ocean are much weaker [*Beltrando and Camberlin*, 1993].

Several studies examined the relationship between the summer rains and circulation anomalies. Very robust relationships with the low-level westerlies and upper tropospheric easterlies over western and central Africa and with the strength of the Somali Jet have been shown [e.g., *Camberlin*, 1995; *Segele et al*., 2009a, 2009b].

Evidence suggests that the enhanced westerlies are associated with high pressure over the Gulf of Guinea and an anomalously deep monsoon trough over the Arabian Peninsula. The westerlies transport moist air from the Congo Basin so that wet seasons are associated with large water vapor transport convergence across Ethiopia and a deep moist layer [*Segele et al*., 2009b; *Williams et al*., 2012].

Other associations with JJAS rainfall have also been documented. *Williams et al*. [2012] found that the southern tropical Indian Ocean strongly influences rain in this season over eastern Africa, consistent with its relationship to the Indian monsoon. They also confirmed that the transport of moist air from the Congo Basin plays a role in interannual variability of boreal summer rainfall in eastern Africa. *Omondi et al*. [2013a] and *Segele et al*. [2009b] suggested that all three oceans influence this season, with the Pacific playing the greatest role. On decadal time scales areas of greatest impact appeared to be the equatorial central Pacific, the tropical and South Atlantic, and the SW Indian Ocean [*Omondi et al*., 2013a]. *Segele et al*. [2009a] found similar links on annual time scales. *Omondi et al*. [2012] found Atlantic SST precursors to the June to August rainfall on decadal time scales. March–May SSTs in the tropical Atlantic correlated positively/negatively with rainfall in western/eastern East Africa. The reverse relationship was found for June–August SSTs.

Segele et al. [2009b] looked at links to global SSTs and atmospheric variables on four time scales evident in the variability of boreal summer rainfall in eastern Africa: seasonal, annual, quasi-biennial (QB), and ENSO. The seasonal and annual cycles were in phase, with wet/dry conditions associated with enhanced/reduced lower tropospheric southwesterlies from the Atlantic. On the QB time scale wet anomalies corresponded to anomalously strong Azores and Saharan highs and a strong TEJ over the Arabian Sea. In contrast, the Mascarene high, the TEJ, and the monsoon trough were all anomalously strong during the wet phase of the ENSO time scale (roughly every 3 to 5 years). Links to SSTs were shown to be strongly positive/negative with SSTs in the equatorial western/eastern Pacific on both QB and ENSO time scales. Opposite conditions prevailed in abnormally dry seasons. *Segele et al*. [2015] found that the influence of Indian Ocean SSTs depends strongly on the magnitude and location of the SST anomalies.

Several studies pointed to additional factors. *Segele et al*. [2015] found that wet boreal summer seasons over the Horn of Africa are associated with enhanced zonal and meridional integrated moisture fluxes and a steep vertical decrease in moist static energy. *Zeleke et al*. [2013] confirmed the links with the TEJ, low-level westerlies form the Atlantic, southeasterly trades over the Indian Ocean, and the Mascarene high and added to the picture the importance of the St. Helena high over the Atlantic. The low-level winds are critical in the transport of moisture into the region. *Diro et al*. [2011a, 2011b] added two further factors in the variability of boreal summer rainfall in Ethiopia, SSTs in the midlatitude Pacific Northwest and in the Atlantic Gulf of Guinea. They also underscored the importance of separately considering subregions of Ethiopia in evaluating the interannual variability.

Parts of the equatorial rainfall region receive some rainfall during the boreal summer, but few papers have examined the interannual variability of these rains. *Segele et al*. [2009a, 2009b] and *Zeleke et al*. [2013] considered some areas in the equatorial regime in their analyses of rainfall throughout Ethiopia. However, little emphasis was placed on those sectors. *Zeleke et al*. [2013] showed a tendency for an opposition in summer rainfall between the summer and equatorial rainfall regions of Ethiopia and found that the most important factors affecting the former had little influence on the latter.

The most detailed study of summer rainfall in the equatorial region is that of *Pohl and Camberlin* [2011], who examined JJAS rainfall within the sector of eastern Africa that receives predominantly transition-season rainfall (i.e., between 5°N and 5˚S). No significant relationships with ENSO, the IOZM, or circulation over the Indian Ocean were found. *Nicholson and Kim* [1997] and *Nicholson and Selato* [2000], respectively, examining El Niño and La Niña, came to a similar conclusion. *Nicholson* [2016b] suggested that the Turkana Jet serves to diminish summer rainfall in this region and hypothesized that interannual variability in this season may be strongly influenced by the intensity of this jet.

5.3. The Short Rains of the Boreal Autumn

In contrast to the long rains, the short rains have been increasing in recent years in the equatorial rainfall region (Figure 12). The most marked changed occurred in 1961, and high rainfall has persisted since that time. Since the mid-1980s rainfall has seldom been far below the long-term mean and a number of really wet years occurred. In the summer rainfall region a few remarkably wet years have occurred within the last two decades. These are discussed further in section 6. However, rainfall generally declined since the early 1970s.

The short rains are strongly coupled to a zonal vertical circulation cell in the central equatorial Indian Ocean [*Hastenrath et al*., 2011; *Mutai et al*., 2012]. This is frequently referred to as a Walker cell or a Walker-type circulation. At low levels westerly winds prevail, a result of a steep eastward pressure gradient. At 200 mbar easterly flow prevails. Rising motion in the east and subsidence in the west, near eastern Africa, complete the cell. Numerous studies clearly demonstrate that the most important physical mechanism in the variability of the short rains is the intensity of this cell, with the low-level westerlies playing a fundamental role in modulating this cell. For rainfall averaged over coastal stations the correlation with these low-level westerlies for the period 1958 to 1997 is −0.85 [*Hastenrath et al*., 2011]. However, the link extends throughout a much larger sector of eastern Africa. *Nicholson* [2015a] found that for the entire region with a bimodal rainy season, the correlation is −0.74 over the period 1874 to 2012. Both that study and *Bergonzini et al*. [2004] showed that the short rains are more strongly modulated by the low-level winds over the Indian Ocean than by ENSO.

Strong westerlies are favored by a steep eastward pressure gradient and abnormally weak trade winds in the South Indian Ocean [*Beltrando and Camberlin*, 1993; *Mutai et al*., 2012; *Dezfuli and Nicholson*, 2013]. The correlation between coastal rainfall and this pressure gradient is −0.70 for the period 1958 to 1997 [*Mutai et al*., 2012]. Changes of pressure in the western Indian Ocean frequently modify the gradient. In some cases, the pressure gradient actually reverses, changing the low-level flow to easterly and resulting in extremely abundant rainfall [*Birkett et al*., 1999].

Further evidence of the importance of this zonal circulation comes from studies showing relationships between the short rains and other components of the vertical cell. The weakening of the equatorial westerlies acts to reduce the subsidence over East Africa. The correlation between this subsidence at 500 mbar and coastal rainfall is +0.60 [*Mutai et al*., 2012]. At the same time the easterly return flow in the upper troposphere is also weakened [*Dezfuli and Nicholson*, 2013; *Nicholson*, 2015a], so that the upper level and low-level zonal winds are inversely coupled. Over the period 1874 to 2012 the correlation between zonal winds at the surface and 200 mbar over the central equatorial Indian Ocean is −0.65 and the correlation between 200 mbar winds and the short rains is 0.50.

Changes over the continent contribute to the picture as well. Some of the earliest literature on interannual variability emphasized the role of low-level westerlies in enhancing rainfall during the short rains [e.g., *Nakamura*, 1969; *Vincent et al*., 1979; *Anyamba*, 1984; *Nicholson*, 1996]. Anomalously high OND rainfall is associated with an increased in upper level easterlies and a decrease in midlevel easterlies across the equatorial belt, as well as an increase in low-level westerlies [*Schreck and Semazzi*, 2004; *Dezfuli and Nicholson*, 2013]. The increased westerly flow out of the Congo basin coupled with the stronger easterlies over the western Indian Ocean enhances low-level convergence in the region.

5.4. The Roles of ENSO and the Indian Ocean in the Interannual Variability of the Short Rains

Conditions over the Indian Ocean are clearly part of the mechanism by which the short rains are modulated. The changes in the Walker circulation described above are part and parcel of the aforementioned IOZM. Warm SSTs in the west and cold SSTs in the east (positive phase of the IOZM) are associated with a weakened Walker circulation over the Indian Ocean, while the reverse SST pattern of the IOZM's negative phase acts to enhance the Walker circulation over the Indian Ocean.

On the other hand, there is also strong evidence of links to the Pacific, particularly ENSO [e.g., *Nicholson and Entekhabi*, 1987; *Ropelewski and Halpert*, 1987; *Schreck and Semazzi*, 2004; *Bowden and Semazzi*, 2007]. October–December rainfall tends to be markedly enhanced during El Niño years and reduced during La Niña years [*Nicholson and Kim*, 1997; *Nicholson and Selato*, 2000; *Dezfuli and Nicholson*, 2013; *Hoell et al*., 2014]. This association between El Niño and above average rainfall/La Niña and below average rainfall is not evident in all episodes. During the period 1901 to 1991, the short rains were above average in only of 12 of 20 El Niño episodes and below average in only 12 of 17 La Niña episodes. Notably, most wet years did occur in conjunction with El Niño but most dry years were not associated with La Niña [*Nicholson and Kim*, 1997; *Nicholson and Selato*, 2000]. However, the links to ENSO vary on a regional basis within East Africa and also depend on the time scale (e.g., interannual versus interdecadal) and on the spatial mode of

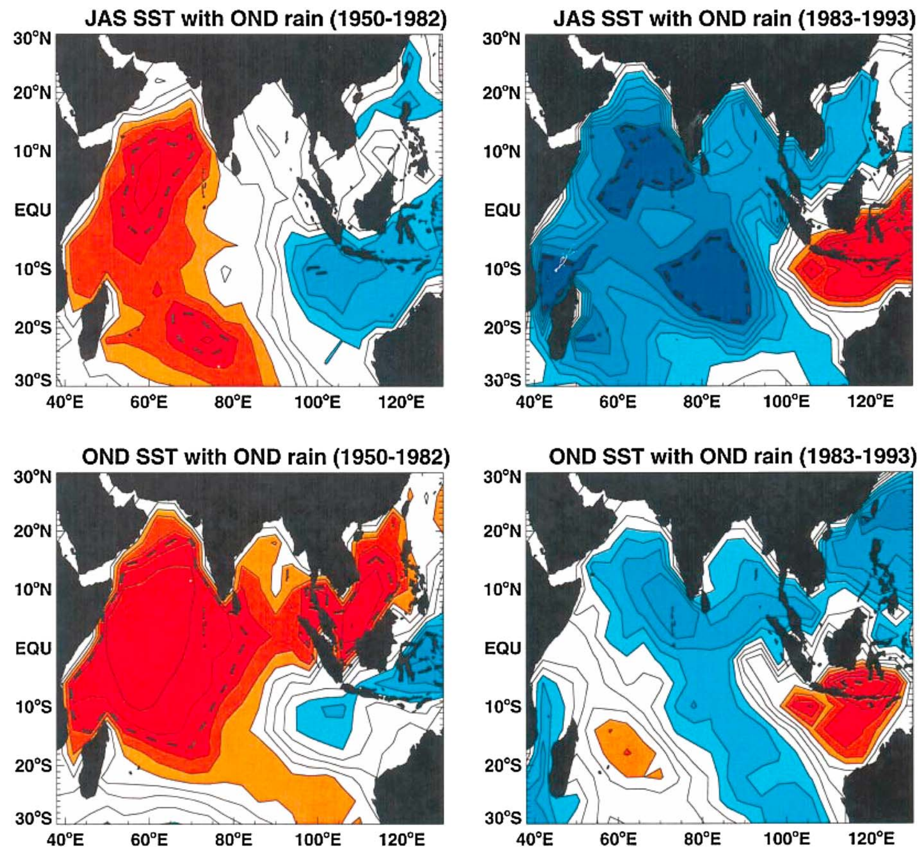

Figure 13. Seasonal correlations of Indian Ocean SST with the OND East African rainfall for the periods 1950 to 1982 and 1983 to 1993. (top) Correlation with July to September SST. (bottom) Correlation with OND SST [from *Clark et al*., 2003].

rainfall variability, being particularly strong for the mode with anomalies of the same sign throughout East Africa [*Omondi et al*., 2012, 2013a, 2013b].

ENSO's impact during the short rains also depends on the characteristics of the episode. Observations showed that the anticipated impact of La Niña/El Niño materialized only when cooling/warming of the tropical Indian and Atlantic Oceans occurred in conjunction with the episode [*Nicholson et al*., 2001]. Modeling experiments of *Goddard and Graham* [1999] produced similar results. *Hoell et al*. [2014] found that the short rains are generally reduced during La Niña but that the degree and spatial consistency of the reduction depends on the nature of the episode.

The strong evidence for control by both the Pacific and Indian Oceans, as well as the Atlantic [e.g., *Onyutha and Willems*, 2015], has resulted in considerable controversy concerning the primary driver of the short rains. *Goddard and Graham* [1999] and others have suggested that the Pacific forces the changes in the Indian Ocean that are linked to interannual variability in eastern Africa. *Black et al*. [2003] postulate that ENSO predisposes the Indian Ocean coupled system to an IOZM event and conclude that only the large events (those that reduce the Indian Ocean SST gradient) can produce extreme rainfall. Some of the observed rainfall teleconnections to ENSO, including the opposition between eastern and southern Africa, are probably related to synchronous occurrences of ENSO and the IOZM [*Manatsa et al*., 2011] or the links between ENSO and the IOZM [*Black*, 2005].

Notably, ENSO is well correlated with parameters over the Indian Ocean that modulate the short rains, such as the low-level and upper level zonal winds [*Nicholson*, 2015a]. During El Niño years the easterly flow over the Indian Ocean is increased and westerly flow over the equatorial continent is increased [*Schreck and Semazzi*, 2004], changes that are typically associated with abnormally high rainfall during the short rains. The high phase of the Southern Oscillation tends to weaken the trade winds in the South Indian Ocean and to

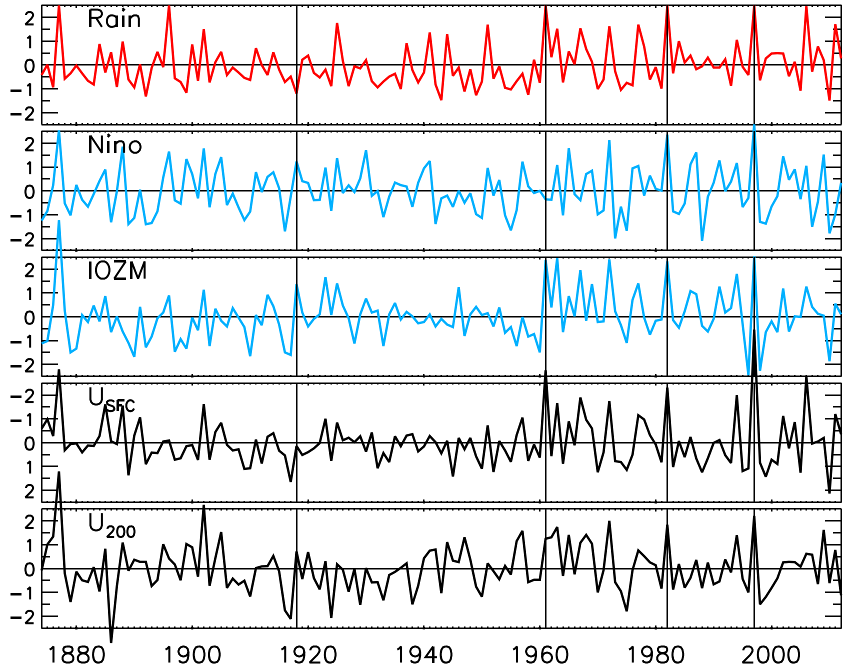

Figure 14. Time series of October–November rainfall, surface zonal winds, 200 mbar zonal winds, IOZM, and Niño 3.4, 1874–2012 [from *Nicholson*, 2015a]. Units for zonal winds and the IOZM are standard deviations. Units for rainfall are the standardized departures, as defined in the text. Note that the *y* axis for the surface winds is inverted, with negative/positive values above/below the zero line. Vertical lines indicate regime shifts identified by several authors.

lower pressure in the east, thus enhancing the pressure gradient [*Mutai et al*., 2012]. This increases the westerly flow as well as the subsidence and moisture flux divergence over eastern Africa [*Hoell et al*., 2014], so that La Niña favors reduced rainfall over eastern Africa.

Studies directly comparing the influence of the Pacific and Indian Oceans have generally concluded that the latter has more control on eastern Africa's short rains on both the interannual [e.g., *Bergonzini et al*., 2004] and interdecadal [e.g., *Omondi et al*., 2013a, 2013b] time scales. *Liebmann et al*. [2014], for example, found that over the period 1979 to 2012 the short rains correlated with the Indian Ocean Dipole index at 0.80, but the correlation with eastern Pacific SSTs was only 0.64. Similarly, *Nicholson* [2015a] found that the correlation during the period 1874 to 2012 was 0.61 with the Indian Ocean Dipole Index but only 0.49 with SSTs in Niño 3.4.

5.5. Changes in the Relationships Among the Short Rains, ENSO, IOZM, and Zonal Winds

The disparity among the various studies might be explained at least in part by the fact that the relationships are nonstationary. *Clark et al*. [2003] were the first to demonstrate the dramatic changes on decadal time scales. The correlation between Indian Ocean SSTs and the boreal autumn of East Africa rains actually reversed during the period 1983 to 1993 (Figure 13). *Nicholson* [2015a] systematically examined these relationships over a period of 139 years. The study considered the SSTs of Niño 3.4, the IOZM, and the zonal winds over the central equatorial Indian Ocean (the sector examined by *Hastenrath et al.* [2011]) at the surface and 200 mbar. These were correlated with each other and with rainfall over the period 1874 to 2012. The study demonstrated abrupt changes in the interrelationships and in the characteristics of variability. These occurred at times identified by other authors as major regime shifts in the tropics [e.g., *Torrence and Webster*, 1999; *Clark et al*., 2003; *Ihara et al*., 2008; *Manatsa et al*., 2012; *Manatsa and Behera*, 2014]: 1918, 1961, 1983, and 1997. These shifts are evident in Figures 14 and 15. Interestingly, each of the last three shifts was marked by a tremendously high rainfall event in October–November and an extremely positive IOZM event. Only the last two shifts coincided with El Niño events.

Prior to 1918, October–November rainfall was generally below average, the exception being the year 1877, in which a major El Niño occurred. ENSO variance was strong [*Torrence and Webster*, 1999], the western pole of the Indian Ocean dipole was anomalously colder than the eastern pole, and negative dipole events

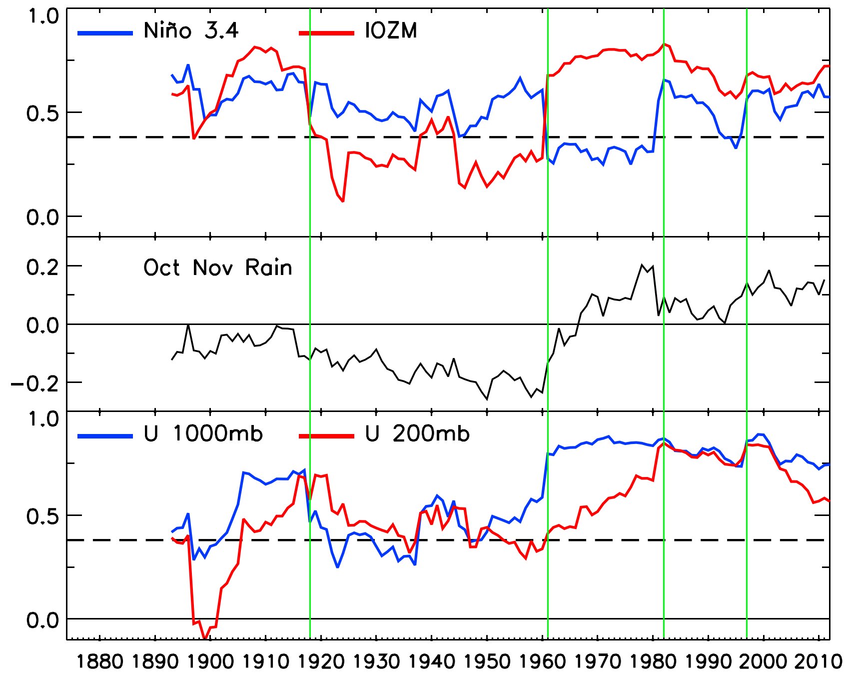

Figure 15. Sliding 20 year correlations between October–November rainfall and four tropical indices [from *Nicholson*, 2015a]. Note that the axis is inverted for surface winds. Values are plotted at the final year of the 20 year interval. Values exceeding ±0.38 are significant at the 5% significance level. Center diagram shows 20 year mean October–November rainfall. Vertical lines indicate regime shifts identified by several authors.

associated with La Niña occurred more frequently than positive dipole events [*Ihara et al.*, 2008]. Coupling was strong among all components of the system, but the dominant factor in rainfall variability appeared to have been the IOZM.

After 1918 October–November rainfall fell markedly and remained very low from then to 1960. Coincident with this, the coupling between rainfall and the other variables became markedly weaker and the coupling among the zonal winds, ENSO, and the IOZM became very weak. *Ihara et al*. [2008] showed that the IOZM and ENSO were relatively independent during this period. The link between the October–November rains and the IOZM was particularly weak and the 20 year sliding correlations with rainfall generally did not even reach the 5% significance level during this interval [*Manatsa and Behera*, 2013; *Nicholson*, 2015a]. ENSO became the dominant factor in rainfall variability.

The variability of ENSO [*Webster et al*., 1999], the IOZM [*Li et al*., 2013; *Manatsa and Behera*, 2013], and surface zonal winds over the Indian Ocean [*Nicholson*, 2015a] was also very low during this period. Collectively, these facts suggest a disruption or at least weakening of the zonal circulation cell over the Indian Ocean.

In 1961, a year of an extreme rainfall event, an abrupt shift from ENSO to IOZM dominance occurred and persisted until 1997. At this time the correlation between ON rainfall and both surface winds and the IOZM increased abruptly [*Manatsa and Behera*, 2013; *Nicholson*, 2015a], while the correlation with Niño 3.4 abruptly decreased and remained below the 5% significant level until the 1980s. The coupling with ENSO was slow to evolve, as was the coupling with the 200 mbar zonal winds. As this coupling evolved, ON rainfall also increased. Between 1961 and 1996 the correlation between rainfall and surface zonal winds fluctuated between 0.75 and 0.90; correlations between the surface zonal winds and the IOZM were similarly high. Note that the abrupt shift in correlations was not just a result of the inclusion of the extreme event of 1961. A similar abrupt shift is evident even when 1961 is excluded from the correlations.

After 1960 the variance of ENSO [*Torrence and Webster*, 1999], the IOZM [*Manatsa et al*., 2012], the zonal winds, and ON rainfall [*Nakamura et al*., 2009, 2011; *Nicholson*, 2015a] also increased abruptly and remained high. Strong and frequent positive IOZM events occurred in conjunction with El Niño [*Ihara et al*., 2008] and higher-frequency events on quasi-biennial and 4 to 5 year time scales became dominant.

The next two regimes shifts circa 1982 and circa 1997 were not as abrupt or as pronounced. *Nicholson* [2015a] demonstrated a shift following 1982, noting vastly different links between circulation parameters and drought after that time. *Clark et al*. [2003] showed that the correlation between OND rainfall along the Kenya-Tanzania coast and SSTs reversed sign throughout the Indian Ocean and western Pacific during the period 1983–1993 (Figure 13). From 1983 to 1993, the variance of rainfall and the zonal winds was extremely weak, despite strong variability in both the IOZM and ENSO, and the IOZM was primarily in the positive phase. The coupling of rainfall to winds, ENSO, and the IOZM also weakened throughout this interval. A subsequent shift occurred in the mid-1990s. *Manatsa et al*. [2012] indicate that it occurred in 1997, in agreement with *Nicholson* [2015a]. However, coupling was clearly re-established during the high rainfall event of 1994, and from then on the IOZM stayed primarily in the negative phase (Figures 14 and 15). Its variance also increased at that time [*Manatsa et al*., 2012].

Manatsa et al. [2015] identified another potential factor in the interannual variability of the short rains, the Southern Annular Mode or SAM. This mode consists of a meridional dipole in the Indian Ocean pressure field, with one pole over the tropics and the other over the extratropics. The regime shifts identified in the SAM, 1917, 1961, and 1997, coincide with those identified in the short rains. However, in contrast to the short rains, the SAM exhibits a pronounced rising trend commencing early in the twentieth century and continuing through the rest of the century.

From 1983 onward rainfall exhibited markedly higher correlations with the two zonal wind indices than with ENSO or the IOZM. The coupling was extremely high among all five variables, although weakest with ENSO. It was particularly high between the surface and 200 mbar zonal winds and between these winds and ON rainfall (Figure 15). Despite the relatively high coupling, the association between drought and large-scale forcing was markedly weaker after 1982.

Nicholson [2015a] examined this by calculating a “failure rate” of four drought indicators: strong negative anomalies in Niño 3.4, strong positive anomalies in the IOZM, strong westerly zonal winds over the central equatorial Indian Ocean, and strong easterly winds above them at 200 mbar. The “failure” rate was defined as the percent of cases in which a given indicator suggested drought, but ON rainfall instead exceeded the long-term mean. Prior to 1982 the failure rate was only 6% to 17%, depending on the indicator considered. Between 1982 and 2012 the failure rate was 44% to 60%. The breakdown of the coupling between drought and ENSO was particularly strong. Before 1982, rainfall exceeded the long-term mean in only 6% of the La Niña 3.4 cases. Of the 10 La Niña episodes between 1982 and 2012, ON rainfall was above average during 4, near average in 3, and drought occurred during only 1.

In summary, major shifts in the magnitude and variability of October–November rainfall and the four circulation parameters occurred in 1918, 1961, 1983, and circa 1994–1997. Prior to 1918 the variability of all five parameters was relatively high, there was substantial coupling among them, the IOZM was predominantly negative, and October–November rainfall was predominantly near or just below the long-term mean. The IOZM appeared to have the greatest control on rainfall. During the period 1919 to 1960, the variability of all parameters was markedly weaker, the coupling between them was extremely weak, the IOZM hovered near the long-term mean, but the magnitudes of Niño 3.4, the two zonal wind indices and October–November rainfall were all markedly lower. ENSO appeared to have the greatest control on rainfall, along with 200 mbar zonal winds. From 1961 to 1982 the IOZM was generally positive and was a stronger factor in rainfall than was ENSO, rainfall was very high, the variance of all parameters was high, and the coupling was strong. However, from 1961 onward the factor most closely linked to rainfall was the surface zonal winds. After 1982 the primary changes included a reduction in rainfall and a reduction in the variance of all parameters except Niño 3.4. The links to ENSO and 200 mbar winds became stronger, that to the IOZM became weaker, but the coupling between rainfall and the IOZM was still stronger than that between rainfall and ENSO.

It is interesting to note that the regime shifts described above were associated with distinct changes in the global Walker circulation [*Nicholson*, 2015a]. The Indian Ocean Walker cell was very weak during the period 1948 to 1960 and became increasingly stronger during the three successive periods (see section 6.3). The descending pole of the Pacific cell became correspondingly weaker in each period. Between 1948 and 1996 the Atlantic Cell became progressively stronger and the ascending pole over the Pacific became progressively weaker. However, this trend was reversed during the 1997 to 2012 period.

The post-1982 increase in the Indian Ocean Walker cell, which would increase both the surface westerlies and 200 mbar easterlies, is consistent with the infrequent occurrence of very wet years after 1982. However, it appears to be inconsistent with the near absence of drought after that time. Between 1983 and 2012 there were 7 years in which drought did not occur despite at least one predictor suggestive of drought and none indicative of above average rainfall. Three of those cases occurred during the period 1983–1994, when links to SSTs were reversed [*Clark et al*., 2003]. *Smoleroff* [2015] showed that in the cases of "failed" drought, the subsidence from middle to high levels over eastern Africa was anomalously weak and the ascent in low levels was anomalously strong, despite the stronger Walker cell over the Indian Ocean. This situation may have resulted from a strengthening of the Atlantic Walker Cell, which reached its maximum intensity during the period 1983 to 1996.

5.6. Nonlinearity of Relationships

The factors most closely associated with the interannual variability of the short rains of eastern Africa include ENSO, the Indian Ocean Zonal Mode, and the surface and upper level zonal winds over the central equatorial Indian Ocean. However, several studies have suggested that the relationships are nonlinear [*Black et al*., 2003; *Mistry and Conway*, 2003; *Manatsa et al*., 2011; *Mutai et al*., 2012; *Dezfuli and Nicholson*, 2013], such that the links to the various factors and the coupling among the factors are much weaker in dry years than in wet years. As an example, ENSO is extremely negative in only 10 of 23 drought years, but it is extremely positive in 15 of 19 exceedingly wet years [*Nicholson*, 2015a].

Most studies of the link to rainfall are based on simple linear correlation. When the consistency of the relationships is evaluated for wet and dry years, it becomes clear that most of the correlation is due to linkages for wet years [*Nicholson*, 2015a]. Moreover, in years with well above normal rainfall, generally either three or four of these factors are involved. In stark contrast, during drought years generally only one or two of the factors contributed. In some drought cases, none of the four factors identified above were operative, indicating that additional factors must be sought. *Manatsa et al*. [2014] have identified one such factor, zonal displacements of the Mascarene High, especially its eastern ridge. It plays a much greater role in drought years than in wet years.

6. Recent Trends and Extreme Events

6.1. Floods Events

Eastern Africa is well known for both extreme floods and extreme droughts, sometimes occurring within the same year. The most extreme flood event within the modern record occurred in 1961 [*Conway et al*., 1982; *Flohn and Burkhardt*, 1985; *Flohn*, 1987; *Conway and Hulme*, 1993]. Lakes Victoria rose several meters. At stations in northern Kenya November rainfall was several times the monthly mean and close to the annual mean [*Nicholson*, 2011]. For example, Wajir, Eldama, and Lokitaung received 612, 402, and 302 mm, respectively, compared to November means of 58, 48, and 39 mm. Rainfall anomalies in Lake Victoria's catchment were as much as 200% of the mean [*Thompson and Mörth*, 1965]. The intensity of the event in these locations may have been exacerbated by feedback related to the low-level jet stream in the Turkana Channel [*Nicholson*, 2016a, 2016b].

The 1961 event was not associated with El Niño, but a similar event occurred during the 1877 El Niño, likewise producing enormous transgressions of the East African lakes [*Nicholson,* 1998a, 1998b, 1999]. Other flood events occurred in 1896, 1982, 1997, 2006, and 2011 [*Nicholson*, 2014b, 2015a, 2016a]. All occurred during the short rains. The 1997 event was associated with El Niño, which generally produces an increase in the short rains of 15% to 25%. However, in 1997 the increase was 20% to 160% and Lakes Victoria, Tanganyika, Malawi, and Turkana each rose roughly 2 m. The event also raised the seasonal minima of the Sudd marshes and Lake T'ana [*Birkett et al*., 1999]. Rainfall anomalies around Lake Turkana were on the order of 100% to 200% of the mean. The impact of El Niño may have been enhanced by a continuation of flooding conditions in December–January 1997/1998 associated with strong warming in the southeastern Indian Ocean [*Latif et al*., 1999].

Fewer meteorological studies have examined floods in the summer rainfall region. *Billi et al*. [2015] documented an increase in the frequency of flash floods in some areas of Ethiopia, most notably around Dire Dawa. Some have suggested that this is linked to land use change, but an increase in precipitation intensity appears to be a more important factor. *Jury* [2011] evaluated floods in the central highlands of Ethiopia during the period 1997 to 2007. The most extreme events occurred in July of 2006 and July of 2007. These were associated with anomalously strong southerly monsoon flow over the western Indian Ocean.

6.2. Recent Trends

In more recent years, droughts have been of bigger concern than floods. They have become more frequent and more severe, extending in some cases over two or more rainy seasons [*Funk*, 2012; *Hoell and Funk*, 2014; *Nicholson*, 2016a]. This has heightened concern about possible climate change over eastern Africa. The greatest changes appear to have occurred during the long rains, which have been declining over decades.

Early evidence of the decline of the long rains was provided by *Funk et al*. [2005, 2008]. *Lyon and Dewitt* [2012] further examined this question and suggested that the decline commenced abruptly around 1999. *Williams and Funk* [2011] and *Liebmann et al*. [2014] found a long, downward trend rather than an abrupt decline. This trend commenced in the early 1980s, after which time the frequency of drought increased in much of the summer rainfall region, as well as in the equatorial region [*Funk et al*., 2013]. *Maidment et al*. [2015] confirmed the March to May decline but found that the assessed magnitude depended on the data set used in the analysis, varying from −14 to −65 mm/yr per decade.

Although both the equatorial and the summer rainfall regions experienced a decline in the MAM rains (Figure 12), the trend was weaker in the latter [e.g., *Rosell*, 2011; *Mengistu et al*., 2014]. Four major droughts occurred in this region in 1978/1979, 1984/1985, 1994/1995, and 2003/2004 [*Bayissa et al*., 2015], but these were not evident in the summer rainfall region as a whole [*Nicholson*, 2016a]. At the stations of Combolcha and Hayk in the central Ethiopian highlands rainfall declined by 25% to 30% after 1996 [*Rosell and Holmer*, 2015]. The decline was offset by an increasing trend in the July to October season [*Rosell*, 2011]. Thus, annual rainfall generally increased. However, droughts have been more frequent throughout Ethiopia since roughly 1999 [*Viste et al*., 2013].

Liebmann et al. [2014] and *Rowell et al*. [2015] confirmed the decline in the equatorial rainfall region in the March to May season and also showed that the October to December season has become wetter. *Degefu and Bewket* [2015] similarly showed a decrease in drought frequency in the Omo-Ghibe River Basin of Ethiopia, which is in the equatorial rainfall region. In the ON season the upward trend is weak because of strong interannual variability forced by ENSO and the IOZM [*Hoell and Funk*, 2014]. In the both seasons the strongest trends were in the central and southern portions of the Horn (northeastern Kenya and southeastern Ethiopia) and in the Lake Turkana basin. In this basin, a regression analysis indicated a decline from roughly 300 mm to 140 mm by the end of the regression period, 1979 to 2012.

Fewer studies have considered the trends in the JJAS season, but they have confirmed a decline similar to that in the boreal spring. This is particularly evident in the summer rainfall region (Figure 12). JJAS rainfall steadily declined since the 1950s [*Williams et al*., 2012]. *Funk et al*. [2012] calculated that the decline was 15.7 mm per decade between 1979 and 2010. In southern Ethiopia rainfall declined in both the July to September season and February to May seasons [*Viste et al*., 2013]. Examining 1960 to 2002, *Cheung et al*. [2008] found a downward trend in July to September in several watersheds in southwestern and central parts of Ethiopia. In the equatorial rainfall region (Figure 12) a weak decreasing trend is evident following the 1960s but there has been some recovery in recent years.

Some exceptions to the above trends have been noted by various authors. However, in most cases a strict comparison is not possible because of the diversity of time periods and geographical locations evaluated. At stations in the Lake Victoria basin, there was a predominantly positive rainfall trend during the last half of the twentieth century [*Kizza et al*., 2009], following the relatively dry conditions of the 1950s. In the Greater Horn of Africa the trends in annual precipitation between roughly 1960 and 2010 were mixed and generally insignificant [*Omondi et al*., 2014]. The negative trends tended to be in the westernmost sectors. During this same period a reduction in lake levels and river flow in the Lake Naivasha Basin occurred, but commensurate trends in rainfall could not be found [*Odongo et al*., 2015]. Studies of rainfall in northern Ethiopia [*Conway et al*., 2004; *Kiros et al*., 2016] found trends in northern Ethiopia to be generally insignificant and spatially diverse. *Schmocker et al*. [2016] came to a similar conclusion for the OND rainfall in the Mount Kenya region.

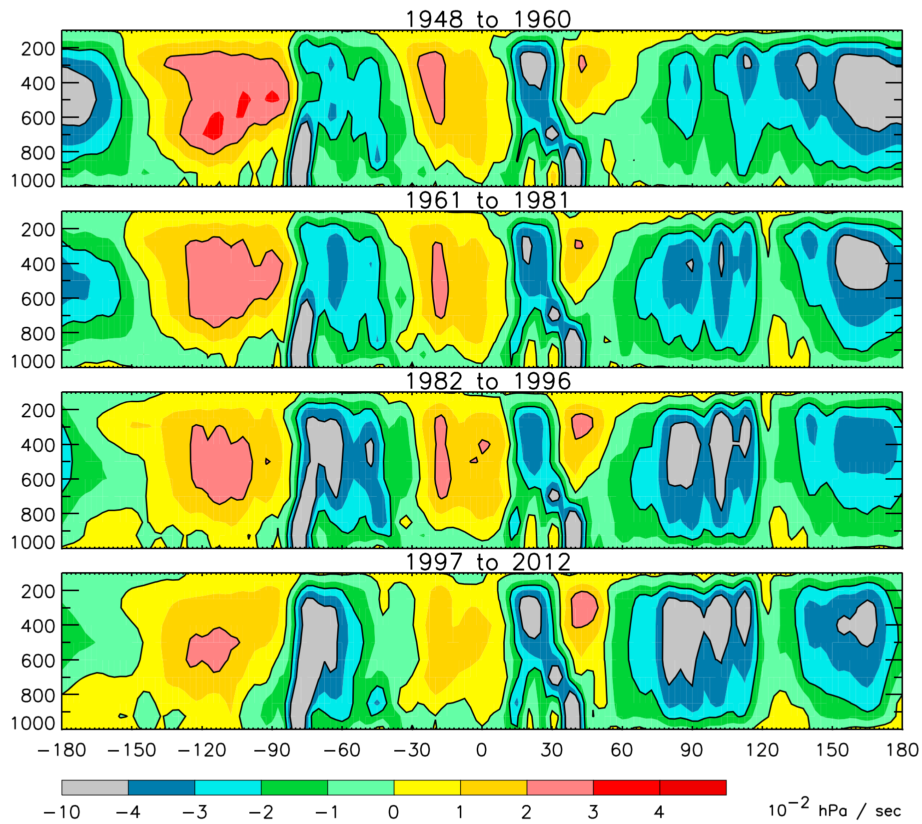

Figure 16. Vertical profile of omega (10–2 hPa s^{-1}) averaged between 10°N and 10°S over four time periods [from *Nicholson*, 2015a].

6.3. Large-Scale Factors Associated With the Trends

Funk et al. [2008] linked the decline of the long rains to warming in the south central Indian Ocean. They suggested that the impact of the warming on eastern Africa is manifested via a disruption of moisture transport. The disruption, in turn, is provoked by increased precipitation over the warm anomalies in the Indian Ocean. *Williams and Funk* [2011] noted that the warming was much greater in the Indian Ocean than in the Pacific and demonstrated that this resulted in a westward extension of the Indo-Pacific warm pool. They further suggested that the midtropospheric diabatic heating associated with the increased precipitation over the Indian Ocean was a major cause of the long rains decline. The midtropospheric heating would presumably send dry air aloft toward eastern Africa.

Lyon and Dewitt [2012], *Lyon et al.* [2014], and *Lyon* [2014] implicated tropical Pacific warming as a cause of the decline, especially post-1999. *Funk et al.* [2013] agreed that the relevant warming included the Pacific but emphasized that the relevant changes span the Indo-Pacific sector, and not the Pacific alone. However, the most direct mechanism by which these SST changes modulate East African rainfall appears to be a weakening of the Somali Jet and moisture transport into East Africa.

The Indo-Pacific SST changes appear to have been a result of radiative forcing [*Funk and Hoell*, 2015], suggesting an anthropogenic component to the warming via both greenhouse forcing and aerosols [*Liebmann et al.*, 2014]. *Rowell et al.* [2015] examined several hypotheses related to the long rains decline and concluded that anthropogenic effects have played a role. They concluded that aerosols are a likely candidate but that land use change is not. Rowell et al.'s conclusion was based on model results, which showed that aerosols affect SSTs in regions that influence MAM rainfall. *Tierney and Ummenhofer* [2015] similarly provided evidence of the role of an anthropogenic component. *Lyon* [2014] reached a contrary view, concluding from observations and modeling results that anthropogenic climate change is not a major factor in the decline. His conclusion was based primarily on the degree to which natural factors alone could prescribe rainfall variability. Overall, the question of anthropogenic effects is still very open, in part because various models produce diverse results.

Liebmann et al. [2014] added an additional factor to the suggested causes of the long rains decline, an enhanced east-west SST gradient in the western Pacific. This was primarily a result of warming in the western Pacific, especially near Indonesia. This would have enhanced the Walker-like circulation over the Indian Ocean, increasing subsidence over eastern Africa. The change in the gradient accelerated in the middle to late 1990s, around the time that *Lyon and Dewitt* [2012] identified as the onset of an abrupt rainfall decline. *Liebmann et al*. [2014] also found that the downward trend in the long rains is associated with enhanced upper level westerlies over the central Pacific and increased upper level easterlies over the western Pacific and Indian Oceans.

Nicholson [2015a] noted a steady intensification of the October–November Indian Ocean Walker cell between 1948 and 2012 (Figure 16). This was accompanied by increased upper level subsidence over eastern Africa. While this would presumably be conducive to drought, the ascending branch of the Atlantic Walker cell over eastern Africa also increased. The juxtaposition of the two cells over eastern Africa suggests that slight changes in the influence of the Indian versus Atlantic Oceans might account for the abrupt shifts between drought and flood during the short rains in recent years [*Liebmann et al*., 2014; *Nicholson*, 2014c, 2015a, 2016a].

6.4. Recent Droughts

The downward decline in the long rains culminated in a series of disastrous droughts commencing in 2005. Since that time both the frequency and duration of droughts increased in eastern Africa and drought was widespread in every year since 2008. Particularly severe droughts occurred in 2009, 2010, and 2011 in both the summer and equatorial rainfall regions [*Lyon and Dewitt*, 2012; *Viste et al*., 2013; *Nicholson*, 2014c, 2016a]. The 2010/2011 drought (Figure 17) was most intense in Kenya, southern Somalia, and southern Ethiopia. Rainfall was at least 50% to 75% below average in these regions. Compounding the already severe situation, high rainfall in November of 2009 and January of 2010 produced a flood situation, as did high rainfall in November of 2011.

There were some surprising aspects to these recent droughts. One is the persistence of drought over several rainy seasons [*Hoell and Funk*, 2014; *Nicholson*, 2014c, 2016a]. Another is the similarity between rainfall trends in the summer rainfall and equatorial rainfall regions. Based on monthly anomalies for 17 years (i.e., 204 correlation pairs), the correlation between the regions is 0.41 (1% significance level = 0.18). The similarity results from trends in all four seasons, but the strongest associations are in January to March (r = 0.58, based on 51 monthly correlation pairs, 1% significance level = 0.35). Correlations for the remaining seasons are April to June r = 0.36, July to September r = 0.53, and October to December r = 0.38. This relationship between the two regions is surprising, given the very different factors identified in interannual variability in them (see section 5).

Some of the similarity may be related to the tendency for rainfall anomalies to persist between July and September, the main rainy months in the summer rainfall region, and October to December [*Nicholson and Entekhabi*, 1986], the months responsible for the lion's share of interannual variability in the equatorial rainfall region. The correlation between the two seasons, based on 17 years/correlation pairs, is 0.79 for the summer rainfall region and 0.66 for the equatorial rainfall region. This suggests some degree of commonality in the factors influencing the two regions and two seasons. Possibilities are discussed in section 6.6.

6.5. Large-Scale Factors Associated With Flood and Drought Events

Most of the studies that have examined individual extreme events have looked at the short rains season. *Hastenrath* [2007] identified an extremely important factor, the equatorial westerlies over the central Indian Ocean. These are the surface manifestation of the Indian Ocean Walker cell. The westerlies drive the Wyrtki Jet in the upper ocean, enhancing the east-to-west temperature gradient along the equator by decreasing/increasing thermocline depth in its western/eastern portion. Increased westerlies then result in cooler waters in the west, near eastern Africa, as well as a stronger Walker cell, stronger easterlies at 200 mbar, and increased subsidence over eastern Africa. Enhanced westerlies were associated with short rains droughts of 2005 and 2010 [*Hastenrath*, 2007; *Hastenrath et al*., 2010; *Nicholson*, 2016a].

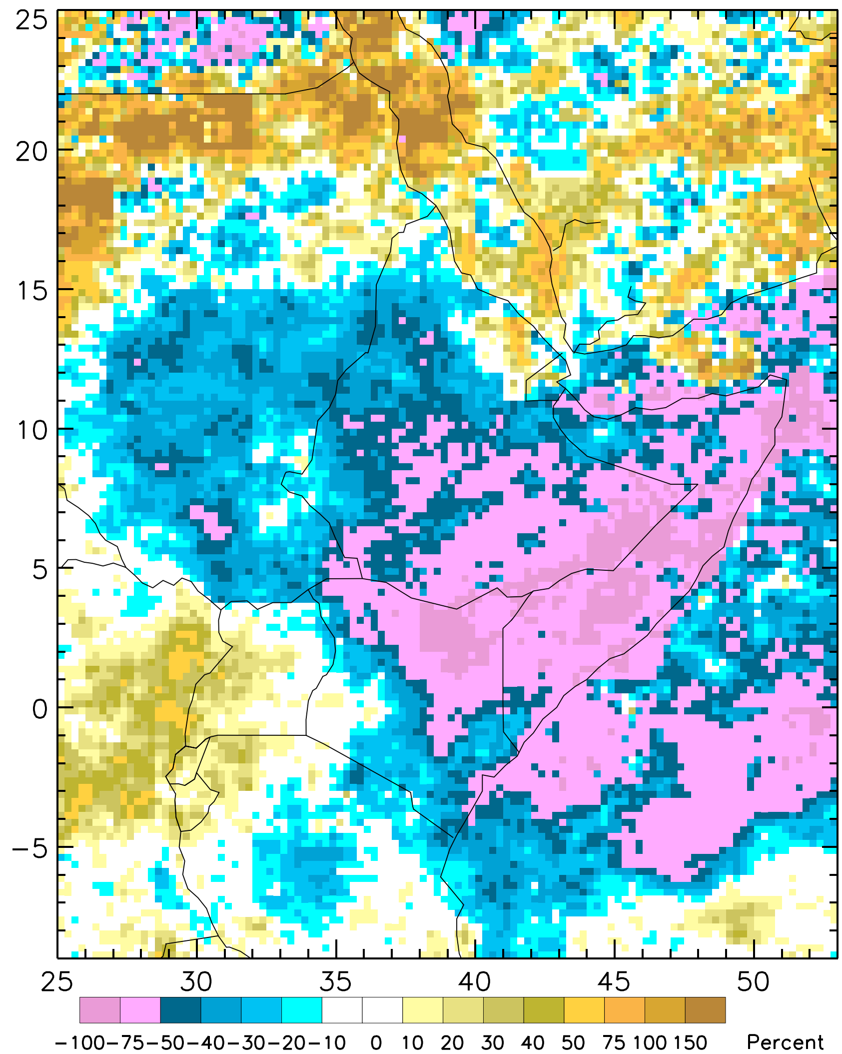

Figure 17. Map of rainfall during the period July 2010–June 2011, expressed as a percent above or below the annual mean [from *Nicholson*, 2014a].

Anomalously weak westerlies and a weak Indian Ocean Walker cell were associated with floods in 1961, 1997, and 2006 [*Kijazi and Reason*, 2005; *Hastenrath*, 2007; *Hastenrath et al*., 2010]. In some cases the temperature gradient actually reverses and the low-level equatorial flow becomes easterly. Such a reversal occurred in the 1997 flood event [*Birkett et al*., 1999]. The warming in the western Indian Ocean started in spring of 1997 and by October easterlies were established between the equator and 10°S and persisted until January 1998 [*Murtugudde and Busalacchi*, 1999; *Murtugudde et al*., 2000]. The easterlies generate cooling in the eastern Indian Ocean, amplifying the reversal of the gradient. *Reverdin et al*. [1986] suggested a similar positive feedback between ocean and atmosphere as a cause of the 1961 flood event.

The association between stronger/weaker westerlies in the boreal autumn and drought/wet conditions clearly explains the abrupt shift from flood to drought in November of 2010 and 2011 [*Nicholson*, 2016a]. In November 2010 the equatorial westerlies over the central Indian Ocean were the strongest of any month during the period 1998 to 2013 and the rainfall anomaly was one of the largest. In contrast, strong easterly anomalies prevailed during the flood event of November 2011. However, at least one additional factor probably played a role. A marked contrast occurred in the winds in the Turkana Channel, which were very strong in November 2010 and very weak in November 2011. These changes appear to amplify the effect of the large-scale factors governing rainfall over eastern Africa [*Nicholson*, 2016a].

Hastenrath [2007] concluded that the relationship between equatorial westerlies and East African rainfall was very robust and operated in the last two decades of the nineteenth century, when lake levels fell and glaciers receded. However, the equatorial westerlies over the central Indian Ocean appear to be a more consistent factor in years of anomalously strong short rains than in drought years. Examining October–November during the period 1874 to 2012, *Nicholson* [2015a] showed that the westerlies were exceedingly weak (at least 0.5 standard deviations below normal) during 17 of 19 "flood" years (ON rainfall at least 0.5 standard deviations above normal). In comparison, they were exceedingly strong in only 9 of 23 drought years (ON rainfall at least 0.5 standard deviations below normal). *Lyon* [2014] found that SSTs in the tropical Pacific and Indian Oceans are a major factor in drought during the short rains but that their links to drought during the long rains are much weaker.

Nicholson [2015a] further showed an interesting contrast between drought and wet years. For both cases four potential causal factors were examined: surface and 200 mbar zonal winds over the central equatorial Indian Ocean, SSTs in Niño 3.4, and the Indian Ocean Zonal Mode. The wet years were generally associated with three or four causal factors occurring simultaneously (16 out of 19 cases), while the dry years were generally associated with only one or two (14 out of 23 cases). In several drought years, none of these factors was operative. This contrast is at least partially explained by a nonlinear response to changes in the causal factors. Each of the factors showed a strong linear relationship to rainfall magnitude during anomalously wet years. Dry years, on the other hand, were associated with a specific sign of the individual factors, but virtually no correlation existed with rainfall magnitude during those years. *Nicholson* [2015a] also noted a change in the factors associated with drought from 1982 onward. In several cases drought did not develop despite the occurrence of factors generally conducive to drought.

Smoleroff [2015] examined those cases and concluded that despite the occurrence of factors that generally produce drought, reduced subsidence over East Africa made drought less likely. She also contrasted conditions associated with both dry years and anomalously wet years before and after 1982. There was marked warming of the tropical oceans post-1982, but the more critical factor appears to have been changes in the zonal circulation cells. That over the Pacific was weaker, and those over the Indian Ocean and over the eastern Atlantic-Africa were intensified post-1982.

Fewer studies have looked at extreme events during the long rains. During this season floods are rare but drought is common. *Shanko and Camberlin* [1998] noted that years with a large number of tropical depressions over the southwestern Indian Ocean tend to be drought years in Ethiopia. An abnormally low number of depressions is associated with heavy rainfall. This relationship is strong for the February to May season but plays some role in June to September as well. The suggested mechanism involves SSTs in the southeastern Indian Ocean, where the depressions originate. The authors also noted anomalously strong equatorial easterlies in the upper troposphere and a reduced southward excursion of the subtropical jet stream during years of weak convective activity over Ethiopia.

Examining drought in Tanzania during the period 1998 to 2005, *Kijazi and Reason* [2005] found strong moisture flux divergence and increased subsidence during the MAM season. This was associated with the descending branch of a Walker-like circulation over the region. *Lyon and Dewitt* [2012] examined the failure of the MAM rains during 2011 and concluded that the drought was related mainly to an abrupt increase in SSTs in the tropical Pacific Basin. *Williams and Funk* [2011] suggested that it was related mainly to the much greater warming in the Indian Ocean than in the Pacific and to an enhanced temperature gradient between the Pacific and Indian Oceans [see also *Hoell and Funk*, 2013]. The modeling study of *Lott et al*. [2013] found that the 2011 drought in the long rains was associated with roughly equal warming in the two oceans.

Pohl and Camberlin [2006a] examined the long rains in the context of intraseasonal and interannual variability, rather than individual drought events or flood years. However, their results are relevant here. They found a strong influence of the Madden-Julian Oscillation (MJO) on the long rains, with dry conditions associated with weak MJO activity. In these cases, westerly wind anomalies are evident over coastal Kenya. This is consistent with the finding of a stronger Walker-like circulation over the Indian Ocean in dry years. *Pohl and Camberlin* [2011] found that extreme events in the long rains and the short rains are associated with cases where the phase of the MJO that enhances convective activity over eastern Africa occurs during an El Niño year.

ENSO plays a major role in droughts and floods in the summer rainfall region of Ethiopia. El Niño suppresses rainfall over Ethiopia in the boreal summer, but La Niña enhances it [*Pohl and Camberlin*, 2011]. Rainfall in this season is strongly positively correlated with SSTs in the equatorial western Pacific but negatively correlated with SSTs in the eastern Pacific [*Segele et al*., 2009a]. Abundant rainfall is also associated with enhanced westerlies across western and central Africa, a strong Somali Jet, and a strong Tropical Easterly Jet [*Segele et al*., 2009b]. SSTs in the southern tropical Indian Ocean also have an influence [*Williams et al*., 2012].

6.6. Commonalities in Factors Influencing the Two Regions and Three Rainy Seasons

In view of the very different factors modulating rainfall variability in the summer and equatorial rainfall regions and in the three rainy seasons, it is difficult to explain the recent tendency for consecutive drought seasons and for similar trends in the summer and equatorial rainfall regions. As for the similar trends in the two regions, the greatest similarity is evident in the January to March and in the July to September seasons. Because January to March is relatively dry in both regions, the focus here is on common factors during July to September.

Some factors can be readily ruled out. Rainfall in the two regions has an opposite response to Pacific SSTs, with anomalously warm western Pacific SSTs increasing/reducing July to September rainfall in the summer/equatorial rainfall region [*Segele et al*., 2009a; *Funk et al*., 2013]. Another potential factor, ENSO, influences the summer rainfall region in the boreal summer, but not the equatorial rainfall region [*Pohl and Camberlin*, 2011].

The most promising mechanism may involve components of the Indian monsoon system. In the summer rainfall region, abundant June to August rainfall is associated with low surface pressure near Bombay, a strong Somali Jet, strong monsoon rains, westerly/easterly wind anomalies in the lower/upper troposphere, and enhanced westerlies across western and central Africa [*Camberlin*, 1995, 1997; *Segele et al*., 2009b; *Omondi et al*., 2012]. The westerly flow brings moist Congo air into the summer rainfall region. This westerly flow is part of a circulation cell over the western Indian Ocean that includes southeasterlies south of the equator. These would presumably be increased as well, bringing moist air into the equatorial rainfall region and enhancing low-level convergence over this region as well.

While this mechanism might account for July to September anomalies of the same sign in both regions, it cannot account for persistence of the anomalies into the boreal autumn season because the seasonal circulation patterns are so different. Another possibility may be the well-documented changes in SSTs in the Indian and Pacific Oceans. *Williams and Funk* [2011] have shown that the stronger warming in the Indian Ocean is a major factor in boreal spring drought in eastern Africa [see also *Williams et al*., 2012]. The observed changes in Indo-Pacific SSTs are also a factor in the more frequent occurrence of drought in eastern Africa during the boreal summer season and are a probable cause of the occurrence of drought over consecutive long and short rains seasons [*Funk et al*., 2013; *Hoell and Funk*, 2014]. The mechanism of changes in precipitation over eastern Africa is shifts in the Walker circulation over the Indian Ocean.

7. Seasonal Prediction

Attempts to predict seasonal rainfall in eastern Africa have spanned nearly three decades. It is impossible to closely compare the various studies because of differences in the time span upon which predictive models were developed or validated and differences in the target area of the predictions. However, some very conclusive results have emerged. One is that the short rains are relatively predictable, while the long rains are not. This conclusion applies both to statistical and dynamical forecasts [e.g., *Dutra et al*., 2013; *Mwangi et al*., 2014]. A second is the consensus that statistical approaches generally possess higher forecast skill than dynamical predictions based on numerical models [*Chen and Georgakakos*, 2015]. The latter tend to poorly predict extreme events and underpredict drought in eastern Africa [*Korecha and Sorteberg*, 2013; *Jury*, 2014]. However, dynamic models show significant forecast skill for the short rains, since they accurately project the Indian Ocean Zonal Mode [*Bahaga et al*., 2016], which is a major driver of the short rains.

Some relatively new approaches have considerable potential. For one, *Philippon et al*. [2002] pointed out that predictive atmospheric signals have been neglected in most statistical models, which tend to rely on surface forcing, generally emphasizing sea surface temperatures. That study and others [e.g., *Nicholson*, 2014a, 2015b] have shown that consideration of atmospheric variables such as zonal winds or low-level circulation can improve forecast skill. *Chen and Georgakakos* [2015] presented a statistical approach based on SST dipoles and found that it provides more accurate seasonal forecasts than canonical correlation or linear regression. Coupled ensemble models or hybrid approaches that combine numerical model results with statistical downscaling to create regional forecasts also have considerable promise [e.g., *Jury*, 2014; *Shukla et al*., 2014a, 2014b; *Segele et al*., 2015]. An alternative approach is to predict hydrologic drought, which forecasts soil moisture and has the benefit of this being a temporally integrative variable [e.g., *Shukla et al*., 2014b; *AghaKouchak*, 2015].

7.1. The Long Rains of the Boreal Spring

The most extensive study of the predictability of the boreal spring rains is that of *Camberlin and Philippon* [2002]. They produced a regression model for a region of Kenya and Uganda that corresponds closely to our equatorial rainfall region. In contrast to most previous studies, they examined atmospheric parameters in addition to SSTs as potential predictors. Four February indices were selected as input to multiple regression and linear discriminant analysis models. These included SSTs in Niño 1.2, zonal wind over the Congo Basin at 1000 mbar, geopotential height of the 500 mbar surface over the Near East, and the east-west moist static energy gradient between the East African highlands and the Sahel. The models were applied for the period 1968 to 1997 and were evaluated using cross validation. For the linear multiple regression model the correlation between the predicted and observed March to May rainfall for the Kenya/Uganda section was 0.66 in the cross-validation mode. The discriminant analysis model correctly classified the seasonal anomalies 70% of the time.

Eden et al. [2014], *Diro* [2008], and *Ntale et al*. [2003] used empirical methods to predict rainfall in the boreal spring. Eden et al. produced a forecast only for Addis Ababa, but the other two focused on Ethiopia as a whole. *Ntale et al*. [2003] applied canonical correlation analysis to predict standardized MAM rainfall totals at a 3 month lead time. They found the optimal predictors to be sea level pressure and SST anomaly fields of the Indian Ocean adjacent to East Africa and in the Gulf of Guinea in the equatorial Atlantic. *Camberlin and Philippon* [2002] similarly found strong local influence (the Red and Arabian Seas) on MAM rainfall in Ethiopia.

Because of the spatial variation in interannual variability and in the annual cycle within Ethiopia, *Diro* [2008] produced separate forecast models for five homogeneous zones. Both multiple linear regression and linear discriminant analysis were applied to four sets of predictors. The study found that the method of selecting predictors had little impact on forecast skill. It was also shown that the models had the most skill in the southern and eastern parts of Ethiopia and that the extreme years were more reliably forecast than the average years.

Moron et al. [2013] used a unique approach to predicting boreal spring rainfall in eastern Africa. They noted that seasonal prediction typically focuses on 3 month rainfall totals at a regional scale. This can be problematic when the season is not homogeneous. The authors demonstrated that both spatial coherence of rainfall and potential predictability from SSTs is greater during the weak rains of March than during the peak season months of April and May, when spatial coherence is particularly weak. By combining the EOFs of both interannual and subseasonal variations with a fuzzy *k*-means clustering, they were able to filter out the noisier variations of the rainfall field and emphasize the most consistent signals in both time and space. This approach provided insight into the seasonal predictability of long dry spells and heavy daily rainfall events at local scale and into their subseasonal modulation.

Funk et al. [2014] developed a statistical forecast model for the long rains (Figure 18). It is based on two SST indices, the western Pacific SST gradient and SSTs in the central Indian Ocean, and these were respectively utilized to predict the first and second PCs of the interannual variability of the long rains. With a 2 month lead time, the correlation between observed and predicted in the cross-validation mode was ~0.6, based on the time period 1981 to 2013.

Nicholson [2014a, 2015b] used multiple linear regression to produce a seasonal forecast model for the boreal spring season in both the summer rainfall region and the equatorial region. In contrast to most other empirical forecast models, atmospheric variables were emphasized. The variables considered included sea surface temperature, sea level pressure, omega, and zonal and meridional winds at 925 mbar, 850 mbar, 700 mbar, and 200 mbar throughout the global tropics. The forecast results were determined for the period 1950 to

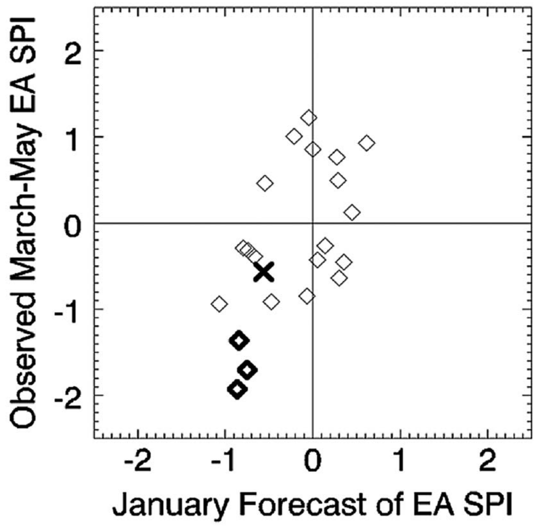

Figure 18. Scatterplot of observed and forecast MAM rainfall from January SST anomalies [from *Funk et al*., 2014]. Cross indicates a forecast based on February SST. Diamonds indicate extreme drought years.

2005. Two notable findings were that atmospheric variables, especially zonal winds, generally provide higher forecast skill than surface variables, such as sea surface temperatures and sea level pressure, and that ENSO and the Indian Ocean Dipole provide less forecast skill than variables associated with them.

The regression models using February input data produced a correlation between predicted and observed rainfall of 0.76 for the equatorial region and 0.63 for the summer rainfall region. In the cross-validation mode, the correlations fell to 0.63 and 0.49, respectively. The model for the equatorial rainfall region based on January input data gave a correlation of 0.74 between predicted and observed rainfall, roughly equal to the correlation based on February input data. In all three cases the root-mean-square error (RMSE) ranged between 0.30 and 0.32.

In view of the weaker skill for the MAM season, compared to JJAS and ON seasons, and the heterogeneity of the long rains [*Camberlin and Philippon*, 2002; *Moron et al*., 2013], *Nicholson* [2015b] developed seasonal forecast models for the three individual months. This approach resulted in markedly improved skill for the equatorial region but somewhat lower skill for the summer rainfall region. As an example, for the equatorial rainfall region the correlation between predicted and observed over the period 1950 to 2005 at 2 month lead time was 0.63 for the MAM season, but 0.78, 0.83, and 0.84, respectively, for the individual months of March, April, and May. Notably, the predictors were almost exclusively atmospheric variables; SSTs and sea level pressure provided little forecast skill. Also, skill was limited to a 2 month lead time, suggestive of the spring predictability barrier of ENSO [e.g., *Torrence and Webster*, 1999; *Wajsowicz*, 2007].

7.2. Summer Rainy Season

The JJAS season accounts for some 50% to 80% of the rainfall over Ethiopia's agricultural regions and for a significant amount of rainfall in the highland and coastal regions of Kenya. Eritrea, Djibouti, northern Ethiopia, northern Uganda, and South Sudan also receive most of their rainfall in the boreal summer. However, with the exception of *Nicholson* [2014a], forecast models for this season have been produced only for Ethiopia [e.g., *Jury*, 2015] or for a region within that country, such as the Blue Nile basin [*Elsanabary and Gan*, 2014].

Using mainly multivariate statistical techniques, *Korecha and Barnston* [2007] obtained significant forecast skill for the JJAS season. They concluded that ENSO is the most significant predictor but that more local factors near Africa and in the Atlantic and Indian Ocean are also important. However, forecast skill was apparent only with relatively short lead times, which the authors attributed to the spring predictability barrier of ENSO. Their regression model used three variables to predict JJAS rainfall: February–March SSTs in the South Atlantic and in the Niño 3.4 region plus May SST in the Niño 3.4 region. The model explained 59% of the overall rainfall variance and 41% of the variance in the cross-validation mode. *Gissila et al*. [2004] similarly utilized statistical forecasting for this season, with SSTs in the Pacific and Indian Oceans as predictors. They improved predictive skill by separately considering several regional sectors of Ethiopia. The degree of skill was found to be regionally dependent.

Diro et al. [2011a, 2011b] also used surface parameters to predict Ethiopian summer rains. In agreement with *Gissila et al*. [2004], they found that the best predictors vary by region, indicating that over Ethiopia statistical forecast models need to take into account local variations in teleconnections. They found forecast skill from SSTs in the equatorial Pacific, in midlatitude regions of the northwest Pacific, and in the Gulf of Guinea. An important part of their study was to provide a justification for the use of predictor regions (such as the Pacific Northwest) for which no clear physical relationship to Ethiopia is apparent. Such predictors can be precursors to the development in later seasons of predictors with a direct physical connection.

Block and Rajagopalan [2007], like the aforementioned studies, used surface variables as predictors in their model for rainfall in the Upper Blue Nile Basin (northwestern Ethiopia). They utilized a nonparametric approach based on local polynomial regression. This approach precluded selecting predictors on the basis of arbitrary correlations. Their models utilized variables of the MAM season to predict JJAS rainfall.

In contrast to many prior studies, *Segele et al*. [2009a, 2009b] found that the strongest links to Ethiopian boreal summer rainfall are not SSTs or ENSO but are various components of the African-Asian monsoon system, such as the Azores, Saharan, and Mascarene highs, the Tropical Easterly Jet, and the monsoon trough. Based on these results, *Segele et al*. [2015] developed forecasts for the summer rains by applying multiple linear regression to ensemble-based prediction of potential predictive signals in regional circulation and global SSTs. Predictions for the JJAS season produced 2 to 3 months in advance achieved excellent results. The correlation between observed and predicted all-Ethiopia rainfall was +0.81 over the period 1970 to 2002 and only somewhat lower for rainfall at individual stations.

Nicholson [2014a] developed multiple linear regression forecast models for summer rainfall for both the summer and equatorial rainfall regions of eastern Africa, thus considering areas well beyond Ethiopia. In contrast to the forecast models developed for other seasons, the majority of predictors were sea surface temperature and pressure. As with *Gissila et al*. [2004], the predictors included SSTs in the midlatitude portion of the equatorial Pacific. Based on May predictors, the model predictions for July to September rainfall correlated strongly with observed rainfall in the summer rainfall region but less skill was evident for the equatorial rainfall region. The models respectively explained 63% and 29% of the seasonal rainfall variance over the period 1950 to 2005. In the cross-validation mode, the model still explained 56% of the variance for the summer rainfall region but could account for only 13% of the variance in the equatorial rainfall region. In agreement with *Korecha and Barnston* [2007], this study's results showed that the spring predictability barrier limits the lead time for the forecasting of summer rainfall over eastern Africa.

Diro et al. [2012] further evaluated the potential of dynamical prediction of Ethiopian rainfall by using a regional model (RegCM3) to downscale European Centre for Medium-Range Weather Forecasts (ECMWF) seasonal ensemble forecasts. They found that the skill of probabilistic forecasts was greater with ECMWF on a grid-point-by-grid-point comparison than for their six homogeneous zones. However, RegCM3 had higher skill at the countrywide scale.

7.3. The Short Rains of the Boreal Autumn

The short rains, although the secondary season in most of eastern equatorial Africa, provide the largest contribution to interannual variability. They also appear to be the most predictable. The potential for seasonal prediction has long been recognized [*Farmer*, 1988; *Kinuthia et al*., 1988; *Hutchinson*, 1992; *Ogallo et al*., 1994]. Statistical forecast models for this season have been developed by *Philippon et al*. [2002], *Mutai et al*. [1998], *Mutai and Ward* [2000], *Ntale et al*. [2003], *Hastenrath et al*. [2004], *Mwale and Gan*, [2005], and *Nicholson* [2014a]. The few studies that utilized numerical forecast models [*Batté and Déqué*, 2011; *Dutra et al*., 2013; *Mwangi et al*., 2014] produced results that were less accurate and with shorter lead times than the best statistical models.

Traditionally, the months of the short rains are considered to be October and November. Some authors, however, consider a longer season, such as September to December or October to December. The coherence of rainfall anomalies during this 2 month season is well known. *Camberlin and Philippon* [2002] and others have shown found that coherence is limited to these 2 months, i.e., the longer seasons are not homogeneous. The exception appears to be the coastal region, where the correlation between the ON season and the September to December season is 0.97 [*Hastenrath et al*., 2004]. However, even in this region there is much greater skill in predicting October and November rainfall than December rainfall [*Hastenrath et al*., 1993].

Mutai et al. [1998] found that the July to September global SST pattern is strongly correlated with October to December seasonal rainfall aggregated for a large sector of eastern Africa extending from

5°N, in Kenya, southward to Malawi at 15°S. They developed a multiple linear regression forecast model based on three rotated EOFs for SSTs in the northwest Pacific, the eastern equatorial Pacific (the ENSO signal), and the South Atlantic. The model showed significant forecast skill, with a correlation between predicted and observed of 0.69 for the period 1945 to 1988 for rainfall averaged over the entire region. The strongest predictor was an SST EOF with maximum variance in the northwest Pacific. In contrast, *Bahaga et al*. [2015], who examined predictability over an entire century, found little forecast skill from ENSO. Instead, the Indian Ocean Dipole, especially its western pole, provided the greatest predictive skill.

Kabanda and Jury [1999] examined the predictability of October to December rainfall in northern Tanzania. They developed a multiple regression model based on three tropical predictors of the July to September season: zonal and meridional wind in the Indian Ocean and the Southern Oscillation Index. The model accounted for about half of the rainfall variance during the 1970s and 1980s.

Philippon et al. [2002] used a multiple linear regression model to predict the OND rainfall in a large sector that included inland Kenya, northern Tanzania, and most of Rwanda, Burundi, and Uganda. Based on September predictors identified from correlations for the 1968 to 1997 period, their model explained 64% of the interannual variance. The predictors included a monsoon index involving the NE and SW wind components at 200 mbar and 850 mbar, respectively, meridional wind at 200 mbar over the southeastern tip of Africa, and an index of circulation over the western Indian Ocean.

Ntale et al. [2003] used canonical correlation analysis to predict standardized seasonal rainfall totals for September to November at 3 month lead time. Predictors included SLP and SST anomaly fields in the Indian and Atlantic Oceans. The strongest association was with SSTs off the Somali and Benguela coasts. *Mwale and Gan* [2005] continued the work, comparing several methods of predicting standardized seasonal precipitation at 21 stations within a homogeneous region that comprises most of East Africa. Skill was higher with a nonlinear model known as artificial neural network than with the more standard linear canonical correlation model. In the latter case, the percent variance explained at individual stations for the 11 seasons 1987 through 1997 ranged from 42% to 66%, with RMSE of 0.4 to 0.75 standardized units. The linear correlation model explained 6% to 56%, with RMSE of 0.4 to 1.2.

Hastenrath et al. [2004] conducted several prediction experiments for the ON season using a linear forecast model and a variable number of predictors, including two experiments with the Southern Oscillation Index as the only predictor. The best predictors were zonal temperature and pressure gradients across the equatorial Indian Ocean, variables suggestive of the IOZM. A cross validation for 1958–1996 based on six predictors produced a correlation between predicted and observed rainfall of 0.45. However, when the model was tested using separate training and validation periods, correlation in the latter was much lower. It also appeared that the correlation with individual predictors changed markedly over time. *Farmer* [1988] reached a similar conclusion, finding that the Southern Oscillation Index provided significant forecast skill for Kenyan rainfall during the period 1943–1984, but not for the period 1901–1942.

Nicholson's [2014a] statistical forecast model also considered the ON season. For the equatorial rainfall region the correlation between predicted and observed rainfall during the years 1950 to 2005 was 0.78 using August input data and 0.80 using May input data (Figure 19). For the summer rainfall region the model performed nearly as well, with the correlation between predicted and observed being 0.77 using August input data and 0.72 using May input data. The contrast between the models for the two seasons was that the majority of August predictors were winds aloft or omega, while roughly half of the May predictors were surface variables such as SSTs. This is consistent with the longer persistence of SSTs, compared with atmospheric characteristics.

Collectively, these studies suggest that seasonal predictability of the short rains is strong, with circulation variables generally providing the greatest skill, especially at short lead times. So far, statistical forecasts appear to outperform dynamical forecasts, but there is considerable disagreement concerning the relative importance of the Pacific, Atlantic, and Indian Oceans in these forecasts. Moreover, the relationship between predictors and predictands may be nonstationary. Because the relationship between forcing factors and rainfall is nonlinear, models such as neural networks that do not assume linearity might outperform models based on linear regression. Forecast skill appears to be greater for October–November than for September to November or October to December, as the longer periods are climatologically less

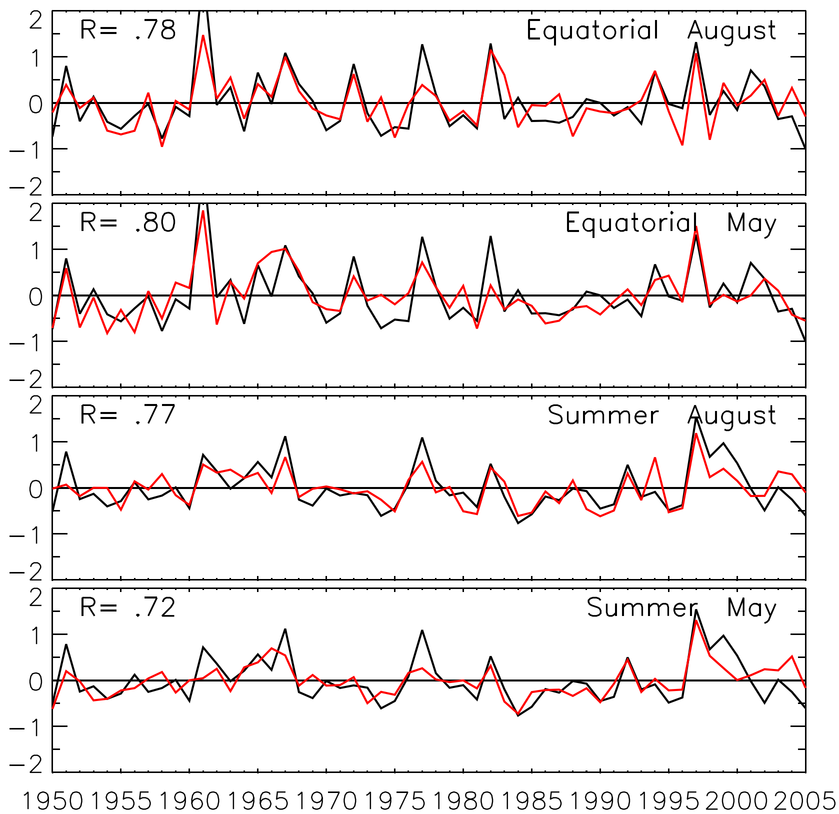

Figure 19. Predicted (red) versus observed (black) ON rainfall for the (first and second panels) equatorial and (third and fourth panels) summer rainfall regions with August predictors in one case (Figure 19, first and third panels) and May predictors in a second case (Figure 19, second and fourth panels) [from *Nicholson*, 2015b]. Correlation *r* between the predicted and observed is indicated in each panel.

coherent. Overall, there is a need to better understand the physical processes providing the link between predictors and predictands.

8. Summary

8.1. Regionalization

The equatorial rainfall region of eastern Africa is strongly homogeneous with respect to interannual variability, particularly during the October–November short rains season. Some exceptions to this include the western highlands of Kenya and the coastal strip. In contrast, the summer rainfall region, Ethiopia, in particular, is spatially heterogeneous with respect to interannual variability and in most research subregions should be considered.

8.2. Seasonal Cycle

The paradigm for the equatorial rainfall region of two rainy seasons occurring as a result of the biannual equatorial passage of the ITCZ cannot be substantiated. Some recent studies have made progress in understanding the seasonal cycle in this region, but further work is required.

Rainfall variability and characteristics within the October–November season are strongly coherent. This coherence does not extend to the longer seasonal groups of September to November or October to December. This strongly argues for defining the short rains as October–November in analysis and prediction.

The long rains of the boreal spring should not be treated as a single season because the character, causal factors, and teleconnections are markedly different in each month. May, in particular, shares many characteristics with the JJAS season. Thus, each month of the so-called long rains season should be treated separately in analysis and prediction. Past consideration of the season as a single entity may help to explain the lower success in seasonal prediction, compared to the short rains, and paucity of atmospheric or oceanic teleconnections.

8.3. Intraseasonal Variability

Two very important factors in intraseasonal variability are low-level zonal winds over and near the continent and the MJO. Intraseasonal variability is much less spatially coherent than interannual. For example, there is an out-of-phase relationship between the East African coast and the highlands.

8.4. Interannual Variability

Four major large-scale factors that modulate the short rains have been identified: ENSO, the IOZM, and zonal winds near the surface and at 200 mbar over the central equatorial Indian Ocean. Their relationship to rainfall is nonlinear. They are positively correlated with the amount of rainfall in anomalously wet years but merely predispose the region to the occurrence of drought. Several factors act together to produce an extremely wet year but generally only one or two are evident during drought years. The causes of numerous droughts in the short rains still need to be identified.

Although the short rains show a close link to both ENSO and the IOZM, the strongest links are to the zonal circulation over the Indian Ocean. These factors are also closely interrelated. However, the magnitude of the link between each factor and ON rainfall is nonstationary, and the degree of dominance of individual factors changes on a decadal time scale. The coupling between the short rains and ENSO has markedly diminished since 1982.

It is well established that the Walker cell over the Indian Ocean is an extremely important factor in the interannual variability of the short rains. However, shifts in other zonal circulation cells appear to play a role as well. Most notably is that over the eastern Atlantic and western equatorial Africa. The close juxtaposition of that and the Indian Ocean cell makes eastern Africa very sensitive to changes in the zonal circulation.

Several more local factors affect both regions and several seasons. Recent research has underscored the importance of the low-level westerlies over both land and ocean, of the Tropical Easterly Jet, and of water vapor transport from the Congo. The role of the latter factor makes the well-known decline of the rain forest in the Congo Basin a cause for concern about eastern Africa's rainfall regime.

8.5. Recent Trends

Over much of eastern Africa the long rains have been declining in recent decades, while the short rains have increased. Factors governing the variability of the long rains have finally been identified, with the MJO having a strong influence on year-to-year variations and with the relative strength of Pacific and Indian Ocean anomalies playing a major role in the downward trend.

Droughts have become longer and more intense and tend to continue across rainy seasons. At the same time, interannual variability has increased so that unusually strong floods have also plagued the region.

8.6. Seasonal Prediction

Considerable progress has been made in the seasonal forecasting of rainfall. The short rains season is markedly more predictable than the long rains and with longer lead time. Prediction of the long rains appears to be impeded by the spring predictability barrier, which precludes long lead times in the forecasts. Atmospheric variables provide more reliable seasonal forecasts than the factors traditionally considered, such as SSTs and ENSO. In contrast, SSTs and sea level pressure were strong predictors of summer rainfall. However, lead times are fairly short for accurate prediction because of the spring predictability barrier.

9. Conclusions and Priorities for Further Research

Three controversial points emerge from this review: the relative importance of the three major oceans in driving variability and providing forecast skill, whether statistical or dynamical forecast models provide greater skill in this region, and the probable impact of global warming on the rainfall regime in eastern Africa. The first two controversies are interrelated and probably relate to the nonstationarity of the relationships and to the lack of controlled comparisons between various approaches to prediction. The cited studies differ with respect to years, regions, and seasons evaluated. The third controversy cannot be resolved without the development of better climate and regional models. The latter are particularly important because they can better resolve the complex geographical controls on the region's precipitation regime.

As for the role of the oceans, clearly the major control shifts over time. During the short rains the most direct and most robust influence is that of the low-level flow in the equatorial Indian Ocean, but even its link to rainfall variability wanes from time to time. While both ENSO and the IOZM modulate this flow, it also exhibits interannual variability that is independent of these global modes. The degree of upper level subsidence over eastern Africa also appears to be a critical factor in rainfall variability. More research is needed to identify large-scale factors that influence the low-level equatorial Indian Ocean and the upper level subsidence. The potential role of midlatitude influences, such as the Southern Annual Mode, has, so far, received little attention.

A first step to resolving the controversy concerning statistical versus dynamical prediction is an intercomparison of the various predictive schemes based on a common precipitation database and common time periods and geographical locations for predictions and validation. Seasonal prediction via statistical models can also be improved via research to determine the robustness of proposed predictors, the physical links between rainfall and predictors, and the appropriate seasonal aggregation.

The ITCZ paradigm cannot adequately explain rainfall seasonality in eastern Africa. Although two papers have provided new insights into this question, further research is still needed. This is particularly true for more northern sectors of the region. A related and disputed question is whether or not a Walker-type circulation prevails during the long rains. While there exists only a weak semblance of such a circulation in the long-term mean, the possibility of a coherent zonal cell during some years has not been ruled out. This should also be further investigated.

Another topic that merits further research is the factors contributing to drought and its magnitude in both the equatorial and summer rainfall regions during the short rains. Also, the mechanism underlying the correlation between the summer and short rains has not been identified nor has the reason for the recent multiseason persistence of droughts. One potential factor that spans several seasons is the Turkana Jet. Further research is required to understand its interannual variability and how this is linked to the interannual variability of rainfall.

Related to this question is the contribution of the West African and Indian monsoon to summer rainfall in eastern Africa and to the early and late months of the long and short rains. Little research has been conducted on these topics. Similarity, further research is needed on the degree to which the seasons themselves are homogeneous with respect to characteristics of variability and causal factors.

A critical open question is the reason for the reduced coupling of the short rains to large-scale factors, especially ENSO, since 1982. Notably, a similar regime change occurred over West Africa in 1982 in July to September season. This may be a global-scale issue. For example, *Miller et al*. [1994] identified a major climate shift in the late 1970s and a major change in the global Walker circulation also occurred around 1982 [*Nicholson*, 2015a]. Related to this is what drives the decadal-scale changes in the dominant factors in rainfall variability over eastern Africa. Additional study of these issues may help in the quest to accurately project the impact of global climate change on this region.

Glossary of Technical Terms

Azores High: A subtropical high-pressure system in the North Atlantic, also known at the North Atlantic High or Bermuda-Azores High.

Bimodal rainy season: A season cycle of rainfall in which two maxima occur.

Cross validation: A method to validate prediction models by systematically eliminating one of the cases upon which the model was calibrated.

Cyclonic flow: Air flow that is counterclockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere, generally around a low-pressure center.

Dynamical forecasting: Forecasting by way of numerical models that are based on primitive equations linking atmospheric variables.

ECMWF: European Centre for Medium-Range Weather Forecasting

El Niño: A warming in the equatorial Pacific that has quasi-global consequences

Empirical orthogonal functions (EOFs): These represent the decomposition of a data set into a group of orthogonal (i.e., independent) functions. In climate studies each EOF most often represents a spatial mode of variability.

ENSO: Acronym for El Niño–Southern Oscillation (see these terms in the glossary).

Geopotential height: This approximates the height of a pressure surface above mean sea level. It includes an adjustment for the impact of gravity on potential energy of an air parcel.

Indian Ocean Zonal Mode (IOZM, also termed Indian Ocean Dipole or IOD): A contrast in sea surface temperatures in eastern and western portions of the equatorial Indian Ocean.

Interannual variability: Variability that occurs from year to year, generally applied to variability that is on time scales less than one decade.

Intertropical Convergence Zone (ITCZ): An area in which the trade winds of the two hemispheres convergence and often associated with low pressure, cloudiness, and precipitation

La Niña: Cooling in the equatorial Pacific that has quasi-global consequences

Madden-Julian Oscillation (MJO): A region of tropical convection that moves eastward near the equator and recurs roughly every 30 to 60 days. Depending on the phase of its local passage, rainfall can be enhanced or suppressed.

Mascarene High: A subtropical high-pressure system in the South Indian Ocean.

Principal component (PC) analysis: A statistical method that transforms a set of observations of several variables into a set of linearly uncorrelated variables termed principal components. Depending on the analysis method, the PCs can represent a spatial mode or a time series. The terms PC and EOF are often used interchangeably.

Quasi-biennial oscillation (QBO): Fluctuations in atmospheric circulation that occur on time scales of roughly two series. The QBO in the tropics has a time scale of about 2.3 years.

Regional climate model: A high-resolution climate model covering a limited area, driven by boundary conditions based on observations or from a lower resolution general circulation model.

Root-mean-square error (RMSE): An estimate of error between observed and predicted values of a time series, calculated as a standard deviation of the residuals (i.e., prediction errors).

Saharan High: A midtropospheric high-pressure cell that is an eastward extension of the Azores High.

Southern Oscillation Index (SOI): An index of a pressure oscillation across the equatorial Indo-Pacific region, most commonly based on the pressure difference between Tahiti and Darwin, Australia.

Spectral analysis: A statistical method that disaggregates a time series into the main time scales contributing to the bulk behavior.

Spring predictability barrier: The rapidly evolving state of the tropical Pacific in spring makes prediction of ENSO from variables occurring prior to spring very difficult.

Statistical forecasting: Forecasting by way of empirical or statistical techniques using large-scale variables to predict local conditions.

Tropical Easterly Jet (TEJ): A jet stream in the upper troposphere that extends from India to West Africa.

Walker circulation: A zonal circulation cell (see below) in the Pacific, with rising motion in the western portion and sinking motion in the eastern portion. Sometimes, the term is applied to a global series of such equatorial zonal cells.

Wyrtki Jet: Strong zonal currents that occur during boreal spring and fall in the tropical Indian Ocean.

Zonal circulation cell: An atmospheric circulation cell oriented in the vertical-equatorial plane.

Acknowledgments
The author was supported by NSF grants AGS 1160750 and AGS 1158984. The contribution of Douglas Klotter in preparing figures and painstakingly creating the reference list is greatly acknowledged. The author would like to thank the reviewers for their thorough reading of the article and their extremely useful comments and suggestions. The precipitation data utilized in this article are available from the author.

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