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EAST AFRICAN HYDROCLIMATIC VARIABILITY: 1950 – 1999

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Dedicated to

Late Dr. M.K. Soman
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Abstract

The interannual variability in precipitation over East Africa is well-understood. Many studies have identified the factors influencing the interannual variability of precipitation such as El Niño – Southern Oscillation (ENSO), the Indian Ocean Dipole Mode (IODM), and Atlantic Ocean sea surface temperature and pressure variations. The relatively arid conditions in much of the East African region are not understood fully. The objective of present study is to determine the meteorological association of aridity over East Africa with regional hydroclimatic variables as well as to find global teleconnections affecting spatial distribution of aridity over East Africa.

The East Africa Aridity index is calculated as the ratio of the mean seasonal precipitation to the mean seasonal potential evapotranspiration (after Budyko, 1974), and is used as a measure of aridity over East Africa. Principal components analysis was performed on the aridity index to identify characteristic modes of the temporal variability of the aridity index across East Africa. Correlation analysis was performed to identify the meteorological association in the interannual variability of aridity over East Africa and to find the global teleconnections, such as with ENSO, IODM, North Atlantic Oscillations (NAO), Tropical Atlantic SST Dipole (TASD), and Quasi Biennial Oscillations (QBO) in it. The first principal component of the aridity index was used for the correlation analysis. Correlations of the normalized difference vegetation index (NDVI) and Palmer Drought Severity Index (PDSI) with the first principal component of aridity index were calculated.

The aridity index over East Africa is driven by precipitation rather than potential evapotranspiration (PET). The PET over East Africa is driven by precipitation rather than temperature. Aridity over East Africa is well correlated with the NDVI and PDSI. The
ENSO influence on interannual variability of precipitation and hence on aridity is very much evident in all the seasons, while IODM influence is evident in the June – September season, the driest season for East Africa. Influence of NAO, TASD, and QBO was observed to be very small compared to that of ENSO and IODM. The teleconnections influencing the rainfall variability of East Africa also influenced variability in aridity.
Chapter 1: Introduction

Meteorologically East Africa is one of the most complex regions of the world. The climatic patterns over East Africa change over short distances, which are illustrated by the rainfall patterns (Nicholson, 1996). Though being at an equatorial position, East Africa receives less rainfall compared to the other tropical regions. Climate variability in seasonal and interannual climate is significant in East Africa, especially in the spatial and temporal distribution of rainfall. East Africa in the present study includes Somalia, Ethiopia, Kenya, Uganda, Tanzania, Rwanda, Malawi, Djibouti, Zambia; much of Sudan and Democratic Republic of Congo; and parts of Central African Republic and Zimbabwe. Different types of climate are found over various regions of East Africa. Though the annual temperature variation is small, the rainfall variability is significant. The Horn of Africa is dry and arid, while western equatorial East Africa is humid. Climate over Ethiopian and Kenyan Highlands is humid, while coastal East Africa is dry and arid compared to interior East Africa.

The spatial distribution of rainfall over East Africa depends mostly on the movement of intertropical convergence zone (ITCZ) over the region, but the local factors such as topography and the presence of lakes also play an important role. The rainfall over most of the East Africa is bimodal. The long rainy season of March-May and the short rainy season of October-December are the two important rainy seasons. Being at equatorial position, most regions of East Africa receives rainfall throughout year, but the rainfall receipt over East Africa is less compared to the other equatorial regions of the world. Rainfall is thus important. Although the greatest precipitation occurs in MAM
season, the OND season experiences larger degree of interannual climate variability (Clark et al., 2003). The interannual rainfall variability is influenced by various local and global factors including ENSO, Indian Ocean Dipole mode (IODM) and Atlantic Ocean SSTs. The study of these factors is important for better prediction of the future climate. El Niño modulates the climate over East Africa independently as well as through the Indian Ocean SSTs. Indian Ocean SSTs have been shown to influence the East African rainfall during non-ENSO years also. Atlantic Ocean SSTs also have a major impact on the rainfall receipt over East Africa (Nicholson and Entekhabi, 1987; McHugh, 2004a).

East Africa experiences floods and droughts often, which have a devastating impact on human health, livestock, forest, rangeland production, and especially on agriculture. The study of these factors is important for better prediction of the future climate. Extreme rainfall events and droughts are also shown to have a global connection. The extreme rainfall events of 1960 and 1997 highly increased the water levels of East African lakes, especially of Lake Victoria (Conway, 2002); increased White Nile discharge (Flohn, 1987); and caused widespread floods (Mistry and Conway, 2003). These extreme rainfall events have been shown related to ENSO (Ropelewski and Halpert, 1987; Ogallo, 1988) and IODM (Saji and Yamagata, 2003). 1997 extreme rainfall event was preceded by drought in 1996 and followed by severe drought in 1998. El Niño (La Niña) and IODM was associated with the extreme rainfall (drought) in 1997 (1998). 1997/98 El Niño event in east Africa produced above normal rainfall that led to widespread flooding and disease outbreaks (Linthicum et al., 1999) and the drought in the following year 1999/00 devastated agriculture production resulting into widespread famine (Anyamba et al., 2002).
The future climate change may add to the complexity of climate over East Africa by increasing the frequency and intensity of extreme events. This region is home to hundreds of thousands of farmers, pastoralists, and others who structure their life in structure with the environment; which is greatly influenced by climate. The climate change thus has social, economic, and demographic implications. Studies have predicted warming by 0.5°C to 0.7°C (IPCC, 2001; Hulme et al., 2001), increases in rainfall (IPCC, 2001), and more increase in potential evapotranspiration resulting in drier climate especially in arid and semi-arid regions of East Africa (Feddema and Freire, 2001) as the predicted future climate for East Africa. The future climate may also bring in addition, land-use changes through population growth and development pressure, altered frequency, intensity and extent of forest fires providing to feedback processes (IPCC, 2001), deficit in food production, and increased desertification.

Increase in CO₂ content of the atmosphere and increase in the frequency of warm phase ENSO events (Trenberth and Hoar, 1997; Timmerman et al., 1999, ICTP, 2001) are important factors among others increasing rainfall and frequency of extreme rainfall events in the predicted future climate of East Africa. The role of increased rainfall in resurgence of malaria in Rawanda and in the East African Highlands has been studied (Loevinsohn, 1994; Hay et al., 2002). Increase in rainfall may cause the environment vulnerable to outbreak and spread of waterborne diseases such as cholera and giardia. Because of the impact of short rain variability (in October to December) on agriculture, epidemic and cattle diseases, especially malaria and rift valley fever, it is important to accurately predict the strength of short rains ahead of time (Clark et al., 2003).
The relatively arid conditions in much of the East African region among other climatic aspects; are not understood fully (Nicholson, 1996). No major factors explain this aridity, while many local and global factors play a role (Trewartha, 1961; Anyamba, 1984). The variability in aridity is influenced by rainfall variability and temperature variability through change in potential evapotranspiration. Rainfall among other climatic factors is of maximum significance for the East African countries as the economies of the East African countries are largely based on the agriculture (Rodhe and Virji, 1976; Indeje et al., 2000; Seleshi and Zanke, 2004). The aridity is an important factor especially in agricultural industry in East Africa as the continued aridity leads to desertification of the land, meaning process of soil degradation. Most of the East Africa, especially Horn of Africa, coastal and southern Eastern Africa has vast arid and semiarid lands. Extended droughts in arid lands can initiate or exacerbate desertification process (IPCC, 2001) as arid lands can respond quickly to the seasonal fluctuations. Arid and semiarid lands thus are vulnerable also to climate change. Desertification is a meteorological as well as a social problem in East African countries. Governments of East African countries are participating in the program for combating desertification.

Intra- and inter-annual precipitation variability over East African region is very well studied. Precipitation dominates over potential evapotranspiration in determining the aridity, as the precipitation variations are large compared to those in potential evapotranspiration. Evapotranspiration is governed by demand of available energy mainly through net radiation and supply of water mainly through precipitation. Local and global factors influencing rainfall receipt also could influence aridity of the region. Study of temporal and spatial variability in aridity is important for the regions of high rainfall
variability, such as East Africa, for better understanding of impacts of climate change affecting rainfall over the region. The desertification and aridity over Sahara region, Sahel and West Africa are well studied, but such studies for East Africa lack.

Aridity over East Africa and its associations with local hydrology variables and with local and global climate, in each season, are studied in the present study using principal component analysis and correlation analysis. Aridity index calculated as ratio of the mean seasonal precipitation to the mean seasonal potential evapotranspiration (after Budyko, 1974) is used as a measure of aridity over East Africa. Principal component analysis is performed on aridity index in order to identify characteristic modes of temporal variability of aridity index across East Africa. Correlation analysis is performed in order to identify the meteorological association in the intra-annual variability of aridity over East Africa and to find the global teleconnections (such as with ENSO, Indian Ocean Dipole Mode, North Atlantic Oscillations and tropical Atlantic SST dipole and Quassibiennial Oscillations) in the interannual variability of aridity over East Africa. Aridity index is driven by precipitation variations more than potential evapotranspiration. Global climatic teleconnections influencing interannual variability in precipitation thus also influence the interannual variability in aridity index. ENSO, Indian Ocean Dipole Mode, Palmer drought severity index (PDSI) and normalized difference vegetation index (NDVI) over East Africa also correlated well with first principal component of aridity index over East Africa.

The objective of present study is to determine the meteorological association of aridity over East Africa with regional hydroclimatic variables as well as to find global teleconnections affecting spatial distribution of aridity over East Africa.
The second chapter is a literature review discussing mainly the interannual climate variability over East Africa and its association with global teleconnections. Aridity over east Africa calculated using precipitation and potential evapotranspiration data and climatic indices calculated using SST data are analyzed in conjunction with the hydroclimatic and circulation variables by performing correlation and principal component analysis. Data used for the present study and the methods are discussed in the third chapter ‘Data and Methods’. Annual climate cycle over East Africa in terms of the hydroclimatic and circulation variables is discussed along with the available literature in chapter four ‘East Africa Annual Climate’. The results are discussed and analyzed in ‘Results and Analysis’, the fifth chapter. Written in the chapter six are the conclusions.
Chapter 2: Literature Review

The inter-annual variability in East equatorial Africa is modulated by number of different phenomena including QBO, ENSO and SST variations in Indian and Atlantic Ocean and sunspot cycle. Spatial and temporal variability in East African rainfall show many major peaks ranging from two to twelve years, connected to different atmospheric phenomena. Ogallo (1980) and Ogallo et al. (1994) showed three major peaks in inter-annual variability of East African rainfall, centered on 2.5-3.7 years, associated with QBO; 4.8-6 year, associated with ENSO; and 10-12.5 years, associated with sunspot cycle. Rhode and Virji, (1976) and Indeje et al, (2000) analyzed trends and periodicity of annual rainfall using spectral analysis and the spectral analysis of time series revealed three major peaks, centered on 2-2.5, 3.5 and 5.6 years. A strong quasi-periodic fluctuation with time scale of 5-6 years in East African rainfall was observed by Nicholson and Nyenzi (1990) and Nicholson (1996) which were associated with ENSO and SST fluctuations in equatorial Indian and Atlantic Oceans. ENSO explains 50% of East African rainfall variance (Ogallo, 1988), and other factors explain the remaining variance. The other factors include tropical storms, easterly waves, jet streams, continental low-level trough and extratropical weather systems (Ogallo, 1989; Indeje et al, 2000). Long-term rainfall fluctuations in eastern Africa are linked to quasi-global climatic fluctuations such as Southern Oscillations (SO) and QBO (Nicholson, 1996). The interannual variability of rainfall receipt on East Africa is shown to be influenced by ENSO, North Atlantic Oscillations (NAO), circulation features associated with southern Atlantic Ocean, and Indian Ocean Dipole Mode (IODM), by different studies, as described in following sections.
In most areas of East Africa normalized difference vegetation Index (NDVI) is a sensitive indicator of the interannual variability of rainfall (Davenport and Nicholson, 1993). Davenport and Nicholson (1993) observed a strong relationship between NDVI and rainfall both spatially and temporally over East Africa. Anyamba et al. (2001) showed the increase in positive NDVI, suggesting greener than normal condition prior to and during the rainy season (from October 1997 to May 1998) over East Africa during 1997-98 ENSO events. Anyamba et al. (2002) also showed the association of temporal and spatial pattern of NDVI over East Africa to Pacific Ocean SSTs during 1997-98 and 1999-00 ENSO events. Anyamba et al. (2000) also showed the close association of increase in NDVI with central Pacific and western Indian Ocean warming during ENSO event of 1997-98 using correlation analysis.

Dai et al. (1997) and Dai et al. (2004) studied global PDSI (Palmer Drought Severity Index) and its time series and performed principal component analysis on PDSI for time period of 1900 to 2002 and used correlation analysis to conclude that soil moisture correlates significantly with PDSI. The study suggested that PDSI represents surface moisture conditions better than precipitation alone (Dai et al., 2004) as the correlation of PDSI with soil moisture was higher compared to its correlation with precipitation over US and Eurasia. PDSI partly reflects soil moisture memory and PDSI has long memory so that summer PDSI could be affected by surface warming and subsequent soil moisture changes during winter and spring. Pattern of global PDSI in Dai et al. (2004) indicated large surface warming since 1950 causing drying in western, central and southern Africa along with some other continents.
An Empirical Orthogonal Function (EOF) analysis of East Africa rainfall and global CMAP rainfall data from 1961-2001 by Schreck III and Semazzi (2004) isolated a separate eastern Africa mode in rainfall consistent with warming of global climate. They found that the rainfall over eastern Africa increases with a time evolution that is consistent with increasing global warming. They plotted the time series of second EOF of regional rainfall has increasing trend from 1979 to 1999 as it is in the time series of global warming index. The time series of gauge-based eigenmode and of surface temperature (from Climatic Research Unite archive) (for 1963 – 1990) also show increasing trend and close association with each other. Both of these comparisons of time series are the confirmations of the evidence that a mode in eastern Africa rainfall is associated with global warming. Conway (2002) also have indicated that larger parts of Somalia, Ethiopia, Sudan and northern Uganda have experienced increase in rainfall during 90s, while these regions had experience reduced rainfall during 1970s-80s.

2.1 ENSO and East Africa Climate Variability

A broad scale phenomena associated with formation of warmer than usual sea surface temperatures (SSTs) in tropical Pacific Ocean along the coast of Ecuador and Peru that would occur around the Christmas time is know as the El Niño phenomena. La Niña is counterpart to El Niño and is characterized by cooler than usual SSTs in tropical Pacific Ocean. During normal circumstances in tropical Pacific Ocean cold deep Ocean water upwells in eastern Pacific Ocean as the southeast trade winds drive water westward. This upwelling results in shallower thermocline in eastern Pacific than in western Pacific. During El Niño, trades weaken, anomalously low pressure and warm SSTs are found in eastern pacific; while anomalously high pressure and cold SSTs are
found in western Pacific; upwelling in eastern pacific is suppressed and thermocline in eastern Pacific deepens; convective activity and rainfall is enhanced in Eastern Pacific, while they are suppressed in western Pacific; and subequatorial counter current strengthens. During La Niña, trades strengthen; anomalously low pressure and cold SSTs are observed in eastern Pacific; while anomalously high pressure and warm SSTs are observed in western Pacific; oceanic upwelling in eastern Pacific is enhanced, making thermocline even shallower; and subequatorial counter current slackens.

The time between two successive El Niño events as well as the intensity of El Niño is irregular. It recurs every 3 to 7 years. El Niño is associated with the changes in air pressure above Pacific Ocean, which is known as Southern Oscillations (SO). SO is a periodic reversal of the pressure pattern across the tropical Pacific Ocean. The SO index (SOI) is a measure of strength and phase of SO and indicates the state of Walker circulation. SOI is calculated as the air pressure difference between Tahiti and Darwin. During El Niño, SOI is strongly negative. The process of deep convection exchanges heat between Earth's surface and the free atmosphere. The changes in SSTs during ENSO affect atmospheric circulation on global scale, thus affecting global climate. SSTs over four regions (Nino 1+2, Nino 3, Nino 4 and Nino 3.4) of tropical Pacific are generally used to monitor ENSO. The largest SST variability is observed in region Nino3, while Nino 3.4 region also has large SST variability. Nino 3.4 region is between 5°S-5°N and 170°W-120°W. ENSO index measured in any Nino region is the SST anomaly observed in that region in the units of degree Celsius.

Many studies have revealed the existence of the connection between interannual rainfall variability in East Africa and ENSO. The extreme rainfall anomalies also show
connection with ENSO (Ropelewski and Halpert, 1987, Ogallo, 1988). According to Ogallo (1988), ENSO explains 50% of East African rainfall variance, and other factors explain the remaining variance. Net impact of ENSO over east African climate is a result of complex interactions and alignments among various climatic factors. ENSO influences the amount of rainfall in East Africa, but this influence is different in different rainy seasons of year. According to Nicholson and Kim (1997), in equatorial East Africa, rainfall is enhanced in short rain season (OND) of the ENSO year, but reduced during the main rain season (MAM). During OND, the rainfall on East Africa increases (decreases) during the warm (cold) ENSO events. The harmonic analysis in this study indicates that ENSO significantly modulates the short rain period and has little impact on the main rain season.

Rainfall activity on East Africa is been shown related to the changes in Walker circulation. A mechanism for this is suggested by Hastenrath et al. (1993) and Hastenrath and Polzin (2003) that the anomalous zonal circulation cell over Indian Ocean is driven by the SO, where SO is an atmospheric component of ENSO. In their study using observed data they hypothesized that during the positive phase of SO, convection is enhanced over eastern Indian Ocean, while the western Indian Ocean is colder and the convection over western Indian Ocean and Eastern Africa is suppressed. This hypothesis suggests that the walker circulation cell over Indian Ocean is influenced by SO. The ascended air in eastern Indian Ocean descends over western Indian Ocean and East Africa. The SO positive phase makes stronger this ascent and descent of air over Indian Ocean, through Indian Ocean SSTs. Thus the rainfall over East Africa is increased (decreased) during a positive (negative) SO event. This mechanism does not explain all
the anomalous rain events of East Africa, for example, the floods in 1961. Few anomalous rainfall events could be explained using the mechanism of Indian Ocean Dipole Mode (IODM), with no influence from SO.

The EOF analysis to find ENSO signals in East Africa rainfall seasons was conducted by Indeje et al., 2000. In this study the East Africa region was restricted to the three countries- Uganda, Kenya and Tanzania. They found above normal rainfall conditions during MAM over costal areas of Kenya and Tanzania during the ENSO onset year, while below normal rainfall conditions over northern Kenya and Uganda, during MAM. They concluded that during the ENSO onset year, the wetter than normal conditions could be expected over coastal regions of East Africa and over central Tanzania during MAM, while, dryer than normal rainfall conditions could be expected over central Kenya and northern Uganda during JJAS. This study also observed a tendency of late onset of long rain season followed by early cessation over most parts of northern East Africa during ENSO onset year. During JF of post-ENSO years, they observed wetter conditions over most parts of East Africa. An east-west dipole pattern in rainfall over equatorial East Africa was observed during ENSO year (Ogallo, 1988, Indeje et al., 2000). The study by Indeje et al., 2000 also discuss the socio-economic impacts of the change in rainfall conditions and patterns over East Africa due to the impact of ENSO. Their study suggests that the guidance about the expected shift in rainfall due to ENSO to the farmers in East Africa will prove beneficial as they can delay planting by month, which will be beneficial for final yield. The study also suggests that the higher than normal JJAS rainfall and poor and late short rains expected during post-ENSO year can fail production of beans and potatoes due to failure of soil to sustain soil
moisture as evaporation rates during July to October are high. Farmers in Tanzania could be advised for planting alternative crops requiring less soil moisture due to the moisture stress during ENSO year.

The occurrence of various droughts in Africa, especially in Southern Africa and the Horn of Africa are caused by physical processes related to the occurrence of ENSO events (Wolde-Georgis, 1997). Ethiopian drought is caused by ENSO, along with SST anomalies in the Southern Atlantic and Indian Oceans combined with anthropogenic activities (Haile, 1988). There is a remarkable correspondence between annual rainfall in Ethiopia and ENSO events. ENSO events and SST anomalies affect rainfall distribution in Ethiopia by displacing and weakening the rain-producing air masses (Haile, 1988). 1997/98 El Niño event in east Africa produced above normal rainfall that led to widespread flooding and disease outbreaks (Linthicum et al., 1999) and the drought in following year 1999/00 devastated agriculture production resulting into widespread famine. Galvin et al. (2001) have reviewed the effect of 1997-98 ENSO event on Massai herders in northern Tanzania. The Massai economy is based on livestock products, livestock sales and agriculture. 1998 ENSO rains affected livestock sales as the waterborne diseases affected livestock significantly, while the drought in 1997 also increased livestock sales.

La Niña modulates rainfall in East Africa in some seasons. Strong positive anomalies are observed in July-September rainfall season, during La Niña year (Nicholson and Selato, 2000). The occurrence of El Niño tends to enhance the MAM and OND rainfall over Coastal East Africa, while La Niña tends to diminish it (Figure 2.1). Anyamba et al. (2001) and Anyamba et al. (2002) studied vegetation response to ENSO
Figure 2.1: Composite of coastal African rainfall showing monthly mean (solid line), monthly composites for El Niño years (dotted line) and monthly composites for La Niña years (dashed line). The shaded region indicates the monthly mean ± 1 standard deviation. (Figure is obtained from Clark et al., 2003)

over East and southern Africa during period of 1997-2000 using satellite measurements of vegetation conditions in the form of normalized difference vegetation index (NDVI). They found that the transition from El Niño to La Niña conditions during period 1997-2000 had significant effects on precipitation and vegetation biomass conditions over East Africa. A marked contrast in vegetation anomaly pattern was observed during the peak period of warm (1997/98) and cold (1999/00) phases of ENSO conditions. Reversal in NDVI response pattern was observed over East Africa from positive during El Niño and negative during to La Niña events. They found a strong relationship between NDVI and Pacific Ocean SSTs (Niño 3.4 region) for East Africa.

ENSO also influence air temperature and pressure in tropics. It is shown that during warm ENSO events, air temperature is higher in most of the tropics including East
Africa (Diaz and Kiladis, 1992) and air pressure is also higher in East Africa (Trenberth and Shea, 1987). Interannual and seasonal variation in these climatic variables may have an impact on crop yields over East Africa (Plisnier, et al., 2000). By correlation analysis of temperature variables over East Africa and Pacific SSTs Plisnier, et al. (2000) observed that Pacific SST anomaly causes lower than usual temperatures in East Africa about 6-8 months later, thus found positive linkage of air temperature and ENSO. Plisnier performed correlation analysis on climatic variables such as air temperature and relative humidity and land surface variables such as NDVI and Pacific SSTs. These variables were found to have different type of teleconnection with ENSO, leading to a complex impact of ENSO on the ecosystem. They observed ENSO-driven changes in NDVI and air temperature indicating changes in surface soil moisture and evapotranspiration. They also suggested that the exact response of vegetation to ENSO induced climate variations depend on land cover. The found that the ENSO impact over ecosystem of East Africa is highly differentiated in space and suggested that surface attributes, as determined by geology, soils and vegetation might also influence the magnitude and time lag of the ENSO impact.

ENSO signal is also observed in Atlantic and Indian Oceans. Both of these oceans affect the rainfall anomalies over East Africa as described in next to sections. Indian Ocean tends to control these rainfall anomalies during the warm phase of ENSO and warm phase response appears more consistently in Indian Ocean, while, in Atlantic Ocean, the cold phase response appears more consistently (Nicholson, 1997). ENSO response appears more consistently over equatorial East Africa during warm phase of ENSO than during cold phase of ENSO (Nicholson, 1997).
2.2 Atlantic Ocean and East Africa Climate

Many studies have demonstrated the association of Atlantic Ocean SST variations and circulation variability and East Africa precipitation receipt. Westerlies over Atlantic Ocean carry moist air masses over to East Africa. These air masses are uplifted due to highlands of western East Africa and produce precipitation. Thus westerly flow from Atlantic is strongly associated with East Africa rainfall, while the strength of this flow depends on fluctuations in location and intensity of ITCZ (McHugh, 2004a).

2.2.1 North Atlantic Oscillations (NAO)

The NAO is a large-scale alternation of atmospheric mass with centers of action near the Icelandic Low and the Azores High. It is the dominant mode of atmospheric behavior in the North Atlantic sector throughout the year, but is most pronounced during northern hemispheric winter. McHugh and Rogers (2001) have shown the association between North Atlantic Oscillations and Southeast Africa rainfall in the months of December-February. They also demonstrated that precipitation significantly decreases during easterly flow episodes prevalent across East Africa between the equator and $20^\circ$S during negative phase of NAO, while during the positive phase of the North Atlantic Oscillation and during westerly flow episodes, precipitation significantly increases in East Africa south of equator. However, according to Nicholson (1996), some outbreaks of moist Atlantic westerlies are associated with precipitation deficits over eastern Kenya, which suggests that the impact of the westerly flow regime may be modulated by local factors in equatorial East Africa. Results of the studies by McHugh (2004a) show that the westerly flow from Atlantic Ocean influences the fluctuations in location and intensity of
the localized convergence zone across east Africa on which depends the precipitation over east Africa.

2.2.2 Tropical Atlantic SST Dipole (TASD)

Tourre et al., 1999, through the analyses of SST and SLP over the entire Atlantic basin found that the decadal pattern in the north Atlantic has a coherent SST pattern in the Tropical Atlantic as well, indicating tropic-extratropic interactions at the decadal time scale. Chang et al., (1997) and Mehta (1997) found a dominant decadal scale SST pattern in tropical Atlantic from both model and data. This pattern has an out of phase relationship between the north and south of the equator in the SST anomalies, (Chang et al., 1997). (See figure 2.17 for Tropical Atlantic SST Dipole and Mechanism of Tropical Atlantic Variability). Tropical Atlantic SST Dipole is shown to have of great importance to climate variability in Northeast Brazil and to the rainfall variability in northeast South America and Sahel. Not much study has been done to see the affects of TASD on climate variability of east Africa. TASD affect the trade wind strength and hence could shift ITCZ to south or to the north of its mean annual position, affecting atmospheric circulation. The global effects of TASD are not widely known.

2.2.3 South Atlantic SST Anomalies

The annual cycle of East Africa rainfall is strongly associated with the seasonal migration of ITCZ. The trade winds converge in ITCZ with the South Atlantic air masses causing rainfall generation over East Africa. Westerly flow from South Atlantic is an important predictor of rainfall receipt over East Africa (McHugh, 2004b) due to the above said association. EOF analysis by McHugh (2004b) has shown that this strong positive linkage is through enhanced low-level moisture convergence; uplift, decreased
lower tropospheric stability, and increased precipitable water during both the rainy seasons (MAM and OND) of East Africa. Moist airflow from south Atlantic is associated with anomalous rainfall receipt across East Africa (Nicholson, 1996; McHugh and Rogers, 2001; McHugh, 2004a). The moist south Atlantic air is cooled adiabatically over East African Highlands generating rainfall is one of the hypothesized mechanisms explaining the association between the south Atlantic moist airflow and anomalous East Africa rainfall (McHugh and Rogers, 2001, McHugh, 2004b). This south Atlantic moist flow converging over East Africa significantly increases near surface specific humidity and precipitable water in the unstable atmosphere over East Africa, while; the convergence and uplift of air causes the atmospheric instability.

Commonly the moisture inflow from South Atlantic is associated with enhanced East Africa rainfall, although, Nicholson, 1996 has shown that it is also associated with rainfall deficit over East Africa in some cases.

Nicholson and Entekhabi (1987) have shown the association of Atlantic SST variability along the Benguela coast and equatorial East Africa rainfall during warm and cold phases of SST variation, implied by rainfall anomalies. Composite analysis in this study showed that the rainfall anomalies over equatorial Africa were of opposite signs during warm and cold phases of SSTs in Atlantic Ocean along Benguela coast in such a way that the rainfall is enhanced during warm phase and subnormal rainfall occurred in cold phase. The rainfall enhancement is particularly marked over equatorial eastern Africa, suggesting the influence of atmospheric and oceanic variables over Atlantic Ocean on rainfall.
2.3 Indian Ocean and East Africa Climate

A study using atmospheric general circulation model by Goddard and Graham (1999) suggested that the Indian Ocean SSTs exert a greater influence over East African short rains compared to SSTs in Pacific Ocean. Indian Ocean SSTs also influence rainfall over East Africa, especially over Horn of Africa and coastal East Africa during JJAS through South Asian monsoon circulation. Empirical studies have shown a strong correlation between western Indian Ocean SSTs and short rains variability (Mutai et al., 1998; Nicholson 1997; Ogallo, 1998).

2.3.1. Indian Monsoon

The influence of Indian Ocean on the annual and interannual climate of East Africa is well studied. Indian Ocean is adjacent to the eastern coast of Africa. The northwest and southeast monsoon winds over Africa are associated with Indian Ocean. When the northwest monsoon winds have continental origin over Sahara region, they are dry, but during the reversal, they originate in northwest Indian Ocean and are much moister. The southeast monsoon winds originate in Indian Ocean. The Somali Jet, originating in Indian Ocean near Madagascar Island, though present throughout the year, is well developed during the onset of south Asian monsoon (Nieuwolt, 1977). During May-July, the inland penetrating Somali Jet brings moisture from Indian Ocean to Horn of Africa as well as to the coastal regions of East Africa.

Vyazilova (2001) studied the influence of Indian Ocean on East Africa rainfall during ENSO events and found that strong rainfall anomalies over East Africa are connected with moisture transport anomalies in low atmosphere over tropical Indian Ocean and Western Pacific during warm and cold events of ENSO.
2.3.2: Indian Ocean Dipole Mode (IODM)

The Indian Ocean Dipole Mode (IODM) is a climate mode that occurs interannually in the tropical parts of the Indian Ocean. Indian Ocean dipole mode was first identified by Saji et al. (1999) and Webster et al. (1999). Some studies prefer to call it as Indian Ocean Zonal Mode (IZOM), as the term dipole suggests simultaneous variation in the east and west parts of Indian Ocean. It is an interannual mode of variability in Indian Ocean SSTs, which has east-west structure. IODM is rooted in subsurface equatorial ocean dynamics (Rao et al., 2002). During a positive IODM event, the sea-surface temperature (SST) drops in the southeastern part of the Indian Ocean; off the northern coast of Australia, the eastern coast of Japan and throughout Indonesia; while the SST rises in the western equatorial Indian Ocean; off the eastern coast of Africa, from the northern half of Madagascar to the northern edge of Somalia. Negative IODM mode is characterized by a reversal of east-west equatorial SST gradient across Indian Ocean such that SST increases in southeastern Indian Ocean, while SST drops in western Indian Ocean. Negative anomalies in the east tend to lead the positive western anomalies by one season (Clark et al., 2003). During the negative IODM, the anomalies reverse. Convective patterns increase in the northern half of Africa, India and off the eastern coast of Africa during IZOM.

IODM appears to have a complicated relationship with ENSO. IODM occurs sometimes simultaneously with ENSO and also at other times when Pacific Ocean SSTs are not anomalous. Beltrando and Camberlin (1993) described the teleconnections between ENSO, Indian Ocean surface fields and Ethiopian summer and Somali autumnal rainfall. Heavy summer rains in east central Ethiopia are associated with lower
pressure in N-W and in east of Indian Ocean and over higher pressure over the eastern part of the Pacific Ocean and to southern Indian Ocean to a lesser extent. Enhanced rainfall in East Africa (e.g. in 1961) may be an integral part of IZOM (Black et al., 2003). Through data analysis, black et al., 2003 have shown that extreme East Africa short rains (Sept-Nov) are associated with Indian Ocean SST anomalies and IZOM. Enhanced rainfall activity over East Africa is produced during IZOM events due to reversal of zonal SST gradient in Indian Ocean. Black et al., 2003 suggested that El Niño can trigger an IZOM event and therefore, heavy rainfall over East Africa under the right circumstances and found that heavy rainfall in East Africa is consistent with the interdecadal variability in the activity of the IZOM and ENSO. The Indian Ocean in connected to the Pacific Ocean through the changes in the Walker circulation and the Indonesian Throughflow. Thus Indian Ocean is highly sensitive to the SST changes in Pacific Ocean. Nicholson and Kim (1997) also found a connection between ENSO and rainfall not just over East Africa, but also over most of the African continent and suggested a linkage through ENSO induced SST anomalies in India Ocean, which modulate the rainfall variability over East Africa.

Figure 2.2: Schematic of the anomalous Walker and Hadley circulation during strong El Niño years showing the development of the local Hadley circulation in the vicinity of the Maritime continent and the associated easterly and southerly wind anomalous over the eastern Indian Ocean. Figure is obtained from Black et al., 2003.
According to Behera et al. (2004), co-occurrence of IODM and ENSO does not necessarily suggest that ENSO influence on East Africa rainfall receipt (especially during short rains) is through IODM as IODM can evolve independent of ENSO. Figure 2.2 shows the schematic of how ENSO and IODM are related.

About 30% of positive IODM events co-occur with El Niño events (Rao et al. 2002). Correlation analysis by Saji and Yamagata (2003) has shown that for Eastern Africa among other regions, rainfall anomalies have negligible correlations with ENSO during IODM years (Figure 2.3). IODM events independent of ENSO occurred in the years 1961, 1967, 1977, 1983, 1994 (all years positive phase), 1958, 1960, 1974, 1989, 1992, 1993 (all years negative phase) and those co-occurred with ENSO in years 1963, 1972, 1982, 1997 (all years positive mode), 1964, 1971, 1975, 1996 (all years negative mode) (Saji and Yamagata, 2003). IODM-independent ENSO years are 1965, 1969, 1976, 1986, 1987 (all years warm phase), 1967, 1970, 1973, 1978, 1984, 1985, 1988 (all years cold phase) (Saji and Yamagata, 2003). The Figure 2.3 also indicates the importance of IODM influencing rainfall anomalies over East Africa. Composite of OND rain anomaly for 19 IODM events (figure 2.3a) and that for 11 ENSO independent IODM events shows high rainfall anomaly over East Africa. OND rain anomaly also correlates well over East Africa to IODM index (Figure 2.3e) during all IODM events including ENSO-independent events (Figure 2.3f).

Although, Black (2003) have shown that the observed teleconnection between ENSO and East African boreal autumn rainfall is a manifestation of the dynamic link between El Niño and the Indian Ocean dipole, IODM independently influences the rainfall anomalies over East Africa. In the study using observed data The hypothesis by
Hastenrath et al. (1993) and Hastenrath and Polzin (2003) that during the positive phase of SO, convection is enhanced over eastern Indian Ocean, while the western Indian Ocean is colder and the convection over western Indian Ocean and Eastern Africa is suppressed is examined and reproved by Behera et al. (2004) in a CGCM study.

**Figure 2.3:** The composite OND rain anomaly over Africa for a) 19 IODM events b) 11 ENSO-independent IODM events c) 20 ENSO events d) 12 IODM-independent ENSO events. Contour interval is ±1. Simple correlation of rain e) on IODM index g) on El Niño index (Nino 3 region). The partial correlation of the rain f) on IODM index independent of El Niño index and h) on El Niño (Nino 3 region) independent of IODM index. Correlation coefficients are multiplied by 10 and are plotted with contour interval of ±2. (Figure is obtained from Saji and Yamagata, 2003).
2.4 Extreme Rainfall Events

Interannual variability in East Africa rainfall receipt shows the presence of extreme rainfall events. In year 1878 heavy rainfall occurred (Nicholson, 1999) which flooded Congo River. Extreme floods occurred in 1916-18. This episode was smaller in magnitude and was limited to a small area in the East Africa, in Lake Victoria, southern Sudan, southwestern Ethiopia (Conway, 2002). Rainfall increased 10 to 40% over these regions between October 1916 and September 1917 (Conway, 1997). The other examples of extreme rainfall events are in 1961 and in 1997. In these years the rainfall was heavy over almost entire East Africa. In both of the years, heavy rainfall occurred during short rains (especially October-November). The heavy rainfall caused flooding in East Africa in these events causing significant economic disruption.

Many studies have been performed to investigate the reasons causing these extreme rainfall events. An analysis of velocity potential over East Africa by Mistry and Conway (2002) proved the hypothesis that the anomalous OND 1961 extreme rainfall episode was a consequence of the strength and the position of local Indian Ocean cell in Walker circulation, to be untrue. They also showed that the Indian Ocean SST anomalies were the important factor influencing the determination of moisture fluxes and supply of latent heat in OND 1961 and in subsequent MAM 1962 rainfall event. They found the 1961 rainfall anomaly as a consequence of reversal in the prevailing wind regime in Indian Ocean; SST anomalies in western Indian Ocean, and cool moist air mass from southern Indian Ocean. The combination of these factors created instability over Lake Victoria region increasing significantly its water level.
Webster et al. (1999) and Saji et al. (1999) provided the driving mechanism of the extreme rainfall events in East Africa in 1961 and 1997 as the dipole reversal in atmospheric circulation and in Indian Ocean SSTs, although, the dipole reversal in Indian Ocean has not always associated with occurrence of extreme rainfall anomalies. The occurrence of extreme rainfall events of 1878 and 1997 also show co-occurrence of ENSO, but 1961 was not an ENSO year. Thus the influence of ENSO on these extreme rainfall events remains unclear. The influence of ENSO and IODM over predictability of East African rainfall is not fully been understood.

2.5 QBO and East Africa Climate

Quasi Biennial Oscillation (QBO) is a quasi-periodic oscillation in stratospheric and tropospheric zonal wind between westerly and easterly modes (Nieuwolt, 1977; Indeje and Semazzi, 2000). QBO are known to have influence on tropical climate variability. QBO plays a role in modulating African rainfall (Mason and Tyson, 1992; Jury et al., 1994). Jury et al., (1994) have shown that during the QBO easterly phase, tropical cyclones in the Indian Ocean are more frequent and convection increases significantly over Madagascar, which affects rainfall over southeast Africa. They also showed the association of QBO with Walker circulation cell connecting Africa and Indian Ocean, which combines with westerly phase QBO resulting in rising tropospheric motion over Africa.

QBO related spectral peaks in East Africa rainfall have been reported observed distinctly in many studies (Rodhe and Virji, 1976; Ogallo, 1994; Nicholson and Entekhabi, 1986; Ogallo et al., 1994). Nicholson and Entekhabi (1986) examined the
temporal characteristics of African rainfall variability. Their results of time series analysis indicated QBO peak in equatorial and southern Africa rainfall. They also showed strong coherence between QBO and SO through cross-spectral analysis, suggesting influence of QBO on equatorial and southern Africa rainfall variability. They found that coherence with the SO is particularly strong in the QBO range of 2.2-2.4 years (Figure 2.5). Their co-spectra analysis suggested an inverse relationship between SOI and rainfall over equatorial Africa, meaning higher (lower) rainfall in low-index (high-index) years.

Figure 2.4: QBO in African rainfall (Nicholson and Entekhabi, 1986). Left figure: coherence-squared between rainfall and Southern Oscillation in the QBO range for each of the regions with the discrete spectra peak in rainfall in this range. Shading indicates values exceeding the 95% confidence level. Right figure: Distribution of spectral peaks in rainfall in QBO range (2.2-2.4 years). Shading indicates 95% confidence level, hatching indicated 90% confidence level.

Ogallo et al. (1994) through spectral analysis identified the peaks in interannual variability of East Africa rainfall at 2.5-3.7 years corresponding to QBO. Their results indicated the 28 months period dominant in zonal wind component in 1966-1987. They
also showed based on the reversal of zonal winds, significant (at 5% level) association between East Africa (Kenya, Uganda and Tanzania) rainfall and QBO signal.

Indeje and Semazzi (2000) performed correlation analysis and composite analysis to investigate teleconnections between MAM rainfall over East Africa and equatorial lower stratospheric zonal winds through simultaneous and lag correlations to search the optimal predictive potential of QBO. Their results indicated strong teleconnections between QBO (lower equatorial stratospheric zonal winds at 30 mb) and East Africa rainfall during MAM. They found the weakening of this relationship with decreasing time lag between the two phenomena. Their contingency analysis indicated the 60 % (63 %) probability of co-occurrence of above (below) normal rainfall over western highlands of East Africa during westerly (easterly) phase of QBO. Results of correlation analysis suggested that 36 % of the variability of the long rains season over East Africa is associated with QBO. They found phase of the QBO before the season to be a good predictor of East Africa long rains and of non-occurrence of drought over East Africa.

The EOF analysis performed by Indeje et al. (2000) showed the annual cycle of East Africa rainfall in which relationship between QBO and seasonal rainfall is shown to be significant during boreal summer (June-August) and the weakest during December-February. The non-significant relationship between long rains of East Africa and QBO has been observed during strong ENSO years (Indeje et al., 2000; Indeje and Semazzi, 2000), for example, during the long rains of 1966, 1973 and 1983. Indeje and Semazzi (2000) also indicated the relation between Indian Ocean dipole mode and QBO. They postulated a mechanism of interaction between Walker circulation cell over Indian Ocean
and QBO (Figure 2.6), which also indicates the association of Indian Ocean dipole mode, QBO and East Africa rainfall.

**Figure 2.5:** Schematic of large-scale horizontal interaction between circulation systems over the African-Indian Ocean sector (Indeje and Semazzi, 2000).
2.6 Past Climate Variability in Equatorial East Africa

A study of 300 years of East African climate variability by Dunbar (2003) presented a result examining a monthly resolution stable oxygen isotope, longest coral climate record acquired from a living coral head from Malindi marine reserve, Kenya, that there is a clear warming trend of about 15° C, which accelerated in 20th century and is superimposed on strong decadal temperature variability. This study claims that East African SST and rainfall are better correlated with Pacific SST (ENSO) compared to Indian Ocean SST for all the study period from 1696 to 1996. These coral records indicate a major cool and dry period from 1750 to 1820, which is the largest multi-decadal anomaly of past 300 years that coincides with historical Lapanarat Drought (Cane and Molnar, 2001), and are strongly linked to multi-decadal tropical cold SST anomalies and continental droughts in East Africa. These coral records are found to be strongly coherent with ENSO indices compared to India or East-African rainfall indices.

Verschuren et al. (2000) presented a decade-scale reconstruction of rainfall and drought over equatorial East African region, based on lake level changes and salinity fluctuations of Lake Naivasha, in Kenya. The proxies used were sediment stratigraphy, species compositions of fossil diatom and midge assemblage. The results of the study reveal that the equatorial East Africa has alternated between contrasting climate conditions, over the past millennium. The results of the study combining sediment records and hydrological response of Lake Naivasha suggest that East Africa was drier during the Mediaeval Warm Period (AD 1000 to 1270) compared to present climate. They report on three prolonged dry periods (1390-1420, 1560-1625 and 1760-1840), which were severe compared to twentieth century droughts. They also reported that
during the Little Ice Age (LIA, AD 1270–1850) fairly wet conditions were present. These wet conditions had breaks of persistent aridity conditions (figure 2.4). This aridity was severe compared to present aridity level. They showed that the decade-scale variations might be related to sunspot cycle and the resultant changes in solar radiation. The study indicated that the highest inferred rainfall of the past 1,100 years coincides with the 'Maunder minimum' of solar radiation.

**Figure 2.4**: Palaeo-environments in eastern Africa as a function of time and latitude since 15 Myr ago. Green: humid environments; brown: more arid woodlands; and tan: more arid grasslands and deserts. (From Cane and Molnar, 2001)
Other studies also indicated the similar influence of LIA on East Africa climate. Johnson et al. (2001) reported colder conditions in tropical Africa between 1570 and 1820. LIA also had impact on hydrological cycle over East Africa, as indicated by droughts, desiccation and low level of East Africa lakes (Nicholson and Yin, 2001).

Warm water in southern Pacific Ocean moving westward along the equator is known as Southern Equatorial Current. This current flows along the north coast of New Guinea to Halmahara eddy, which is present to the East of Halmahara Island, and turns to flow eastward in North Equatorial Countercurrent. A study by Cane and Molnar, 2001, proposed that 3-5 million years (Myr) ago, when New Guinea was to the 2°-3° south of its present position and when Halmahara was a small island, warm water from south Pacific should have passed into the Indian Ocean, increasing SSTs in Indian Ocean and precipitating a rainier climate in East Africa. Figure 2.4 indicates that, climate in eastern Africa has evolved from moist and warm to arid and slightly cooler over past 3-4 Myr (Cane and Molnar, 2001). This study suggests that the aridification of East Africa at around 4 Myr ago was caused by northward movement of New Guinea and increasing the size of Halmahara island, closing the Indonesian seaway between Pacific and Indian oceans and thus by preventing the warm water of southern Pacific from entering into Indian Ocean.

2.7 Climate Change and East Africa

Africa is very vulnerable to climate change, the overall capacity for Africa to adapt to climate change being low (IPCC, 2001). Numbers of factors including climatic (global warming, change in sea level, change in rainfall, exacerbation of desertification) and sociopolitical (water management, decline in agricultural production, population
growth, food security at risk) factors play important roles in making Africa vulnerable to climate change. 0.5°C to 0.7°C warming has been observed over Africa in 20th century (IPCC, 2001, Hulme et al., 2001). This warming is slightly larger in JJA and SON seasons than in DJF and MAM seasons (Hulme et al., 2001). A decrease in rainfall over Sahel and increase in rainfall over East Africa is also been observed (IPCC, 2001). A study by Intergovernmental Panel on Climate Change (IPCC) indicates warming across Africa in future, ranging from 0.2°C to 0.5°C, with greatest warming over interior semi-arid margins of Sahara and central southern Africa. The dominant impact of global warming is predicted to be a reduction in soil moisture in subhumid zones and a reduction in runoff (IPCC, 2001). Study using GCMs by Hernes et al. (1995) has suggested warming of the equatorial Africa countries of about 1.4°C. These studies and a study by Joubert et al. (1996) have suggested a sea level rise around coast of Africa under the climate change scenario. A warming is observed over Ethiopia and Sudan along with few other African countries (Hulme et al., 2001). This study by Hulme et al. (2001) using different emission scenarios have suggested 0.9°C to 2.6°C warming by 2050 globally, while warming of below 0.2°C to over 0.5°C per decade for Africa.

Soil degradation and global warming are the two important factors affecting the future climate of Africa (Feddema and Freire, 2001). Though the overall precipitation is predicted to increase over Africa, the increase in the potential evapotranspiration will be greater, resulting in the drier climate especially over the semiarid and arid regions of Africa. The impact due to soil degradation is less compared to that of global warming (Feddema and Freire, 2001).
As explained in section 2.1, ENSO is an important factor affecting interannual climate over East Africa. Climate change may increase the frequency of ENSO warm phases (IPCC, 2001), thus increasing precipitation over East Africa, East Africa being in phase with warm ENSO episodes. This increase in frequency of warm ENSO episodes could be by increasing warm pool in the Tropical western Pacific or by reducing the efficiency of heat loss (Trenberth and Hoar, 1997, Timmerman et al., 1999).

Precipitation and temperature are important climatic factors controlling desertification. Extended droughts (for example in Ethiopia) can initiate or exacerbate desertification. The global warming and change in ENSO frequency, along with other factors may affect the temperature, rainfall receipt and intensity and frequency of extreme events like drought and floods, affecting aridity over East Africa. Change in temperature and rainfall will have many negative impacts on human health (IPCC, 2001), for example, on sanitary conditions and on outbreak of diseases.

Association of incidence rates of diseases including Malaria and Rift Valley fever with the change in rainfall is been observed before (Loevinsohn, 1994, Epstein et al., 1998, Zhou et al., 2004, Linthicum et al., 1999) in parts of East Africa. A study performed by McHugh (2004c, CO2) using nineteen coupled general circulation models with 1% per annual increase in CO2 scenario, evaluated model output for impacts on East Africa. Few models showed increase in rainfall receipt over East Africa in CO2 enriched future atmosphere, the reasons being modification of ITCZ over East Africa and changes in the frequency of extreme events especially over interiors of East Africa. It was observed that ITCZ may travel slowly over East Africa causing increase in rainfall rate and duration during MAM. The frequency of extreme wet periods may double in future
with increased CO₂ in non-coastal regions of East Africa, with larger increases towards
the equator with increase up to 11 times. This study suggests that the increased rainfall
rate and frequency of extreme rainfall events over East Africa may result from increase in
atmospheric CO₂, and may contribute to higher incidences of the above said diseases in
future across East Africa.
Chapter 3: Data and Methods

3.1 Data Description

Table 3.1 shows the source, resolution and time period available for each data set used. The description of data is written in the following sections.

Table 3.1: Data sources, resolution and time period

<table>
<thead>
<tr>
<th>Data Name</th>
<th>Source</th>
<th>Resolution</th>
<th>Time Period</th>
</tr>
</thead>
<tbody>
<tr>
<td>Terrestrial water budget data</td>
<td>Climate Research Center, University of Delaware</td>
<td>0.5° x 0.5° lat/lon</td>
<td>1950-1999</td>
</tr>
<tr>
<td>Atmospheric data</td>
<td>NCEP/NCAR reanalysis</td>
<td>209 km x 209 km</td>
<td>1942 onwards</td>
</tr>
<tr>
<td>SST data</td>
<td>UK Met Office, Hadley Center</td>
<td>5° x 5° lat/lon</td>
<td>1871 onwards</td>
</tr>
<tr>
<td>Aerosol Index data</td>
<td>NASA/GSFC Ozone Processing Team</td>
<td>1° x 1.5° lat/lon</td>
<td>1980-2001</td>
</tr>
<tr>
<td>NDVI data</td>
<td>NOAA/NASA Pathfinder AVHRR Land program</td>
<td>1° x 1° lat/lon</td>
<td>1982-1992</td>
</tr>
<tr>
<td>PDSI data</td>
<td>Climate and Global Dynamics Division of NCAR</td>
<td>2.5° x 2.5° lat/lon</td>
<td>1860-1995</td>
</tr>
<tr>
<td>QBO Index data</td>
<td>NOAA-CIRES Climate Diagnostics Center</td>
<td>-</td>
<td>1948-2004</td>
</tr>
<tr>
<td>NAO Index data</td>
<td>Polar Research Center, Ohio State University</td>
<td>-</td>
<td>1874-1999</td>
</tr>
<tr>
<td>Nino 3.4 Index data</td>
<td>KAPLAN and Climate Prediction Center</td>
<td>-</td>
<td>1856-2004</td>
</tr>
</tbody>
</table>

3.1.1 Terrestrial Water Budget Data

Monthly mean terrestrial water budget variables (Table 3.2) were obtained from the Climate Research Center, University of Delaware. To prepare this data archive, the Global Historical Climatology Network (GHCN version 2) and Legates and Willmott's (1990) station records of monthly and annual mean total precipitation have been used.

Time series of gridded monthly total potential evapotranspiration and total precipitation are obtained from the ‘Terrestrial Water Budget Data Archive’ and
Table 3.2: Terrestrial water budget variables

<table>
<thead>
<tr>
<th>Variable Name</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Precipitation (P)</td>
<td>Mm</td>
</tr>
<tr>
<td>Potential Evapotranspiration (PET)</td>
<td>Mm</td>
</tr>
<tr>
<td>Actual Evapotranspiration (ET)</td>
<td>Mm</td>
</tr>
<tr>
<td>Surplus (S)</td>
<td>Mm</td>
</tr>
<tr>
<td>Deficit (DEF)</td>
<td>Mm</td>
</tr>
<tr>
<td>Mid-monthly Soil Moisture Depth (W)</td>
<td>Mm</td>
</tr>
</tbody>
</table>

‘Terrestrial Air Temperature and Precipitation: Monthly and Annual Time Series (Version 1.02) Archive’. Other climatic water budget variables are estimated based on the semiempirical relationship between the observed monthly total precipitation and estimated potential evapotranspiration, according to the Thornthwaite method. Other water budget variables such as snow melt and mid-monthly snow cover are not used in this study, as these variables do not significantly affect the terrestrial water budget of the study area.

Station averages of precipitation are interpolated to a 0.5-degree x 0.5-degree of latitude/longitude grid, where the grid nodes are centered on 0.25 degree. For spatial interpolation, a spherical version of Shepard's distance-weighting method (Shepard, 1968; Willmott et al., 1985) is used. An average of 20 nearby stations is considered that influence a grid-node estimate, which has resulted in smaller cross validation errors and visually more realistic air-temperature and precipitation fields. A neighbor finding algorithm based on spherical distance is also used for data interpolation. To increase the accuracy of spatially interpolated average air temperature, an average air temperature lapse rate is used which incorporates the station height. 6.0 °C km⁻¹ is used as the average environmental lapse rate. Monthly precipitation fields are estimated using
Climatologically Aided Interpolation (CAI) (Willmott and Robeson, 1995). To increase the accuracy of spatially interpolating time series of monthly climate variables, interpolated estimates of available climatology are used. To do this, CAI is performed on the available monthly time series of climate variables at individual stations to obtain the difference between average climatology and the station value. Traditional methods of interpolation are then used on these station differences to compute the gridded difference field. This field is then added to the interpolated estimates of climatology at each grid point.

Potential evapotranspiration is calculated using air temperature. Potential evapotranspiration is climatic water demand. Thornthwaite’s expression (Thornthwaite, 1948) for the relation between potential transpiration and air temperature is

\[ E^o = 1.6 \left(10 \times \frac{T}{I}\right)^a, \text{ for } 0 \leq T < 26.5^\circ C \]

Where, \( E^o \) = monthly unadjusted potential evapotranspiration in cm,
\( T \) = mean monthly temperature in degree Celsius,
\( I \) = annual heat index

Annual heat index I is the sum of the 12 monthly heat indices \( i \), where,

\[ i = \left(\frac{T}{5.0}\right)^{1.514} \]

And,

\[ a = (6.75 \times 10^{-7} I^3) - (7.71 \times 10^{-5} I^2) + (1.79 \times 10^{-2} I) + 0.49 \]

For other temperature ranges, monthly, unadjusted potential evapotranspiration is

\[ E^o \text{ (in mm/month)} = \begin{cases} 0, & \text{for } T < 0^\circ C \\ -415.85 + 32.24T - 0.43T^2, & \text{for } T \geq 26.5^\circ C \end{cases} \]
An adjustment is made to account the variable number of days in month and variable hours of daylight in days. Unadjusted potential evapotranspiration $E^0$ is adjusted to $E$ as

$$E\ (\text{mm/month}) = E^0 \left[\left(\frac{\varrho}{30}\right) \left(\frac{h}{12}\right)\right]$$

Where, $\varrho = \text{length of the month in days}$

$h = \text{duration of daylight in hours on the fifteenth of the month.}$

To compute the monthly mean of potential evapotranspiration for $n^{th}$ month, $I$ (annual heat index) is calculated using monthly heat indices for the $(n-11)$ to $(n-1)$ months and for the $n^{th}$ month.

Thornthwaite’s soil moisture budget (Willmott et al., 1985) is governed by the following expression

$$\frac{\partial w}{\partial t} = P - E' - S$$

where, $w = \text{moisture available in root zone}$

$P = \text{rate of precipitation (mm/hour)}$

$E' = \text{rate of evapotranspiration (mm/hour)}$

$S = \text{surplus (mm/hour)}$

Evapotranspiration depends upon rates of precipitation and potential evapotranspiration and on the available soil moisture. Thus,

$$E' = P + \beta(w, w^*) \left(E^0 \ (T', h) - P\right), \quad \text{for } P < E^0 \ (T', h)$$

$$= E^0 \ (T', h), \quad \text{for } P \geq E^0 \ (T', h)$$

Where, $w^* = \text{soil moisture storage capacity}$

$T'$ = daily average surface air temperature
B = a function that relates \([\frac{(E' - P)}{(E^o - P)}]\) to \((w/w^*)\) (Willmott et al., 1985)

According to Thornthwaite's method, S includes both surface and subsurface runoff from root zone. Thus the expression of surplus is,

\[
S = P - [E' + (w^* - w)], \quad \text{for } P > [E' + (w^* - w)]
\]
\[
= 0, \quad \text{for } P \leq [E' + (w^* - w)]
\]

In the preparation of this terrestrial water budget data set variables \(w^*\) has been held constant at 150 mm.

The data of precipitation and potential evapotranspiration were used to calculate Aridity Index over East Africa (Section 2.2.1) as a ratio of mean seasonal precipitation and mean seasonal potential evapotranspiration (modified from the calculation method of Aridity Index after Budyko, 1974 (detailed description in section 3.2.2.1). The principal components analysis was performed over Aridity Index thus obtained as explained in methods section. The other terrestrial water budget variables were correlated to time series of first principal components of Aridity Index in order to investigate the effect of variability in these water budget variables on the Aridity Index.

3.1.2 Atmospheric Data

The data set of atmospheric hydroclimatic variables used in this study is the NCEP/NCAR reanalysis data described by Kalnay et al. (1996). The objective of the NCEP/NCAR Reanalysis Project is to reanalyze historical data using state-of-the-art models and perform data assimilation using past data and analysis/forecast systems. The data are collected from variety of sources and by different agencies including National Aeronautics and Space Administration/Goddard Laboratory for Atmosphere
(NASA/GLA), the European Center for Medium-Range Weather Forecasts (ECMWF), and the Center for Ocean-Land-Atmosphere Interactions (COLA).

The different sources rely on direct observations collected worldwide at meteorological observatories, rawinsondes, ships, buoys, aircrafts and satellites. Conventional meteorological observations collected at meteorological stations are using conventional instruments. A wider range of variables collected at various stations globally, such as, temperature, precipitation, surface pressure, humidity, wind speed, and direction, cloud cover, snow depth, visibility, solar radiation and current weather (Shea et al., 1995), where measurements at various levels of the atmosphere are recorded. Upper-air observations are made by radiosondes or rawinsondes. Ocean data are provided by ships and by moored and drifting buoys for the variables such as water temperature, salinity, dissolved oxygen, various nutrients and tracers (Shea et al., 1995) at ocean surface and/or at the vertical levels below ocean surface. Satellites data are used for a broad range of geophysical quantities such as the vertical distribution of atmospheric temperature and moisture, clouds, winds, atmospheric gases and SSTs (Shea et al., 1995). The reanalysis data are produced from 1948 onward.

The climate model used for the reanalysis project is a global spectral model with horizontal spectral resolution of T62 (209 km) and with 28 vertical levels. The model has 5 levels in the boundary layer and about 7 levels above 100 hPa with the lowest model level at about 5 hPa from the surface, and the top of the model atmosphere is at about 3 hPa. This vertical structure was chosen to reasonably resolve the boundary layer as well as to make stratospheric analysis without getting affected by boundary conditions. The model includes parameterizations schemes for major physical processes, such as
convection, large scale precipitation, shallow convection, gravity wave drag, radiation with diurnal cycle and interaction with clouds, boundary layer physics, an interactive surface hydrology, and vertical and horizontal diffusion processes (Kanamitsu, 1989; Kanamitsu et al., 1991).

Quality control is performed at all the steps for data assimilation module. A preprocessor is used for the data entering into the data assimilation module to reformat the data coming from many different sources, while the analysis output is monitored by quality control monitoring system. The model used in data assimilation module is the T62 model with 28 vertical levels. The data assimilation system includes spectral statistical interpolation analysis and improved error statistics. The complex quality control of rawinsonde data provides certain atmospheric parameters. Final screening is obtained using optimal interpolation quality control, developed for all the observations entering data assimilation system, which detects and withholds the data with gross errors from entering into the system.

The classification of reanalysis gridded data fields into four classes (A, B, C, and D) is based on the relative influence of observational data and model output (Kalnay et al.; 1996), where Class A analysis variable is strongly influenced by observed data, Class B analysis variable is influenced strongly by model - although there are observational data directly affecting the value of the variable, Class C analysis variables are derived solely from model fields and Class D reanalysis fields are obtained from climatological values and not depend on model output. The reliability of the variables is not uniform. Class A variables are the most reliable variables.

The atmospheric data variables used in this study and their reliability are
described in Table 2.1. This data set was used for global and regional studies due to its reliability as well as the availability. Though data were obtained for all mandatory pressure levels, data at only surface level were used in this study. Upward longwave radiation flux (OLR) is calculated for the top of the atmosphere and precipitable water (PWAT) is calculated as the vertically-integrated specific humidity from the surface to the 300 hPa pressure level. Most of the variables obtained from the reanalysis data are of Class A (such as sea level pressure, geopotential height and zonal and meridional winds) or Class B (such as air temperature, vertical velocity, specific humidity, PWAT and dew point depression) and few are of Class C (such as rainfall rate and OLR).

**Table 3.3** Reanalysis data variables used in this study

<table>
<thead>
<tr>
<th>Variable Name</th>
<th>Unit</th>
<th>Class</th>
<th>Levels Available</th>
</tr>
</thead>
<tbody>
<tr>
<td>Air temperature (T)</td>
<td>K</td>
<td>B</td>
<td>All pressure levels</td>
</tr>
<tr>
<td>Zonal wind (U)</td>
<td>m s⁻¹</td>
<td>A</td>
<td>17</td>
</tr>
<tr>
<td>Meridional wind (V)</td>
<td>m/s⁻¹</td>
<td>A</td>
<td>17</td>
</tr>
<tr>
<td>Sea level pressure (slp)</td>
<td>Pa</td>
<td>A</td>
<td>Surface</td>
</tr>
<tr>
<td>Geopotential height</td>
<td>Gpm</td>
<td>A</td>
<td>17</td>
</tr>
<tr>
<td>Vertical velocity (VVEL)</td>
<td>m s⁻¹</td>
<td>B</td>
<td>All pressure levels</td>
</tr>
<tr>
<td>Specific humidity</td>
<td>kg kg⁻¹</td>
<td>B</td>
<td>1000 – 300 hPa</td>
</tr>
<tr>
<td>Precipitable water</td>
<td>kg m⁻²</td>
<td>B</td>
<td>Surface</td>
</tr>
<tr>
<td>Rainfall rate</td>
<td>Kg m⁻² s⁻¹</td>
<td>C</td>
<td>Surface</td>
</tr>
<tr>
<td>Dew point depression</td>
<td>K</td>
<td>B</td>
<td>850 hPa</td>
</tr>
<tr>
<td>Upward Longwave Radiation flux (OLR)</td>
<td>W m⁻²</td>
<td>C</td>
<td>Surface</td>
</tr>
</tbody>
</table>

The data for precipitation and PET used in this study are not reanalysis fields because of the reduced reliability of the data. Instead, those variables were obtained from Climate Research Center, University of Delaware. The reanalysis data obtained were used for two purposes, (a) to describe the annual climate over East Africa and (b) to compute the correlation of those fields with the time series of first principal components.
(See methods section) of Aridity Index to investigate the association between the hydroclimatology of East Africa and the changes in aridity over East Africa.

3.1.3 Sea Surface Temperature (SST) Data

The SST data used in the present study are obtained from the UK Met Office Hadley Center. This dataset of sea ice and SST data set, HadISST1, is developed at UK Met Office Hadley Center for Climate Prediction and Research. The purpose of creating this data is to force atmospheric global circulation models in the simulation of recent climate and to evaluate coupled atmosphere-ocean models to improve the understanding of natural and human-induced climatic variation and to allow evaluation of model performance (Rayner et al., 2003). HadISST1 data are available from 1871 onwards and for 1° latitude/longitude resolution. The SST data are taken from the Met Office Marine Data Bank (MDB). From 1982 onwards data received through the Global Telecommunications System (GTS) is also included. If no MDB data are available, monthly median SSTs for 1871-1995 from the Comprehensive Ocean-Atmosphere Data Set (COADS) were used (http://hadobs.metoffice.com/hadisst/). The sea ice data are taken from a variety of sources including digitized sea ice charts and passive microwave retrievals (http://hadobs.metoffice.com/hadisst/).

In HadISST1, an Empirical Orthogonal Functions (EOFs) based Reduced Space Optimal Interpolation (RSOI) (Kaplan et al., 1997) is used to reconstruct the broad-scale fields of SST and to extend the analysis over the most of the data sparse ocean region. This reanalysis data uses in situ SST observations and remotely sensed satellite SST data.

The in situ data are adjusted for the biases such as ocean water depth from which the water sample is collected; non-insulated or partially-insulated fabric of the buckets
used to collect the water sample, their exposure to the wind on the deck, and fixed or drifting buoys, especially for the data obtained before 1941. The reconstructed SST data using RSOI is combined with the refined gridded SSTs to capture the variability of SST anomaly with the greater accuracy. Satellites cover almost the entire ocean region and thus the satellite SST data could be used especially for the not-monitored ocean areas and along with the observed data for the monitored ocean region. Monthly SSTs from the Advanced Very High Resolution Radiometer (AVHRR) are used for January 1982 onwards, which are obtained from the operational US National Oceanographic and Atmospheric Administration (NOAA) satellite- born AVHRR instruments. Only night AVHRR SST data are used in HadISST1 because of its easier biases retrievals. The data are adjusted for biases such as unrepresentative atmospheric conditions, presence of warm/cold top clouds, sea ice, stratospheric aerosols resulting from volcanic irruptions, tropospheric aerosols, and instrument calibration error. Combination of in situ SSTs and bias-adjusted AVHRR SSTs are used 1982 onwards.

The SST data were used to compute climatic indices explained in section 2.4. The climatic indices were then correlated with the Aridity Index as well as with the first principal components of Aridity Index and with the water budget variables seasonally over East Africa.

3.1.4 Aerosol Index Data

The source for the aerosol data is NASA/GSFC (Goddard Space Flight Center) Ozone Processing Team. The data are collected using Earth probe TOMS (Total Ozone Mapping Spectrometer). The instrument ‘Earth Probe TOMS’ on NASA spacecraft

* Data for aerosols, NDVI, PDSI and QBO index were downloaded from KNMI climate explorer website (http://climexp.knmi.nl/selectfield.cgi?someone@somewhere).
AURA allows the observation of aerosols as the particles cross the land-sea boundary. It is possible to observe phenomena such as desert dust storms, forest fires and biomass burning using the TOMS data. The data are available in the form of an aerosol index (AI). The TOMS aerosol index is a measure of the difference between the backscatter ultraviolet (UV) radiation wavelength dependence from an atmosphere containing aerosols and the backscatter UV radiation wavelength dependence from a pure molecular atmosphere$^1$. Mie scattering, Rayleigh scattering, and absorption occur in atmosphere containing aerosols, while, pure Rayleigh scattering occurs in pure molecular atmosphere. Quantitative definition of aerosol index (AI)$^1$ is given by

$$ AI = -100 \log_{10} \left[ \frac{I_{331}^{\text{Meas}}}{I_{360}^{\text{Calc}}} \right] $$

where, $I_{331}^{\text{Meas}} = \text{measured 331 nm EP-TOMS radiance}$

$I_{360}^{\text{Calc}} = \text{calculated 360 nm EP-TOMS radiance for a Rayleigh atmosphere}$

Under most conditions, the AI is positive for absorbing aerosols and negative for non-absorbing (scattering) aerosols$^1$.

The TOMS AI data$^1$ is available for the time period of 1980 to 2001. These data are used to calculate the correlation with first principal components of Aridity Index (EOF1) and with the climatic indices (see section 3.2.2) in order to examine the possible connection between aerosol flow and precipitation and aridity over East Africa.

$^1$ Data for aerosols, NDVI, PDSI and QBO index were downloaded from KNMI climate explorer website (http://climexp.knmi.nl/selectfield.cgi?someone@somewhere).

$^{[1]}$ Reference Website: http://toms.gsfc.nasa.gov/aerosols/AI_definition/ai_def.html
3.1.5 NDVI Data*

The NDVI (Normalized Difference Vegetation Index) data set has been produced as part of the NOAA/NASA Pathfinder AVHRR Land (PAL) program. The data is derived from the visible (0.58 µm to 0.68 µm) and near-infrared channel (0.73 µm to 1.10 µm) reflectance and is available in 8 km and 1-degree resolutions for the time period of 1982-1992. The value of NDVI ranges between –1 and +1. NDVI is a measure of greenness of a region. The data with 1-degree resolution is used in the present study.

AVHRR sensor measures emitted and reflected radiation in five channels of the electromagnetic spectrum, which covers visible, near infrared and infrared spectral regions as described in Table 3.4. Chlorophyll considerably absorbs incoming radiation in visible wavelength (measured by channel 1), while spongy mesophyll leaf structure considerably reflects radiation in visible wavelengths. The ratio of transform (dividing one band by the other) measures this contrast between responses of the two bands. The NDVI is one such ratio transform which highly correlates with vegetation parameters such as green-leaf biomass and green-leaf area \(^2\). NDVI hence is used for vegetation discrimination (Justice \textit{et al.}; 1985).

The NDVI is calculated as

\[
NDVI = \frac{CH_5 - CH_1}{CH_5 + CH_1}
\]

The Pathfinder data are processed with automated quality control for validating the data and for consistency in fields such as date and satellite or scan times \(^2\). The dataset

---

* Data for aerosols, NDVI, PDSI and QBO index were downloaded from KNMI climate explorer website (http://climexp.knmi.nl/selectfield.cgi?someone@somewhere).
\(^2\) Reference website for NDVI data: http://daac.gsfc.nasa.gov/CAMPAIGN_DOCS/FTP_SITE/readmes/pal.html
contains errors in the computation of solar zenith angle, which affects reflectance of channel 1 and channel 2 [2]. Channel 1 is used in the computation of NDVI. NDVI data is used to calculate its correlations with the first principal components of Aridity Index (EOF1), water budget variables and with climatic indices in order to find the relation between vegetation cover over East Africa and other hydrological and climatic variables.

**Table 3.4:** Channels, wavelengths and spectral regions measured by the AVHRR sensor on NOAA-7, NOAA-9, and NOAA-11 satellites [2] (Kidwell, 1991).

<table>
<thead>
<tr>
<th>Channel</th>
<th>Spectral Region</th>
<th>Wavelength Band</th>
<th>Purpose</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Visible</td>
<td>0.58 to 0.68 µm</td>
<td>Daytime cloud and surface mapping</td>
</tr>
<tr>
<td>2</td>
<td>Near-Infrared</td>
<td>0.725 to 1.1 µm</td>
<td>Surface water delineation and vegetation cover mapping</td>
</tr>
<tr>
<td>3</td>
<td>Mid-Infrared</td>
<td>3.55 to 3.93 µm</td>
<td>SST and nighttime cloud mapping</td>
</tr>
<tr>
<td>4</td>
<td>Thermal Infrared</td>
<td>10.5 to 11.5 µm</td>
<td>Surface temperature and day and night cloud mapping</td>
</tr>
<tr>
<td>5</td>
<td>Thermal Infrared</td>
<td>11.5 to 12.5 µm</td>
<td>Surface temperature mapping</td>
</tr>
</tbody>
</table>

### 3.1.6 QBO Index Data*

The Quasi-Biennial Oscillation (QBO) Index is obtained from NOAA-CIRES Climate Diagnostics Center’s Climate Indices dataset [3]. QBO index data are calculated at the Climate Diagnostic Center (CDC). QBO Index data are calculated as the zonal average of the 30 mb zonal wind at the equator, where the zonal wind data are obtained from the NCEP/NCAR Reanalysis. The QBO Index data are available from 1948 – 2004. Correlations of QBO index with climate indices (such as ENSO Index, IOD Index, NAO

---

* Data for aerosols, NDVI, PDSI and QBO index were downloaded from KNMI climate explorer website ([http://climexp.knmi.nl/selectfield.cgi?someone@somewhere](http://climexp.knmi.nl/selectfield.cgi?someone@somewhere)).
[3] Reference website for QBO index data: [http://climexp.knmi.nl/getindices.cgi?FUBData/qbo+QBO+i+someone@somewhere](http://climexp.knmi.nl/getindices.cgi?FUBData/qbo+QBO+i+someone@somewhere)
Index and TASD Index), with water budget data variables, with Aridity Index and with first principal components (EOF1) of Aridity Index, in order to investigate the effect of QBO on hydroclimatology of East Africa as well as on the parameters affecting the hydroclimatology of East Africa.

3.1.7 PDSI Data*

The Palmer Drought Severity Index (PDSI) was developed by W.C. Palmer in 1965 to measure departure of moisture supply (Palmer, 1965). The PDSI provides standardized measurements of moisture conditions. PDSI is a drought index responding to abnormally dry or wet weather conditions. The drought index developed by Palmer was only for the USA. Dai (1998) have developed the PDSI for use globally.

The PDSI data used here were obtained from Climate and Global Dynamics Division of NCAR (Dai PDSI data)[4]. The PDSI is calculated using monthly air temperature (T) (from Hansen and Lebedeff, 1987 and the updates; Jones and Moberg, 2003 and updates) and precipitation (P) (from Dai et al., 1997; Chen et al., 2002) based on moisture balance of the ground (Dai et al., 1998). Anthropogenic changes in surface conditions that alter runoff changes or drainage catchments are not considered in calculation of PDSI (Dai et al., 1998) and Soil-holding capacity was fixed to climatological maps from Webb et al., 1993. The PDSI value generally ranges between –7.0 and +7.0, where –7.0 is considered as dry and +7.0 as wet. The adjustments are made to correct station record-derived T and P data (Dai et al., 1997). The PADI data is available for the time period of 1860 – 1995 for 2.5° x 2.5° latitude/longitude resolution.

In the present study, PDSI is been correlated with EOF1, water budget variables, with aerosols, NDVI, PDSI and QBO index were downloaded from KNMI climate explorer website (http://climexp.knmi.nl/selectfield.cgi?someone@somewhere).

over East Africa and with climatic indices.

3.2 Methods

The objective of present study is to determine the meteorological association of aridity over East Africa with regional hydroclimatic variables as well as to find global teleconnections affecting spatial distribution of aridity over East Africa. The theoretical basis of techniques used and the methods used are described below.

The Aridity Index was calculated over the East Africa region for the four seasons, which were decided based on rainfall receipt over East Africa. The Aridity Index over a region represents the mean state of climate over that region as it is the ratio of mean (annual or seasonal) precipitation (supply) and the mean (annual or seasonal) PET (atmospheric water demand). Principal components analysis (PCA) was performed on Aridity Index in order to identify characteristic modes of temporal variability in the Aridity Index across East Africa. The time series of the first principal components of Aridity Index was further correlated with regional mean seasonal hydroclimatic variables to identify the meteorological association in the intra-annual variability of aridity over East Africa. The time series of hydroclimatic variables was plotted against the time series of the climatic indices to view the practical significance of the range of hydroclimatic variables that corresponds to the climatic indices.

3.2.1 Area and Seasons

The East Africa region is considered as a region between 12°N-20°S and 20°E-50°E in present study. This area includes The Horn of Africa and tropical East Africa. Four seasons were decided on the basis of the precipitation receipt by different regions of
East Africa in an average year. Indeje et al. (2000) have considered same seasons for East Africa region, although their boundaries of East Africa do not exactly match.

Over most of the East Africa, annual rainfall has a bimodal distribution. The long rains of March to May (MAM) and short rains of October to December (OND) are the transition periods between the southeast (from June to September (JJAS)) and the northeast (January-February (JF)) monsoons over East Africa. This division of seasons allows incorporating changes in amount of rainfall receipt as well as the changes in the circulation features over East Africa during an average year. The figures showing annual climate as well as the results are arranged seasonally.

3.2.2 Definitions of Indices

3.2.2.1 Aridity Index

Aridity differs from drought. Drought is normal, recurring short-term feature of climate, which can occur virtually in all climatic regions. Drought is related to climate variability of the region. Aridity on the other hand, is permanent or long-term and characteristic feature of climate over low rainfall regions.

The Aridity Index is defined as a ratio of mean annual precipitation (P) and the mean annual evapotranspiration (PET), after Budyko (1974). PET is the amount of moisture that would be removed from a given land area by the methods of evaporation and transpiration. PET is defined as the amount of water evaporated (both as transpiration and evaporation from the soil) from an area of continuous uniform vegetation that covers the whole ground and that is well-supplied with water. Further more, it is the amount of moisture that, if available, would be removed from a given land area by evaporation and is expressed as the rate of latent heat transfer per square centimeter or depth of water
PET is a function of soil moisture, surface air temperature and humidity. Temperature and humidity influence the rate of evaporation by affecting the moisture holding capacity of atmosphere. For a region annual mean PET is governed by the demand of available energy (net radiation) and supply of precipitation (Arora, 2002). The Aridity Index being ratio of precipitation to PET, measures the water balance of a region.

Aridity of the region may change with change in precipitation and/or in PET. Temporal and spatial intra-annual and interannual variability in these variables may affect the variability in Aridity Index. To accommodate the seasonal changes in these variables, Aridity Index in this study is calculated seasonally as the ratio of mean seasonal precipitation to mean seasonal PET. Though the means are calculated seasonally instead of annually, the method of calculations essentially is the same. Aridity Index is generally used to determine the climatic regimes or aridity zones as described in Table 3.6. The smaller the Aridity Index; the higher the aridity over the region and vice versa.

Table 3.6: Aridity Index and associated aridity zones

<table>
<thead>
<tr>
<th>Aridity Index (ArI)</th>
<th>Aridity zone</th>
</tr>
</thead>
<tbody>
<tr>
<td>ArI &lt; 0.05</td>
<td>Hyper Arid</td>
</tr>
<tr>
<td>0.05 &lt; ArI &gt; 0.2</td>
<td>Arid</td>
</tr>
<tr>
<td>0.2 ≤ ArI &gt; 0.5</td>
<td>Semi-Arid</td>
</tr>
<tr>
<td>0.5 ≤ ArI &gt; 0.65</td>
<td>Dry Subhumid</td>
</tr>
<tr>
<td>ArI &gt; 0.65</td>
<td>Moist Subhumid</td>
</tr>
</tbody>
</table>

3.2.2.2 El Niño (NINO) Index

The El Niño Index for region Niño 3.4 is used in this study. Niño 3.4 region is the region between 5°S-5°N and 170°W-120°W (Figure 3.1) and the NINO 3.4 index represents the SST anomalies in the Niño 3.4 region. While NINO 3 region has largest
SST variability, NINO 3.4 region also has large SST variability on El Niño time scales. The units of the Niño 3.4 Index are degrees Celsius. The SST in this region has the strongest effect on shifting rainfall in the western Pacific (Trenberth, 1997) and in turn, shifting the location of rainfall from the western to central Pacific modifies greatly the location of the heating that drives the majority of the global atmospheric circulation.

![Figure 3.1](image.png)

**Figure 3.1:** Nino 3.4 region in equatorial Pacific. (Modified from website of University of Arizona, Institute for the Study of Planet Earth.)

### 3.2.2.3 Indian Ocean Dipole Mode (IODM) Index

The IODM is a climate mode that occurs interannually in the tropical parts of the Indian Ocean. Indian Ocean Dipole index is calculated as the difference between SSTs averaged over western and eastern Indian Ocean. The region between 50°E - 70°E and 10°S - 10°N is considered to calculate average SST for western region, while the region between 90°E - 110°E and 10°S – equator is considered to calculate average SST for eastern region (Saji *et al.*, 1999).

\[
\text{IOD}_{W} = \text{average SST (50°E - 70°E, 10°S - 10°N)}
\]

\[
\text{IOD}_{E} = \text{average SST (90°E - 110°E and 10°S – 0°)}
\]
IOD Index = IOD$_W$ - IOD$_E$.

3.2.2.4 North Atlantic Oscillation (NAO) Index$^5$

The NAO is a large-scale alternation of atmospheric mass with centers of action near the Icelandic Low and the Azores High. The NAO Index is calculated as the difference in sea level pressure between Ponta Delgada, Azores (38°N, 26°W) and Akureyri, Iceland (66°N, 18°W). NAO Index was devised by Sir Gilbert Walker in the 1920s. NAO index he calculated incorporated pressure, air temperatures, and precipitation at several Atlantic coastal stations. The NAO index now used is a simplification of the original NAO Index, which incorporates surface pressure difference only. The data are available for the 1874 – 1999 periods. Though the NAO Index also could be calculated using the pressure data from other nearby regions, the NAO Index calculated using Azores – Iceland data produces very realistic estimates of the magnitude of the NAO dipole during non-winter seasons$^5$. The subtropical center at the Azores lies in the eastern Atlantic Ocean. The Azores data better represents the strength of the Atlantic westerlies (Hurrell and van Loon, 1997). The data is obtained as ‘Rogers (1984) updated NAO Index’ from Polar Research Center, Ohio State University.

3.2.2.5 Tropical Atlantic SST Dipole (TASD) Index

TASD is the dominant decadal-scale SST pattern in tropical Atlantic and an out of phase relationship between the north and south of the equator in the SST anomalies Tropical Atlantic SST Dipole Index is calculated as the difference between SSTs averaged over Northern and Southern tropical Atlantic Ocean. The average SST for northern tropical Atlantic are calculated over the region between 5°N-20°N, 40°W-20°W

$^5$Reference website http://polarmet.mps.ohio-state.edu/NAO.
and the average SST for southern tropical Atlantic are calculated over the region between 5°S-5°S, 15°W-5°E (Chang et al., 1997).

\[
\text{TNA} = \text{average SST (5°N-20°N, 40°W-20°W)}
\]

\[
\text{TSA} = \text{average SST (5°S-5°S, 15°W-5°E)}
\]

\[
\text{TASD Index} = \text{TNA} - \text{TSA}
\]

### 3.2.3 Principal Components Analysis

PCA is also known as Empirical Orthogonal Functions (EOF) analysis. The principal components are orthogonal to each other, hence are called empirical (-arising from data) orthogonal (-uncorrelated over space) functions. PCA is a data transformation technique.

Singular value decomposition (SVD) stands at the base of PCA (Preisendorfer, 1988). PCA was used independently for analysis in different study areas at different places at different times. The first clear geometric visualization of PCA was first given by Pearson in 1901. Its current, generalized, and relatively abstract appearance is due to the efforts of Loève (1963). The first application of PCA in meteorology was in Massachusetts Institute of Technology by G.P. Wadsworth and his colleagues J.G. Bryan and C.H. Gordon (Preisendorfer, 1988).

PCA enables fields of highly correlated data to be represented adequately by a small number of orthogonal functions and corresponding orthogonal coefficients, which account for much of the variance in their spatial and temporal variability (Indeje et al., 2000). Variance is a measure of dispersion of values around mean. If the variables are associated with each other, they have covariance. The variance of a whole data set is a sum of all individual variances. In PCA, the data are transformed in such a way as to
describe the total variance with the same number of axes. The first principal components accounts for the maximum variance, while other principal components explain the remaining variance. The principal components are uncorrelated to each other.

PCA focuses on explaining the total variation in the observed variables on the basis of the maximum variance properties of principal components (Dunteman, 1989). Principal components represent the variation in the original variables. In PCA, geometrical representation of first principal components is the line of closest fit to the n observations in p dimensional variable space, while the second principal components is the line of closest fit to the residuals from the first principal components.

For example, if a set of \( p \) observations, which could be characterized as the \( p \) dimensional vector \( (x_1, x_2, \ldots, x_p) \) can be linearly transformed into principal components \( y_i \)'s, the first principal components \( y_1 \) is a linear combination of \( x_1, x_2, \ldots, x_p \). Thus,

\[
y_1 = a_{11}x_1 + a_{12}x_2 + \ldots + a_{1p}x_p,
\]

Where, \( a_1, a_2, \ldots, a_p \) are weights. In PCA, the weights are determined mathematically to maximize the sum of the squared correlations of the principal components with original variables (Dunteman, 1989). Thus,

\[
y_1 = \sum_{i=1}^{p} a_{1i} x_i
\]

Variance of \( y_1 \) is maximized given that the sum of the squared weights is equal to one. Mathematically,

\[
\sum_{i=1}^{p} a_{1i}^2 = 1
\]
Similarly, the second principal components $y_2$ is

$$ y_2 = a_{21}x_1 + a_{22}x_2 + \ldots + a_{2p}x_p = \sum_{i=1}^{p} a_{2i}x_i, $$

and,

$$ \sum_{i=1}^{p} a_{2i}^2 = 1 $$

This procedure continues until the number of components equals to the number of variables. If PCA is performed on a random vector of $p$ variables $(x_1, x_2, \ldots, x_p)$, which can be considered as a point in $p$ dimensional space, where $(x_1, x_2, \ldots, x_p)$ represent coordinate axes, PCA determines a lower dimensional ($n$) space ($n < p$), that explains together all the variance in original points. Thus the original $p$ coordinates could be represented on this best fitting subspace with $n$ coordinate axes (Dunteman, 1989). Only first $k$ components could be taken into consideration for analysis if they together explain most of the total variance. PCA yields a spatial as well as temporal distribution of variable/s.

Various methods have been suggested to determine number of principal components to be retained. Kaiser (1960) recommended that those principal components of correlation matrix with latent roots less than one ($\lambda < 1.0$) could be dropped, without any harm to the analysis; while Jolliffe (1972) suggested dropping of principal components with latent roots less than 0.7 ($\lambda < 0.7$). A “scree” plot also could be used to determine number of principal components to be retained (Cattell, 1966). Latent roots are plotted in scree plot and a point $k$ is determined such that the line joining the points is steep to the left of point $k$, while is not steep to the right of point $k$ (Dunteman, 1989).
Then k principal components are retained (Dunteman, 1989).

Rotation is suggested in several cases where specific problems may afflict the PCA results. Problems such as domain shape dependence, numerical instability and observed Buell-sequences (Legates, 1991 and 1993; Richman, 1993) limit the utility of PCA results and merit rotation procedures such as Varimax, or perhaps an oblique rotation. However, rotation is often applied “automatically” by users and may be applied incorrectly when PCA results are not afflicted with the above problems. In this study the aim of the PCA is to produce the time series scores associated with EOF1; the pattern of variability exhibited by the associated spatial field (the PCA loadings) is considered unimportant. Given this goal, rotation is not applied and even if it were, the time series scores would not change; only the spatial pattern (the loadings) would be affected. Spatial patterns of loadings are shown, but not used or interpreted in this study.

In the present study, PCA was performed on the Aridity Index calculated over East Africa region. The first principal component of Aridity Index (EOF1) was used for correlation analysis. EOF1 thus obtained was further correlated with the terrestrial water budget variables (other than P and PET) obtained from Climate Research Center, University of Delaware and with the number of atmospheric variables obtained from NCEP/NCAR reanalysis dataset as well as with the climatic indices for all four seasons as described in next section.

3.2.4 Correlation Analysis and Correlation Maps

3.2.4.1 Correlation Analysis

Correlation analysis is a statistical tool used to measure the degree of linear association between two variables. Sir Francis Galton is considered to be the first user of
correlation analysis in 1850-1852. An English scientist, Karl Pearson, provided the mathematical framework for correlation and regression analysis and developed the correlation coefficient in 1891-1892. Another English scientist, Weldon, introduced negative correlation coefficient in 1892.

The correlation coefficient is measure of the strength of the linear association between two variables. Correlation coefficients are always expressed by numbers between -1 and +1. A correlation coefficient of +1 indicates that two variables are directly proportional and perfectly related in a positive linear sense, while a correlation coefficient of -1 indicates that two variables are inversely proportional and perfectly related in a negative linear sense, and a correlation coefficient of 0 indicates that there is no linear relationship between the two variables. Correlation analysis can not be used to establish the cause-effect relationship; rather it only indicates the extent or degree of linear relationship between the two variables.

Correlations of Aridity Index with precipitation and with PET were calculated to investigate the dominant factor among precipitation and PET in determining of the Aridity Index over a region.

3.2.4.2 Correlation Maps

Correlation fields were calculated and mapped between the EOF1 of the Aridity Index time series calculated over East Africa and various hydroclimatic and circulation variables obtained from the NCEP/NCAR reanalysis data set. The contours of equal correlation coefficient (solid contours for positive correlation coefficients and dashed contours for negative correlation coefficients) were mapped. The correlations significant at 95 percent, 99 percent and at 99.9 percent confidence levels were plotted as shades of
gray (from lighter to darker). Statistical significance of the correlation coefficients is calculated using a Student’s t-test. T statistics calculated are two-tailed and treated as such when assessing significance. Maps of the correlation coefficients and statistical significance are produced to provide a spatial depiction of the associations between the two variables. As described in Section 3.2.2.1, higher Aridity Index indicates less severe aridity and vice versa. Thus the positive correlation between the atmospheric variable and Aridity Index means that the Aridity Index (aridity) increases (decreases) with the increase in the intensity of the atmospheric variable.

The correlation analysis was also performed to investigate the association between the Aridity Index and climatic indices such as, NINO Index, IODM index, NAO Index and TASD Index and correlations were mapped as described above. Correlations of climatic indices with precipitation and PET were also plotted similarly to find the dominant factor in deciding Aridity Index over the region. Correlation coefficients were calculated for the correlations between these climatic indices and EOF1 of time series of Aridity Index over East Africa for each season.
Chapter 4: East Africa - Annual Climate

4.1 Atmospheric Circulation and Precipitation

4.1.1 Annual Atmospheric Circulation

Three major air masses converge over the East Africa region (Figure 4.1), which are westerly, southeasterly Congo airflow and southeast and northeast monsoon airflow. Both monsoons are thermally stable and are associated with subsiding air, and are therefore dry (Nicholson, 1996). Congo airflow on the other hand is humid and convergent, bringing rain (Nicholson, 1996). The other convergent airflow over East Africa is the Inter Tropical Convergence Zone (ITCZ), which separates the two monsoons. In equatorial Africa, ITCZ is distorted between its mean seasonal positions over the Atlantic and Indian Oceans (Hills, 1979). The seasonal patterns of rainfall in East Africa follow the seasonal migration patterns of the ITCZ. The ITCZ is not a zone of rainfall, but rather a zone of instability (Edwards et al., 1983). A number of factors in ITCZ can lead to a triggering of the rainfall mechanism (Edwards et al., 1983) such as convergence of airmasses. The convergence zone is determined by the convergence of inter-hemispherical monsoonal wind system, in which trade winds converge.

A southerly flow of wind in tropical Africa prevails when the Sun is in the northern hemisphere, while northerly wind prevails when the Sun is in the southern hemisphere (Nicholson, 1996). This seasonal reversal of the wind is called locally as southeast and northwest African monsoons respectively. But this monsoon circulation is actually a part of the western margins of South Asian monsoon circulation (Trewartha, 1961). Like the South Asian monsoon, it is not a result of differential heating of land and sea. Both the southeast and northwest African monsoons are of divergent and subsiding
nature over a large area of East Africa. The origin of the northwest monsoon winds is continental, over Sahara. Thus, it brings dry air over East Africa. Both the monsoon currents run largely parallel to coastline and highland edges (Nieuwolt, 1977) unable to advect much rainfall over East Africa.

Figure 4.1: Near surface airflow (arrows) and convergence zone (dashed line) over southeastern Africa (after Torrance, 1972, obtained from McHugh and Rogers, 2001).

Seasonal mean atmospheric circulation is shown in Figures 4.2.1 to 4.2.2 over Africa (vectors), along with seasonal mean sea level pressure (contours) and seasonal mean rainfall rate (in units of kg m\(^{-1}\) day\(^{-1}\), in shades of grey) for JF, MAM, JJAS, and OND, respectively. The anticyclonic flow generated from the subtropical high-pressure areas over the southern Atlantic and Indian Oceans are important as they influence the atmospheric circulation over East Africa. A branch of anticyclonic flow generated from the subtropical high over southern Atlantic Ocean converges with easterly flow generated
from anticyclonic flow over southern Indian Ocean; weakened over the southeastern African coast. MAM is a season of lowest SLP over East Africa. The subtropical highs in both the hemisphere are weak from January to May and gain strength during JJAS.

Northeast monsoonal winds during JF flow parallel to the coast and are weakened over land. Dry northerly winds originating over northern Africa prevail over central East Africa during JF. In this season the Atlantic air masses converge with easterly monsoon winds from the Indian Ocean over southeastern Africa, distorting the ITCZ in a complex north-south orientation. The reversal of monsoonal winds occurs during the transition season of MAM, in which divergent flow from southern Indian Ocean starts intensifying. Easterly and southeasterly flow prevails over southern East Africa, while northerly winds continue to prevail over northern East Africa during MAM. The intense southeasterly trade winds from the Indian Ocean penetrate inland as far as the Atlantic coastline (McHugh, 2004b). The Atlantic airmasses are not directed toward East Africa during MAM.

Complete reversal of JF winds occurs during JJAS. The strong southeasterly monsoonal winds from Indian Ocean bring moisture over to the coastal East Africa and the Horn of Africa. Northerly wind prevails over most of the interior East Africa during JJAS.

The East African low-level jet, also known as Findlater jet or Somali jet, is the low level jet with its core developed around 1500m or 850mb (McGregor and Nieuwolt, 1998). This jet exists throughout the year, but its geographical extent changes from month to month. During the onset of Indian summer monsoon, the land-ocean temperature and pressure gradient between the Indian subcontinent and Indian Ocean
Figure 4.2.1: Mean seasonal circulation for JF and MAM seasons is plotted as vectors. Length of the vector proportional to 5 m s\(^{-1}\) wind velocity is shown in upper right corner. Seasonal mean sea level pressure (in hPa) is plotted as contours of equal pressure and seasonal mean rainfall rate (in units of kg m\(^{-1}\) day\(^{-1}\)) is plotted in shades of grey.
Figure 4.2.2: Mean seasonal circulation for JJAS and OND seasons is plotted as vectors. Length of the vector proportional to 5 ms\(^{-1}\) wind velocity is shown in upper right corner. Seasonal mean sea level pressure (in hPa) is plotted as contours of equal pressure and seasonal mean rainfall rate (in units of kg m\(^{-1}\) day\(^{-1}\)) is plotted in shades of grey.
causes the shift of Somali jet and it penetrates inland of East Africa during May-July (Figure 4.3). Maximum ascent of air in this jet core occurs along the coast of East Africa. The jet core brings with it the cloudiness and moisture. The surface airflow in southern Somalia and Ethiopia is prevailing from southeast. These winds parallel the coast.

![Figure 4.3: Monthly propagation of the East African low level jet core (McGregor and Nieuwolt, 1998)](image)

OND is a transition season between the southeast and northwest monsoon. Over the coastal region easterly winds prevail, while the Horn of Africa is a northeasterly airflow. During OND, the trades originate in south Indian Ocean and the easterly flow prevails over most of the East Africa region. The westerly flow, which is the south Atlantic moist air mass, converges with trades over East Africa, producing enhanced rainfall during short rains.
4.1.2 Annual Rainfall Rate and Precipitable Water

Rainfall rate is a measure of an intensity of rainfall. It is calculated as the amount of rain that would fall over a given interval of time. The rainfall intensity is assumed constant over that time period. The high rainfall rate regions are located along the ITCZ, parallel to latitudes, (Figures 4.2.1 to 4.2.4) and move seasonally over East Africa, along with the ITCZ and so are the high precipitable water regions (Figures 4.4.1 and 4.4.2). In the region of East Africa where the ITCZ has north-south orientation as well as over highlands of Ethiopia and Kenya, rainfall rate and precipitable water seem to have out of phase relationship. The high rainfall rate regions are to the north of the equator during JJAS, to the south of the equator during JF and over the equatorial regions during MAM and OND. Over coastal East Africa, the rainfall rate is less in all the seasons, except during JF, when higher rainfall rates are found over coastal East Africa to the south of equator. High precipitable water is embedded in large amounts in Atlantic air masses as compared to in southeast trades.

Though the precipitation rate is generally lower over the coastal regions of East Africa, the precipitable water is higher. The lower precipitable water region is located inland, parallel to the coast, in all four seasons. Central Africa has high precipitable water in all seasons.

4.1.3 Annual Precipitation

The rapidly changing complex climatic patterns of East Africa are illustrated by the rainfall patterns over the region. The seasonal migration of the ITCZ, regional orography, large-scale monsoon circulation (Black et al, 2003), large lakes, variations in vegetation type, and land-ocean contrast give rise to temporal and spatial variations in
Figure 4.4.1: Precipitable water in kg m\(^{-2}\) for (a) JF and (b) MAM.
Figure 4.4.2: Precipitable water in kg m\(^{-2}\) for (a) JJAS and (b) OND.
precipitation pattern over East Africa (Indeje et al, 2000). Due to the complex rainfall seasonality, changing within tens of kilometers, diverse climate exists in East Africa ranging from desert to forest, over a relatively small region (Nicholson, 1996). Despite this diversity, interannual fluctuations of rainfall are uniform over the region and appear to be governed by the same factors (Nicholson, 1996). Using cluster analysis, Indeje et al. (2000) have identified eight homogeneous regions of climate variability of East Africa, between 10°S-4°N and 30°E-43°E.

4.1.4.1 Rainfall Seasons

In much of the region of East Africa, the annual cycle of rainfall is bimodal, except northern and southern extremes of East Africa, where it is unimodal. These regions are to the north or to the south of the maximum extent of the ITCZ location in respective hemispheres. The regions with unimodal rainfall cycle receive the maximum rainfall during the summer season of the respective hemispheres (Nicholson, 1996).

The seasonality changes significantly within tens of kilometers (Nicholson, 1996). The two main synoptic subdivisions of annual pattern in East Africa according to J.F. Griffiths (1972) are, the period from early June to early October and the period from November to end of April. He considers May and late October as transitional stages between annual pattern changes. Nicholson and Kim (1997) consider two main rainy seasons in East Africa, as ‘short rains’ (October – December), and ‘main rainy season’ (March – May). June-September also could be considered as third rainfall season during which, coastal and northwest East Africa receives rainfall due to south Asian monsoon circulation. About 42 percent of annual rainfall is received during main long rainy season. During long rains the highest intensity of rainfall is observed near Lake Victoria.
and highlands of East Africa (Indeje et al., 2000). June – September rainfall season accounts for 15%, while, October – December rainfall season accounts for 25% of total regional annual rainfall. During short rains rainfall is distributed in whole East African region (Indeje et al, 2000).

4.1.4.2 Climatology of Precipitation over East Africa

It can be seen from this Figure 4.4 that MAM and OND are the seasons when entire East Africa region receives rainfall, while in JF the region to the south of the equator receives significant rainfall and in JJAS northwestern East Africa receives and coastal East Africa to the south of equator receives significant rainfall. During MAM and OND the spatial distribution of rain though not identical, is very similar.

In JF season, the major concentration of rainfall is over southern Tanzania, northern Mozambique, Zambia, and Malawi, where the rainfall is over 200 mm. Kenya, Rwanda, Burundi, Uganda, northern Tanzania, and the Democratic Republic of Congo (Congo) receives about 50 mm to 150 mm of rainfall, while Somalia, Ethiopia, and Sudan receives less than 50 mm rainfall. The seasonal rainfall during JF is associated with the extreme southward location of ITCZ and partly with moisture influx from the Indian Ocean (Indeje et al., 2000).

MAM has large spatial rainfall variability. This season is considered as the main rainy season, during which total amount of rainfall is similar to that during OND, but is distributed over a longer time period. The ITCZ moves slowly over East Africa during this season (Black et al., 2003) prolonging the time span of precipitation receipt. This produces a less coherent rainfall distribution (Camberlin and Phillipon, 2002) compared to that during OND (McHugh, 2004b). During MAM the northern part of Congo and few
Figure 4.5: Climatological seasonal precipitation over Africa. Contours are drawn at 50 mm.
small regions in Uganda and Tanzania receive over 150 mm of rainfall. The remaining East Africa receives 50 mm to 150 mm of rainfall, except northern Somalia, where the rainfall is less than 50 mm. The spatial variability is due to the dominance of local factors affecting and causing rainfall during MAM.

During JJAS, Ethiopia and northwestern Congo receives rainfall over 150 mm, while all southern East Africa, Kenya, and Somalia receive less than 50 mm of rainfall. Coastal East Africa receives rainfall from moist air associated with the intensified Somali Jet as the core of this jet is best developed at the beginning of this season. The rainfall is concentrated only over few regions, especially over highlands and coastal regions. JJAS is the driest season for the East Africa region.

OND is a short rainy season during which the rainfall is well-distributed especially over equatorial and southern East Africa. The ITCZ moves rapidly across East Africa during this season. In OND, the Congo and few small regions in Kenya, Tanzania, Zambia, Mozambique and Malawi receive over 150 mm of rainfall, while the rest of East Africa receives 50 mm to 150 mm of rainfall, with the exception of Somalia and northern Ethiopia, where the rainfall is less than 50 mm. Dominance of large-scale weather systems may be responsible for the special homogeneity of rainfall during OND (Indeje et al., 2000).

Throughout the annual cycle the Horn of Africa; especially Somalia receives very little rainfall making its land the most arid among the East African countries.

4.1.4.3 Effect of Topography and Lakes on East Africa Rainfall

In equatorial region, relatively uniform pressure and temperature patterns are
observed. The small pressure gradients often modify the flow pattern on the surface of continents in the equatorial regions (Edwards et al., 1983). In equatorial East Africa, the interaction between the flows and topographic barriers (such as the Ethiopian highlands and Rift Valley of Kenya) and large lakes (such as those in the Great Rift Valley) gives rise to local forced convection and to heavy rainfall (Edwards et al., 1983). During main rainy season of March-May, the highest intensity of rainfall is observed near Lake Victoria and the highlands of East Africa (Indeje et al., 2000). This shows that the local factors such as topography and the presence of water bodies are dominant factors, modulating the rainfall pattern during the main rainfall season. During the southwest monsoon (JJAS), the moisture influx from Indian Ocean results in high rainfall over northern East Africa, for example to the windward side of Ethiopian Highlands, while, southern East Africa remains dry as the monsoon flow is diverted away by the high topography of coastal regions of East Africa and Madagascar.

The highlands and lake regions of East Africa, such as, areas of highest relief surrounding Mount Kenya and Mount Kilimanjaro in northwest, the hilly regions northeast of Lake Malawi, northwest of Lake Victoria, and the Ethiopian highlands, receive the most rainfall (Nicholson, 1996). These large water bodies modulate the moisture content of the local atmosphere as well as the diurnal small scale circulation and thermal stability of the atmosphere above.

Lake Victoria and its catchment region is a discrete zone within the East Africa (Ogallo et al., 1988) and is identified as a homogeneous rainfall region (region VI in Figure 1b in their paper) by Indeje et al. (2000). This region receives more rainfall compared to surrounding regions. The nocturnal lake-breeze circulation produces
convergence over the lake, as the temperature over the lake is approximately 3\(^\circ\) lower (Ba and Nicholson, 1998) compared to the surrounding region. This thermal instability of the boundary layer over the lake enhances the convergence. As a result of this convergence, large cumulonimbus cold clouds develop over the lake region during the night (maximum of cold clouds occur at local midnight) and the Lake Victoria and its catchment area receives enhanced rainfall (Ba and Nicholson, 1998). This nocturnal circulation over the Lake Victoria produces about 50 percent of rainfall over the land especially on the western side of the lake (Ba and Nicholson, 1998). The Lake Tanganyika region also receives enhanced nocturnal rainfall due to similar mechanism, especially in November (Ba and Nicholson, 1998). Mistry and Conway, (2003) have shown that the variations in rainfall receipt over East Africa are evident in the Lake Victoria level fluctuations and also have shown through spectral analysis that ENSO is a predominant factor responsible for long-term variability of Lake Victoria rainfall.

4.1.5 Aridity

The origin of the general rainfall deficiency over East Africa lies in the atmospheric circulation above East Africa (Trewartha, 1961). Though being at equatorial position, East Africa does not receive much rainfall. The reasons for this dryness are the divergent character of both the northeast and southwest monsoons, the shallow depth of the southwest monsoon; the strong meridional flow in all but the transition seasons; and the stable stratification aloft together with a marked decline in the moisture content (Trewartha, 1961). The general deficient characteristic of equatorial East Africa rainfall is intensified towards the north of equator and in the Horn of Africa (Trewartha, 1961) and also from west to east (Nieuwolt, 1977).
The relatively arid conditions in much of the East African region are not understood fully. No major factors explain this aridity, while many local and global factors play a role (Trewartha, 1961; Anyamba, 1984). Atmospheric circulation could be one of the important factors. Thermal and dynamic stability of both the monsoons, shallow moist air streams, dry and stable easterly air originating from anticyclones over Sahara, Arabia and from the Mascarene High, anticyclonic flow aloft East Africa are the global factors, while the Ethiopian highlands, the low level Turkana jet (Kinuthia, 1992), cold upwelled water along the Somali coast, and the low-level Somali jet running parallel to the east coast of East Africa, enhancing the frictionally induced subsidence (Nicholson, 1996) are the local factors responsible for arid conditions. Aridity over Ethiopian Highlands is enhanced due to the divergent flow resulting from regional pressure patterns and extreme heating in summer (Nicholson, 1996). Rainfall variability determines the region’s water balance.

Figure 4.6 shows the climatological aridity index over East Africa. Northern East Africa is drier and arid in JF, while, southern East Africa is wet. The higher precipitation receipt over southern East Africa makes it less arid. The regions with the highest aridity index are along ITCZ in JF. Aridity index over most of the East Africa is below 2.0 in MAM. The Horn of Africa of Africa and northern East Africa to the north of 8N and to the south of 16 S are very arid. During JJAS northern East Africa has high aridity index, while southern East Africa is very arid in this season. The Horn of Africa of Africa is highly arid also during this season. In OND, higher aridity index values are along equator. The Horn of Africa region has aridity index lower than 0.5 in all seasons, making that region highly arid.
Figure 4.6: Climatological aridity index over East Africa (Method 1) for the four seasons. The contours of equal aridity index are drawn at 0.5.
4.2 Annual Cycle of Other Hydroclimatic Variables

Most of the hydroclimatic variables were plotted at 850 hPa pressure level as the surface pressure over East Africa is generally below 1000 hPa. Plots for hydroclimatic variables at 1000 hPa would not have represented the actual conditions.

4.2.1 Surplus and Deficit

Figures 4.7 and 4.8 show the climatological surplus and deficit respectively for Africa. The higher surplus regions are seen along the ITCZ positions in each season. The Horn of Africa and coastal East Africa have higher deficit values (>100 mm) in all seasons. These are the dry and arid regions of East Africa. Though coastal East Africa receives rainfall during MAM and OND (Figure 4.5), the region experiences water deficit.

4.2.3 Mid-Monthly Soil Moisture

Central equatorial Africa has higher soil moisture (>100 mm) in all seasons. The regions of higher soil moisture are seen along the position of ITCZ in each season. The highlands of Ethiopia and Kenya have higher soil moisture as these regions receive precipitation in all seasons. The Horn of Africa is the arid region with soil moisture less than 10 mm in all seasons. East Africa to the south of the equator has lower soil moisture (20 mm to 60 mm) in JJAS and OND.

4.2.4 Climatology of Potential Evapotranspiration (PET)

PET is a measure of ability of atmosphere to remove moisture from the surface through the processes of evaporation and transpiration, assuming no control on water supply. This ability of the atmosphere depends upon number of factors including air temperature, humidity and winds, as these variables decide the capacity of atmosphere
Figure 4.7: Climatological seasonal surplus for Africa. Contours are drawn at 20 mm.
Figure 4.8: Climatological seasonal deficit for Africa. Contours are drawn at 20 mm.
Figure 4.9: Climatological seasonal soil moisture depth (mm). Contours are drawn at 10 mm.
to contain water vapor. The higher the air temperature, higher is the water containing
capacity of atmosphere and the higher the humidity in the atmosphere lower is the rate of
evapotranspiration. Winds help to remove water from surface through the process of eddy
diffusion. Higher wind speeds could increase rate of evapotranspiration. The rate of
evapotranspiration is also associated with the gradient of vapor pressure between the
surface and the atmosphere. Other factors influencing PET are surface type (water or
land, which provides the available water for evapotranspiration), soil type (for bare soil),
and vegetation. PET is an important factor affecting aridity over the region.

Figure 4.9 shows the climatology of potential evapotranspiration over East Africa.
During JF, high PET values (above 200 mm) are found in a few regions in southern
Somalia, southern Ethiopia, and in southwestern Tanzania. Most of the coastal East
Africa has PET between 150 mm and 200 mm, while equatorial and southern East Africa
has PET less than 100 mm. In the long rain (MAM) season, high PET values (>200 mm)
are all toward northern Africa, especially in Somalia, coastal Kenya, most of the
Ethiopia, southern Sudan, and in southwestern Tanzania, while, coastal southern East
Africa still has PET between 150 mm and 200 mm. During MAM, central and southern
East Africa has PET less than 100 mm. In JJAS, most of the East Africa to the north of
equator has PET between above 100 mm, except in Ethiopia, Uganda, and Kenya. Eritrea
and few pockets in coastal Somalia show PET above 150 mm. most of the central and
southern East Africa has PET less than 100 mm. A small region to the southeastern
border of Lake Tanganyika has PET above 150 mm. During the short rains (OND)
Ethiopia, Uganda, Kenya, northern Somalia, and central equatorial East Africa has PET
less than 100 mm. Pockets in southern Somalia, Uganda, Kenya and Tanzania show high
(>200 mm) PET. The rest of the East Africa has PET between 100 mm and 150 mm. The climatological PET values of East Africa do not show a relation to climatological precipitation. A few permanent features can be observed in climatological PET, such as, in the higher coastal East African PET (> 150 mm) in a small region in Tanzania, the higher (> 200 mm) in southeastern boundary of Lake Tanganyika and lower PET (< 100 mm) Lake Victoria and its catchments.

4.2.5 Temperature

For most of the region of East Africa, the annual range of surface air temperature variations is between 4°C to 7°C. At the 850 hPa pressure level, air temperature is in the range of 15°C to 20°C. The temperature varies little during annual cycle with the migration of Sun to the north or to the south of equator. Maximum temperature variation is observed over Somalia compared to other countries of East Africa as Somalia receives very less rainfall in a typical year and does not have high elevation regions.

4.2.6 Dew Point Depression (DPD)

DPD is calculated as the difference between air temperature and dew point temperature at given pressure level. DPD is an important indicator of the moisture content of air and the extent to which air must be lifted for condensation. The lower the DPD, moister the air is. An increase in DPD decreases the likelihood of cloud formation. Though the air temperature does not show variation in association with movement of the ITCZ, regions of low DPD show association with the ITCZ movement over East Africa (Figure 4.6.1 and 4.6.2), as ITCZ is generally a high moisture zone. Low DPD (below 7.5 °C) is observed over southern East Africa during JF and over northern East Africa during JJAS. An east-west gradient in DPD is observed over equatorial Africa with higher DPD.
Figure 4.10: Climatological potential evapotranspiration for Africa. Contours are drawn at 50 mm.
Figure 4.11.1: Temperature in degree Celsius for (a) JF and (b) MAM seasons.
Figure 4.11.2: Temperature in degree Celsius for (a) JJAS and (b) OND season
over the eastern coast and lower interiors. Equatorial regions have low DPD during
MAM and OND. The high elevation regions of Ethiopia and Kenya have higher DPD
compared to the surrounding regions. The Horn of Africa and coastal East Africa have
higher DPD (>15 °C) in all the seasons despite the movement of ITCZ. The spatial
distribution of DPD over East Africa is similar to that of precipitable water and specific
humidity over East Africa. Regions of high (low) rainfall rate and higher (lower)
precipitable water generally have lower (higher) DPD.

4.2.7 Specific Humidity (SH)

SH is calculated as ratio of weight of water vapor to the total weight of air in a
moist air sample. Thus higher the moisture content of the air, the higher is the SH. SH is
higher over East Africa in both rainy seasons MAM and OND compared to other two
seasons. The high SH region is well-spread over entire East Africa in MAM compared to
that in OND. During JJAS, high SH is found over the highlands of Ethiopia and near the
large lakes of East Africa. The SH is high over the regions over which ITCZ is located
and thus have north-south orientation. But, the high SH regions do not move much over
East Africa with ITCZ movements. The lowest SH is found during JJAS, which is the
driest season of East Africa. The SH along with air temperature modulates the
evapotranspiration rate over the region.

4.2.8 Moist Static Energy (MSE)

MSE is obtained as combined internal energy, latent heat, and gravitational
potential energy of moist air. MSE is a measure of thermodynamic state of air. It is a
thermodynamic variable (analogous to equivalent potential temperature), calculated by
hypothetically lifting air adiabatically to the top of the atmosphere and allowing all water
Figure 4.12.1: Dew point depression in degrees Celsius for (a) JF and (b) MAM seasons at 850 hPa pressure level.
Figure 4.12.2: Dew point depression in degrees Celsius for (a) JJAS and (b) OND seasons at 850 hPa pressure level.
Figure 4.13.1: Specific humidity in kg kg⁻¹ for (a) JF and (b) MAM seasons.
Figure 4.13.2: Specific humidity in kg kg\(^{-1}\) for (a) JF and (b) MAM seasons.
vapor present in the air to condense and release latent heat.

\[ \text{MSE} = C_p T + gz + L_v R \]

Where, \( g \): gravitational acceleration,

\( L_v \): latent heat of vaporization,

\( C_p \): specific heat at constant pressure for air \((1005 \text{ J kg}^{-1} \text{ K}^{-1})\),

\( T \): absolute temperature \((\text{K})\),

\( z \): height above some reference level (start either \( z = 0 \),

or the height where the ambient pressure is 100kPa), and

\( R \): water vapor mixing ratio in the air.

Figure 4.14.1 and 4.41.2 shows the MSE over Africa at 850hPa. A high MSE region is a region of deep tropical convection. Being at tropics, throughout the year, MSE remains almost the same and high \( (> 336 \text{ kJ/kg}) \) over entire East Africa region at 850 hPa pressure level. During JF, a small region of high MSE \( (>340 \text{ kJ/kg}) \) appears along 15°S.

Moist static instability (MSI) occurs in an atmosphere where actual environmental lapse rate is greater than the dry adiabatic lapse rate (Wallace and Hobbs, 1977). The release of instability promotes rising of air due to positive buoyancy and accelerates the formation of thunderstorm systems.

An atmosphere with high instability enhances convective motions and influences the rainfall receipt over the region. MSI (Figures 4.15.1 and 4.15.2) is positive over East Africa in all seasons, except in JJAS, when southern East Africa and coastal East Africa has low negative MSI. This indicates that the atmosphere above most of the easternmost part of Africa is quite stable during average year, thus suppressing convective motion and rainfall receipt. On the other hand, regions of
Figure 4.14.1: Moist static energy in kJ kg\(^{-1}\) for (a) JF and (b) MAM seasons
Figure 4.14.2: Moist static energy in kJ kg\(^{-1}\) for (a) JJAS and (b) OND seasons
Figure 4.15.1: Moist static instability in kJ kg\(^{-1}\) for (a) JF and (b) MAM seasons
Figure 4.15.2: Moist static instability in kJ kg\(^{-1}\) for (a) JJAS and (b) OND seasons
East Africa such as equatorial central and western East Africa and the Ethiopian and Kenyan highlands have higher MSI, while coastal East Africa and the Horn of Africa, which are the driest regions, have lower MSE compared to the surrounding areas.

4.2.9 Outgoing Longwave Radiation (OLR)

OLR values (<200 W/m²) are usually associated with high tropical convection and thus low OLR values are found over the location of ITCZ in the equatorial region and over the regions with colder temperatures. Thus, equatorial land masses and highlands generally have lower OLR values. Central and western Africa have lower OLR values, while coastal East Africa has higher (>250 W/m²) OLR values, suggesting the suppressed convection over coastal and southern East Africa. The spatial distribution of OLR does not change much over East Africa during an annual cycle.

4.2.10 Vertical Velocity and Geopotential Height

The subsiding nature of the atmospheric circulation over East Africa can be observed in the Figures 4.17.1 and 4.17.2 showing vertical velocity. The vertical velocity is negative (indicating downward motion of atmosphere) over most of the East Africa during an annual cycle. Vertical velocity is lower over coastal East Africa and the Horn of Africa compared to that over central and the part of western East Africa. The spatial distribution of precipitable water is consistent with that of vertical velocity.

Geopotential height (HGT) is a measure of geometric height that accounts for the dependence of gravity on latitude and height. Convergence regions have higher HGT. HGT does not change much seasonally over East Africa (Figures 4.18.1 and 4.18.2). During JJAS at the 850 hPa pressure level, lower HGT (1440 gpm-1480 gpm) could be seen above central Africa,
Figure 4.16.1: Upward longwave radiation flux in W m\(^{-2}\) for (a) JF and (b) MAM seasons.
**Figure 4.16.2:** Upward longwave radiation flux in W m$^{-2}$ for (a) JJAS and (b) OND seasons.
Figure 4.17.1: Pressure vertical velocity in Pa s$^{-1}$ for (a) JF and (b) MAM seasons.
Figure 4.17.2: Pressure vertical velocity in Pa s\(^{-1}\) for (a) JJAS and (b) OND seasons.
Figure 4.18.1: Geopotential height in geopotential meters (gpm) for (a) JF and (b) OND seasons.
Figure 4.18.2: Geopotential height in geopotential meters (gpm) for (a) JF and (b) OND seasons.
while comparatively the higher (>1480 gpm) HGT is over coastal East Africa and southern Africa. HGT field shifts northwards as ITCZ shifts northward. The higher HGT field expands towards north during MAM up to 40°N, while an even higher (>1520 gpm) HGT field appears above southern Africa south of 15°S. During JJAS, above central equatorial and northern East Africa, HGT increases between 1480 gpm-1520 gpm, while the higher HGT field expands above southern East Africa gains HGT up to 10°S. HGT lowers during OND with southward shifting of ITCZ.

4.2.11 NDVI

NDVI has been shown to have association with precipitation. The precipitation promotes vegetation growth and thus in general, higher precipitation regions have higher NDVI. Figure 4.19 shows the climatological seasonal NDVI over Africa.

Central Africa has relatively high NDVI (> 0.4) in all seasons, while, the Horn of Africa has lower NDVI (< 0.2) in all seasons. The highlands of Ethiopia and Kenya have higher NDVI as this area receives precipitation in most of the months from orographic enhancement.

4.2.12 PDSI

Palmer Drought Severity Index (PDSI) is a drought index responding to abnormally dry or wet weather conditions. The PDSI is calculated using monthly air temperature and precipitation based on the surface moisture balance. The PDSI has been shown to correlate well with the soil moisture over US. Theoretically, PDSI varies between -10 and +10. The lower (higher) the PDSI, the drier (wetter) the region. Central equatorial Africa and the highlands of Ethiopia and Kenya have positive PDSI in all
seasons. The Horn of Africa and coastal East Africa as well as the central East Africa have negative PDSI values implying dryness over those regions.

**Figure 4.19**: climatological seasonal NDVI over Africa. Contours are drawn at 0.05.
Figure 4.20: Climatological seasonal PDSI over Africa. Contours are drawn at intervals of 5.
Chapter 5: Results and Analysis

The following analysis describes the results of principal components analysis (PCA) and correlation analysis performed to examine (a) the spatial and temporal correlations of aridity index over East Africa with hydroclimatic variables, and with climatic indices such as El Niño – Southern Oscillation (ENSO) index (NINO Index), North Atlantic Oscillation (NAO) index, Indian Ocean Dipole Mode (IODM) index, Tropical Atlantic SST Dipole (TASD) index and Quasi Biennial Oscillation (QBO) index, (b) the association of first principal components (EOF1) of the aridity index seasonal spatial distribution with the hydrological variables such as precipitation, PET, surplus, deficit and mid-monthly soil moisture; hydroclimatic variables such as sea surface temperature (SST), sea level pressure, precipitable water and surface air temperature; and other variables such asNormalized Vegetation Difference Index (NDVI), Palmer Drought Severity Index (PDSI) and aerosol index; and (c) global teleconnections such as ENSO, NAO, IODM, TASD, and QBO in the spatial and temporal seasonal distribution of EOF1.

PCA was performed on aridity index to identify characteristic modes of temporal variability of the aridity index and the associated spatial patterns across the East Africa. The analysis was to produce the time series and associated spatial patterns of hydroclimatic anomalies and the climatic indices to find the global teleconnections. Time series of area-averaged hydrological variables, NDVI, PDSI, and climatic indices were plotted along with time series of EOF1 in order to show the absolute changes in those variables associated with the EOF1 time series.
5.1 Aridity Index

Aridity index was calculated as the ratio of mean seasonal precipitation to mean seasonal potential evapotranspiration (PET) using 2 methods. In first method, aridity index was calculated for each month using monthly precipitation and PET data. If precipitation value in any month is zero, it was replaced by $10^{-6}$ mm, before calculating the aridity index to avoid the division by zero. Thus obtained monthly aridity index was then averaged seasonally. In second method, seasonal of precipitation and potential evapotranspiration was calculated which was used to calculate seasonal aridity index. Both the methods give very similar results as shown in Figures 5.1 and 5.2. The difference in the results using two methods was found to be negligible. The first method was then used for all further analysis. To prove that the difference in the aridity index calculated is negligible, NINO 3.4 Index was correlated with aridity index calculated using both the methods and the Figures 5.3 and 5.4 show that the correlation patterns thus obtained have negligible difference.

Correlations of aridity index with precipitation and with PET were plotted in order to investigate the comparative influence of precipitation and PET in deciding aridity index over East Africa. The contour maps of these correlations with precipitation are shown in Figure 5.4. During JF, precipitation is highly correlated to aridity index over East Africa except over the Horn of Africa, where correlation coefficients are less than 0.2. Correlation coefficients are greater than 0.6 over East Africa to the south of equator as this region receives maximum rainfall during JF, while correlation coefficients are between 0.2 and 0.5 over East Africa to the north of equator. During MAM correlation coefficients between precipitation and the aridity index are between 0.8 and 1.0 over East
Figure 5.1: Spatial distribution of climatological aridity index over East Africa for the four seasons. Contours of equal aridity index are drawn at the interval of 1.2. (Method Two)
Figure 5.2: Spatial distribution of correlations of aridity index (Method One) over East Africa and NINO Index for the four seasons. Contours of equal aridity index are drawn at the interval of 1.2.
Figure 5.3: Spatial distribution of correlations of aridity index (Method 2) over East Africa and NINO Index for the four seasons. Contours of equal aridity index are drawn at the interval of 1.2.
Africa, during which almost entire East Africa receives rain. During JJAS, correlation coefficients are greater than 0.6 to the north of equator and less than 0.6 to the south of equator. Over a small region between latitudes of 5°S and 10°S and longitudes of 20°E and 30°E, this correlation was negative indicating that the aridity index is not linearly proportional to the amount of precipitation. During OND, over the northwest East Africa correlation coefficients are less than 0.4 and are in the range of 0.4 to 1.0 over the rest of East Africa.

Figure 5.5 shows the correlations of PET and aridity index. PET is related inversely to the aridity index and thus the negative correlations between PET and aridity index. During JF, significant negative correlations are over southern East Africa, with correlation coefficients between 0.3 and 0.5, which also is a region of high precipitation during this season. During MAM and OND, most of East Africa has correlation coefficients above 0.5. During JJAS, significant negative correlations are to the north of the equator. The Horn of Africa is the only region over which a low positive correlation was found between PET and aridity index.

The time series of precipitation and that of PET (Figure 5.6) shows that the magnitude of change in average precipitation over East Africa is much more compared to the magnitude of change in PET of the region. The values of precipitation vary from 30 to 100 mm, while the values of potential evapotranspiration vary from 75 to 120 mm. The variability in precipitation is much more compared to that of PET.

5.2 Principal Components Analysis

PCA was performed on aridity index over East Africa where in one case the annual cycle of the aridity index was removed before performing PCA (Figure 5.7) and in
Figure 5.4: Spatial distribution of correlation of aridity index and precipitation over East Africa. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.5: Spatial distribution of correlation of aridity index and PET over East Africa. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
other case annual cycle was retained (Figure 5.8). When the annual cycle of aridity index was removed before performing PCA, the first principal components of aridity index EOF1 (acr – annual cycle removed) explained 14.82 percent of the total variance, and EOF2, EOF3, and EOF4 explained 9.54 percent, 8.03 percent, 6.43 percent of total variance, respectively (Figure 5.7). When the annual cycle in aridity index was kept while performing PCA, the first principal components EOF1 (wac – with annual cycle) explained 47.46 percent of the total variance and EOF2, EOF3 and EOF4 explained 15.44 percent, 5.04 percent and 4.41 percent of the total variance, respectively (Figure 5.8). Only EOF1s in both the cases were used for the further analysis. Contours of equal loadings were plotted for each principal components of aridity index.

**5.2.1 Correlations of EOF1 with Hydrology Variables**

**5.2.1.1 Precipitation**

Figures 5.9 and 5.10 show the correlation of precipitation over East Africa with
Figure 5.7: First four principal componentss (EOF1 to EOF4) of aridity index with annual cycle removed (acr). Contours of equal loadings are drawn (contour interval is different for EOF1 to EOF4).
Figure 5.8: First four principal components (EOF1 to EOF4) of aridity index with annual cycle (wac). Contours of equal loadings are drawn (contour interval is different for EOF1 to EOF4).
EOF1 (acr) and with EOF1 (wac) respectively. There is considerable difference in the correlation pattern when the annual cycle is removed before calculating EOF1 time series. The influence of ITCZ shift is clearly evident in Figure 5.10 as the annual cycle of aridity index is intact. The significant negative correlations are seen over southern East Africa throughout the year except JJAS seasons. In JJAS, correlations are positive (coefficients between 0.4 and 0.8, significant at 99.99 percent level), in most of the northern East Africa and over central southern East Africa where correlation coefficients are between 0.2 and 0.3, significant at the 95 percent and 99 percent significance level. The correlations over the Horn of Africa and coastal northern East Africa are very small and non-significant (correlation coefficients between 0 and ± 0.1), where correlations are slightly positive in JJAS and are small negative in other three seasons. Figure 5.9 shows that the significant correlations generally are observed over East Africa between 10°N and 15°S. The correlations are positive in all seasons. Significant correlations (at 99 percent level and above) are observed over Horn of Africa and northern coastal East Africa in all seasons, expect in JJAS, when significant positive correlations are over northern and central equatorial East Africa and over southern coastal East Africa.

5.2.1.2 Potential Evapotranspiration

Figures 5.11 and 5.12 show the correlation of PET with EOF1 (acr) and EOF1 (wac) respectively. The correlations are negative as PET is inversely proportional to the aridity index. In JF, significant correlations are over western equatorial East Africa around Lake Victoria and Lake Malawi. In MAM, significant correlations are over most of the East Africa. In JJAS, positive correlations are seen over Horn of Africa and coastal East Africa. Significant negative correlations are observed over northern East Africa and
Figure 5.9: Spatial distribution of correlation of EOF1 (acr) and precipitation over East Africa. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.10: Spatial distribution of correlation of EOF1 (wac) and precipitation over East Africa. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.11: Spatial distribution of correlation of EOF1 (acr) and PET over East Africa. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.12: Spatial distribution of correlation of EOF1 (wac) and PET over East Africa. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
over southeast East Africa in JJAS. In OND significant correlations are over central and equatorial coast East Africa.

5.2.1.3 Surplus

Figures 5.13 and 5.14 show the correlation of surplus over East Africa with EOF1 (acr) and with EOF1 (wac), respectively. In Figures 4.16, the correlations are positive in all seasons. Higher surplus can correspond to the higher precipitation and thus to lower (higher) the aridity (aridity index). Significant correlations are observed over equatorial East Africa in JF, MAM, and OND seasons and over northern East Africa in JJAS. Significant correlations are absent over coastal equatorial East Africa during MAM. In Figure 4.17, significant correlations are negative in JF, MAM and OND seasons, while are positive in JJAS. Significant correlations are found over East Africa south of equator during JF, MAM and OND, while are found to the north of equator during JJAS. The positive correlations observed in JJAS over northern East Africa, between surplus and aridity index can be easily understood but the negative correlation of surplus and EOF1 (wac) in other seasons is spurious.

5.2.1.4 Deficit

Figures 5.15 and 5.16 show the correlations of deficit with EOF1 – (acr) and with EOF1 (wac), respectively. Deficit corresponds to lack on rainfall, and thus to higher (lower) aridity (aridity index) and hence the negative sign of correlation coefficient. This explanation does not hold in Figure 5.16 as the correlations are high positive over most of the regions of southern East Africa in JF, MAM, and OND seasons and over northern East Africa in JJAS. Figure 5.15 shows that the significant negative correlations are observed over northwestern East Africa and northern coastal East Africa in JF, over
Figure 5.13: Spatial distribution of correlation of EOF1 (acr) and surplus over East Africa. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.14: Spatial distribution of correlation of EOF1 (wac) and surplus over East Africa. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.15: Spatial distribution of correlation of EOF1 (acr) and deficit over East Africa. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.16: Spatial distribution of correlation of EOF1 (wac) and deficit over East Africa. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
northern East Africa to the north of 5°S and over southern East Africa to the south of 10°S in MAM. Negative correlations are not significant over southern equatorial East Africa in MAM. In JJAS significant correlations at 99 percent significance level are observed over north-central and equatorial central East Africa, while negative correlations are significant at 95 percent significance level over southern East Africa. Over the Horn of Africa and coastal East Africa positive correlations are observed as these areas receive rainfall in JJAS.

5.2.1.5 Mid-Monthly Soil Moisture

Figures 5.17 and 5.18 show the correlation of mid-monthly soil moisture with EOF1-annual cycle removed and with EOF1-with annual cycle respectively. In Figure 5.17, the correlation of EOF1 and soil moisture is positive in all seasons. Higher (lower) soil moisture corresponds to lower (higher) aridity and higher (lower) aridity index and hence the positive correlation. Significant correlations (at 99.99 percent significance level) are observed over northern East Africa in JF and OND. Correlations are significant at 95 percent and 99 percent levels over most of East Africa in MAM except over coastal East Africa, where correlations are not significant. In JJAS, significant correlations (at 99.99 percent significance level) are over north and north-central East Africa. Positive correlations over East Africa are seen in Figure 5.18. The significant correlations are over southern East Africa in JF, MAM, and OND, while they are over northern East Africa in JJAS.

5.2.2 Correlations with Hydroclimatic Variables

The time series of the EOF1 of aridity index was correlated with hydroclimatic variables such as surface air temperature, SST and precipitable water. The higher (lower)
Figure 5.17: Spatial distribution of correlation of EOF1 (acr) and mid-monthly soil moisture over East Africa. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90 percent, 95 percent and 99.9 percent significance levels, respectively.
Figure 5.18: Spatial distribution of correlation of EOF1 (wac) and mid-monthly soil moisture over East Africa. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
the aridity index; lower (higher) is the aridity, all the correlations were multiplied by \(-1\) for the ease of analysis. Thus in all the Figures in this section negative (positive) correlations refer to higher (lower) the magnitude of the hydroclimatic variables and lower (higher) aridity, and vice versa.

5.2.2.1 Surface Air Temperature

Surface air temperature was correlated with EOF1 (acr) (Figure 5.19) and with EOF1 (wac) (Figure 5.20). The near-surface air temperatures are important in assessing surface aridity than those at higher altitudes. The significant negative correlations indicate that higher surface air temperatures are associated with high aridity. In JF, Surface air temperature over central Africa and over southern Africa correlated negatively with EOF1 (acr). Surface air temperature over the western Indian Ocean and southeastern Atlantic Ocean also correlated negatively. The most significant negative correlation is over southern Africa. The surface air temperature over East Africa does not correlate much with EOF1 (acr). Surface air temperature over Africa to the south of equator correlated positively with EOF1 (wac), while the temperature over northern Africa correlated negative with EOF1 (wac). Significant positive correlations are over equatorial central Africa, while significant negative correlations are over northeastern Africa. Surface air temperature over southwestern Atlantic Ocean and over western Indian Ocean correlated positively with EOF1 (wac).

In MAM, surface air temperature over most of the Africa and over southeastern Atlantic Ocean and over western Indian Ocean correlated negative with EOF1(acr). Significant negative correlations are over sub-Saharan Africa and southwestern Indian Ocean. The surface temperature over most of Africa and the western Indian Ocean
Figure 5.19.1: Spatial distribution of correlation of EOF1 (acr) over East Africa and surface temperature for JF and MAM seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90 percent, 95 percent and 99.9 percent significance levels, respectively.
Figure 5.19.2: Spatial distribution of correlation of EOF1 (acr) over East Africa and surface temperature for JJAS and OND seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90 percent, 95 percent and 99.9 percent significance levels, respectively.
correlated positive with EOF1 (wac), with significant positive correlations over the southwestern Indian Ocean and over North Africa and Mediterranean Sea.

In JJAS, EOF1 (acr) correlated negative over Africa to the south of 10°N and over the southwestern Atlantic Ocean and western Indian Ocean. Significant negative correlations are over western Indian Ocean and Middle East. Surface air temperature over northwest Africa correlated significantly positively with EOF1 (acr). Surface air temperature over Africa between 10°N and 20°S, western Indian Ocean and eastern Atlantic Ocean correlated negative with EOF1 (wac).

In OND, surface air temperature over northern Africa correlated positive with EOF1 (acr), while southern Africa, western Indian Ocean, and the eastern Atlantic Ocean correlated negatively with EOF1 (acr), with significant correlations over northwest Africa and South Africa. Surface air temperature over Africa, and the Atlantic and Indian Ocean does not correlate well with EOF1 (wac).

5.2.2.2 Precipitable Water (PWAT)

In JF, PWAT over southern Africa (to the south of 15°S) and over Arabian Sea correlated significantly positively with EOF1 (acr), while that over southeastern Atlantic Ocean correlated significantly negatively. In JF PWAT correlated positively with EOF1 (wac) over most of the East Africa to the north of 15S, with significantly positively correlations over southern Sudan and Central republic of Africa. PWAT over the Horn of Africa does not show strong association with EOF1 (wac).

PWAT over north central Africa correlated significantly positively with EOF1 (acr), while the rest of Africa and the surrounding ocean did not correlate in MAM. EOF1 (acr) did not correlate with PWAT in OND. PWAT over East Africa did not correlate
Figure 5.20.1: Spatial distribution of correlation of EOF1 (wac) over East Africa and surface temperature. Contours of equal correlation coefficient are drawn at the interval of correlations over southern Sudan and Central republic of Africa. PWAT over Horn of Africa did not show much of an association with EOF1 (wac). 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.20.1: Spatial distribution of correlation of EOF1 (wac) over East Africa and surface temperature. Contours of equal correlation coefficient are drawn at the interval of correlations over southern Sudan and Central Republic of Africa. PWAT over the Horn of Africa did not show much of an association with EOF1 (wac). 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
much in MAM and OND as in JF with EOF1 (wac). PWAT over southern East Africa
correlated only slightly positively and that over northern East Africa showed low
negative correlations in MAM. This pattern is reversed in OND, when an opposite sign of
correlation was observed over the respective regions. PWAT over South Africa
correlations were significantly positive in MAM, and significantly negative in OND.
Over the region of Africa where the ITCZ is located during MAM, generally negative
correlations with PWAT were observed.

In JJAS, PWAT over northern Africa, the northern Atlantic and Indian Ocean
correlated positively with EOF1 (acr) and the significant positive correlations are over
Middle East, northern Africa, and Mediterranean Sea. PWAT over southern Africa,
Indian and Atlantic Oceans correlated negatively with EOF1 (acr) and significant
negative correlations are over the equatorial eastern Atlantic Ocean. Southern and
northern regions have correlation coefficients of opposite sign. The out of phase
relationship of PWAT over southern and northern East Africa were observed during
MAM, and enhanced during JJAS with EOF1 (wac). PWAT over northern East Africa
correlated significantly negative, except over Somalia, where correlations were but
significant negative. Likewise, PWAT over southern East Africa correlated positive, but
the correlations are not significant. Over the region of Africa where the ITCZ is located
during JJAS, generally negative correlations with PWAT were observed. PWAT over the
equatorial western Atlantic Ocean correlated significant positive, while PWAT over
northwestern Indian Ocean correlated negative with EOF1 (wac).

5.2.2.3 Global SSTs

In JF, Indian Ocean SSTs along the coast of India, to the south of Australia, along
Figure 5.21.1: Spatial distribution of correlation of EOF1 (acr) over East Africa and precipitable water for JF and MAM seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.21.2: Spatial distribution of correlation of EOF1 (acr) over East Africa and precipitable water for JJAS and OND seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.22.1: Spatial distribution of correlation of EOF1 (wac) over East Africa and precipitable water for JF and MAM seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.22.1: Spatial distribution of correlation of EOF1 (wac) over East Africa and precipitable water for JJAS and OND seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
the coast of southeast Africa and over equatorial central Indian Ocean showed significantly negative correlations with EOF1 (acr) (Figure 5.23.1). SSTs over the southern Pacific and Atlantic Oceans and over Antarctic Ocean correlated significant negative with EOF1 (acr). Central north Pacific SSTs in mid-latitude correlated significant positive, while SSTs over north eastern Pacific Ocean correlated significantly negative with EOF1 (acr). Indian Ocean SSTs to the south of Australia, along the coast of India, along the west coast of Singapore and along the coast of Mozambique; correlated negatively while a small region over central southern Indian Ocean along 45°S correlated positively with EOF1 (wac) (Figure 5.24.1). Atlantic Ocean SSTs in the central south Atlantic along 45°S correlated negatively with EOF1 (wac). SSTs of central subtropical Pacific Ocean correlated positively while Pacific Ocean SSTs off the west coast of North America and Antarctic Ocean SSTs correlated negatively with EOF1 (wac).

In MAM, SSTs over the Indian Ocean to the north and south of Australia, in Alaska Bay and in the southeastern Pacific Ocean correlated significantly negatively with EOF1 (acr), while central pacific SSTs correlated significant positively with EOF1 (acr) (Figure 5.23.1). SSTs off the southern coast of Indonesia and the northern coast of Australia and also the SSTs in southern Indian Ocean to the south of Australia correlated negatively with EOF1 (wac) (Figure 5.24.1). SSTs of Alaska Bay, the Pacific Ocean off the coast of Argentina, and the Antarctic Ocean over most of the region correlated significantly negative. A small ocean region off the East coast of US to the north of Bahamas in western Atlantic Ocean correlated significantly positive.

In JJAS, SSTs in the Indian Ocean the southern and equatorial Atlantic Ocean, in Arctic Ocean, and the southeastern and equatorial western Pacific Ocean correlated
significantly positively, while SSTs in northwestern Atlantic Ocean correlated significant negative with EOF1 (acr) (Figure 5.23.2). Indian Ocean SSTs correlated significantly positively during JJAS with EOF1 (wac) (Figure 5.24.2). Antarctic Ocean SSTs between longitudes 60°W and 150°W and between longitudes 120°E and 150°E correlated significant positive, while Antarctic SSTs between 50°E and 120°E and also between 150°E and 150°W correlated significantly negative with EOF1 (wac). Arctic Ocean SSTs correlated significantly positively with EOF1. Southern Atlantic Ocean SSTs to the south of 30°S correlated significantly positively, while Northern Atlantic SSTs off the coast of Greenland correlated significantly negatively with EOF1. The western Pacific Ocean off the East coast of China correlated significantly positive while eastern and central Pacific Ocean did not show any correlation.

In OND, southern central Pacific Ocean SSTs and southern Atlantic Ocean SSTs correlated significantly positively with EOF1 (acr), while SSTs in equatorial and northern Atlantic Ocean and in northwestern Pacific Ocean correlated significantly negatively with EOF1 (acr) (Figure 5.23.2). Indian Ocean SSTs did not correlate significantly with EOF1 (acr) or EOF1 (wac) during OND (Figure 5.24.2). Tropical and northern Atlantic Ocean SSTs correlated significantly negatively with EOF1 (wac). As during MAM, a small region off the East coast of US to the north of the Bahamas in the western Atlantic Ocean correlated significantly positive with EOF1 (wac).

5.2.3 Correlations of EOF1 with NDVI, PDSI and Aerosol Index

5.2.3.1 NDVI

Figures 5.25 and 5.26 show the correlations of NDVI with EOF1 (acr) and with EOF1 (wac) respectively. The NDVI over Africa are correlated with EOF1 of aridity
Figure 5.23.1: Spatial distribution of correlation of EOF1 (acr) over East Africa and SSTs for JF and MAM seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.23.2: Spatial distribution of correlation of EOF1 (acr) over East Africa and SSTs for JJAS and OND seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.24.1: Spatial distribution of correlation of EOF1 (wac) over East Africa and SSTs in JF and MAM seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
**Figure 5.24.2:** Spatial distribution of correlation of EOF1 (wac) over East Africa and SSTs for JJAS and OND seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
index over East Africa. NDVI has a positive correlation with precipitation (Figure 5.25). Correlations of NDVI and EOF1 are positive over most of East Africa, except over the Horn of Africa and coastal East Africa in MAM and JJAS. These regions receive less rainfall compared to the rest of East Africa. The positive correlations indicate that the higher (lower) the NDVI, higher (lower) is the aridity index and lower (higher) is the aridity. In JF, significant correlations are over north-central East Africa and equatorial western East Africa. The correlations are insignificant in MAM and JJAS, while significant correlations with correlation coefficient 0.5 and above are seen over the Horn of Africa, and coastal and central East Africa. When NDVI is correlated with EOF1 (wac), the significant correlations are found only in OND, when significant negative correlations are observed over southern East Africa and significant positive correlations over northwestern East Africa.

**5.2.3.2 PDSI**

Figures 5.27 and 5.28 show the correlations of the PDSI with EOF1 (acr) and with EOF1 (wac), respectively. The PDSI correlates positively with EOF1 (acr) (Figure 5.27) and PDSI has shown to correlate well with soil moisture and precipitation. Lower (higher) the PDSI, lower (higher) is the aridity index and higher (lower) is the aridity, hence the positive correlation. Most of East Africa has significant positive correlation with PDSI in JF, MAM, and OND, while significant correlations are over the northern and southern parts of East Africa in JJAS. When PDSI is correlated with EOF1 (wac), the correlations are positive over most of the East Africa and significant correlations are over southern East Africa in all the seasons, and are also over northern East Africa in JJAS.
Figure 5.25: Spatial distribution of correlation of EOF1 (acr) over East Africa and NDVI. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
**Figure 5.26:** Spatial distribution of correlation of EOF1 (wac) over East Africa and NDVI. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.27: Spatial distribution of correlation of EOF1 (acr) over East Africa and PDSI. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.28: Spatial distribution of correlation of EOF1 (wac) over East Africa and PDSI. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
5.2.3.3 Aerosol Index

Figures 5.29 and 5.30 show the correlations of the aerosol index with EOF1 (acr) and with EOF1 (wac), respectively. The aerosol index over East Africa is missing. In JF, the aerosol index over northern East Africa and over the Middle East correlated significantly negatively to EOF1, while the aerosol index over the equatorial eastern Atlantic Ocean correlated significantly positively with EOF1. In MAM, the aerosol index over sub-Saharan Africa and over northeast and equatorial east Atlantic Ocean correlated significantly negatively with EOF1. In JJAS, the aerosol index over most of the equatorial and southern Africa, over eastern equatorial Atlantic Ocean and over Middle East correlated negatively with EOF1, while the aerosol index over Sahara and North Africa did not correlate with EOF1. In OND, the aerosol index correlated significantly negatively over northern Africa and Middle East, and correlated mostly positively over eastern Atlantic Ocean, with EOF1. The aerosol index correlated negatively over most of the Africa and east Atlantic Ocean in JJAS and over North Africa, Middle East and equatorial east Atlantic Ocean in OND with EOF1 (wac). Aerosol index over southeast Atlantic Ocean correlated positive with EOF1 (wac). In MAM it correlated significant positive over North Africa, Middle East and eastern equatorial Atlantic Ocean, while; it correlated significant positive with aerosol index over North Africa and Middle East and negative with aerosol index over sub-Saharan Africa and eastern equatorial Atlantic Ocean in JF.

5.3 Correlation Analysis with Climatic Indices

Variability in aridity index is driven by the variability in precipitation over the region. Thus, the contour maps of correlation between any climatic indices considered
Figure 5.29: Spatial distribution of correlation of EOF1 (arc) and aerosol index over East Africa. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.30: Spatial distribution of correlation of EOF1 (wac) and aerosol index over East Africa. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
here and aridity index show striking similarity with the contour maps of correlation between the considered climatic index and the seasonal mean precipitation.

5.3.1 Correlational Analysis of NINO (Niño 3.4 Region) Index

5.3.1.1 Aridity Index, Precipitation and PET

El Niño has been considered to be an important atmospheric phenomenon affecting the climate over East Africa. Figures 5.2, 5.31, and 5.32 show the correlations of aridity index, precipitation, and PET, respectively with NINO Index. The maps of correlations for aridity index and precipitation with NINO Index are very similar for all seasons. PET over East Africa correlated positively with NINO Index in all seasons.

Northern and southern regions of East Africa are out of phase in terms of the sign of the correlation coefficient of aridity index and NINO Index. Aridity index (Figure 5.31) and precipitation (Figure 5.2) over northern East Africa correlated positively, while aridity index (Figure 5.2) and precipitation (Figure 5.31) over southern East Africa correlated negatively with the NINO Index. The negative correlation over southern East Africa means that during the warm (cold) phase of ENSO aridity is higher (lower) in JF. The negative (positive) correlation between NINO Index and precipitation over southern (northern) East Africa could mean that the warm (cold) phase of ENSO decreases (increases) the rainfall receipt over southern (northern) East Africa. PET correlated significant positive with El Niño index over East Africa to the south of equator (Figure 5.32), meaning that during warm (cold) phase of ENSO, PET is higher (lower). ENSO affects precipitation and aridity in the opposite manner over northern and southern East Africa in JF.

In MAM, the NINO Index correlated negatively with aridity index (Figure 5.2)
Figure 5.31: Spatial distribution of correlation of precipitation with NINO Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.32: Spatial distribution of correlation of PET with NINO Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
and precipitation (Figure 5.31) over East Africa. Significant negative correlations were found over Mozambique, Zimbabwe, and Zambia and also over central equatorial regions of Congo and Tanzania. PET (Figure 5.32) correlated significant positive over equatorial East Africa in MAM. This suggests that the reduced rainfall during main rains of warm ENSO year (Nicholson and Kim, 1997) and increased PET results in high aridity over East Africa.

Similar to JF, an out of phase relation is observed over northern and southern East Africa during JJAS, but the sign of the correlation is opposite to that in JF. During JJAS, a positive correlation of 0.2 in coastal Horn of Africa is associated with southwest monsoon winds. The Somali jet brings moisture over the coastal Horn of Africa and thus precipitation in the monsoon months of JJAS. Aridity is higher over Horn of Africa and southern Africa during JJAS, where aridity index and precipitation, and PET showed low positive correlation with NINO Index. This could mean that during the cold phase of ENSO, rainfall and PET are both reduced, but the rate of decrease of rainfall is higher compared to that of PET, causing higher aridity. Over northern East Africa, excluding the Horn of Africa, the lower aridity could be due to higher rate of precipitation than that of PET, during the cold phase of ENSO.

In OND, aridity index and precipitation over central equatorial East Africa and the Horn of Africa correlated significantly positively with the NINO Index and PET did not show much of a correlation. Over this region aridity is higher compared to over northern and southern East Africa; where negative correlations between aridity index and precipitation and positive correlation with PET were observed. This could mean that the
enhanced short rains (Nicholson and Kim, 1997) during the warm phase of ENSO in OND reduce the aridity over equatorial East Africa.

The time series of NINO Index is plotted with the time series of area-averaged precipitation over East Africa. Though NINO Index correlates well with precipitation (Figure 5.33), especially in MAM and OND seasons, in which East Africa receives much more rainfall compared to JF and JJAS seasons, the correlation coefficient of the two time series is -0.04 (significant 63 percent significance level), indicating not much of a correlation in the two time series. Precipitation in the range of -40 mm – 90 mm is associated with the NINO Index. In a few years, the higher area-averaged precipitation over East Africa is observed during the high positive NINO Index, such as, in 1950-51, 1963-64, 1982-84, and 1997-98. In 1953-54, 1976-77 and 1989-90, the high negative NINO Index corresponds to the lower area-averaged precipitation, while in 1973, 1983, and 1990-95, out of phase relationship could be seen in the area-averaged precipitation and NINO Index.

In most of the years, the time series of area-averaged PET anomaly is in phase with the NINO Index time series (Figure 5.34). The NINO Index correlates well (correlation coefficient = 0.48, significant at 99.99 percent significance level) with area-averaged PET anomaly. PET anomaly in the range of -48 mm – 12 mm is associated with the NINO Index variations. Out of phase relationship could be seen in area-averaged PET anomaly and NINO Index during 1957-59 and 1995-1999. Time series of area-averaged PET anomaly appears to have more related to NINO Index time series compared to that of precipitation anomaly.
Figure 5.33: Time series of NINO Index and area-averaged precipitation anomalies over East Africa. The first number above the figure indicates the correlation coefficient and the number in the parenthesis indicates the significance.

Figure 5.34: Time series of NINO Index and area-averaged PET anomaly over East Africa. The first number above the figure indicates the correlation coefficient and the number in the parenthesis indicates the significance.
5.3.1.2 Hydrology Variables

The NINO Index correlated negatively with surplus in JF to JJAS, but positively in OND (Figure 5.35). In JF, significant correlations are over south-equatorial East Africa, while in MAM, the significant correlations are over equatorial and southern East Africa. In JJAS, significant correlations are over northern East Africa, while in OND, significant correlations are over south equatorial East Africa. The negative correlation mean that positive (negative) NINO Index corresponds to low (high) surplus, while positive correlations mean positive (negative) NINO Index correspond to high (low) surplus over those regions.

Time series of area-averaged surplus anomaly and that of NINO Index is plotted in Figure 5.36. The correlation coefficient of the two time series -0.14 is significant at 99.95 percent significance level). Surplus anomaly in the range of -40 mm – 60 mm is associated with the NINO Index. Area-averaged surplus is in phase with NINO Index in many years, for example, peak in surplus follows high positive NINO Index in 1961-62, 1977-78, and 1997-98.

Figure 5.37 shows the correlation of deficit and NINO Index. In JF, correlations are mostly negative over northern East Africa to the north of 10S, while significant positive correlations are over southern east Africa. In MAM and JJAS, correlations are positive over most of East Africa, with significant correlations over equatorial and southern East Africa in MAM, and over northern East Africa in JJAS. In OND, negative correlations are over the Horn of Africa and southern equatorial East Africa between equator and 15S. Negative correlations over the Horn of Africa are significant at 99 percent significance level. Significant (at 95 percent and 99 percent significance level)
Figure 5.35: Spatial distribution of correlation of surplus with NINO Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
positive correlations are over northern and southern East Africa. Positive correlations mean that a positive (negative) NINO Index corresponds to higher (lower) deficit, while opposite is true for negative correlations. Time series of area-averaged deficit anomaly is plotted against NINO Index time series in Figure 5.38. The correlation of the two time series have coefficient of 0.13, significant at 99.89 percent significance level). Deficit anomalies in the range of -30 mm – 30 mm is associated with the NINO Index variations. Mid – monthly soil moisture correlated negatively over most of the East Africa to NINO Index in JF, MAM, and JJAS (Figure 5.39). In JF, significant correlations are over Horn of Africa, equatorial East Africa (95 percent significance level), and southern East Africa (99 5 – 99.9 percent significance levels). In MAM, significant negative correlations are over equatorial and southern East Africa. In JJAS, significant negative correlations are over north East Africa, while significant positive correlations are over Horn of Africa.
**Figure 5.37:** Spatial distribution of correlation of deficit with NINO Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
In OND, significant negative correlations are over northern East Africa, while significant positive correlations are over the Horn of Africa, coastal East Africa and equatorial East Africa. Positive correlations mean that positive (negative) NINO Index corresponds to higher (lower) soil moisture, while negative correlations suggests that positive (negative) NINO Index corresponds to lower (higher) soil moisture.

Figure 5.40 shows the time series of area-averaged mid-monthly soil moisture anomalies plotted against NINO Index time series. The correlation coefficient of the two time series is -0.13, significant at 99.81 percent significance level. Soil moisture anomalies in the range of -20 mm – 30 mm is associated with the NINO Index variations. The two time series are in phase, where peaks in area-averaged mid-monthly soil moisture anomalies follows high positive NINO Index peaks in many years, for example in 1961-62, 1968-69, 1982-83, and 1997-98.
Figure 5.39: Spatial distribution of correlation of mid-monthly soil moisture with NINO Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.40: Time series of NINO Index and area-averaged mid-monthly soil moisture over East Africa. The first number above the figure indicates the correlation coefficient and the number in the parenthesis indicates the significance.

5.3.1.3 EOF1

Table 5.1: Correlation coefficients of time series of first principal components of aridity index over East Africa and NINO Index time series for four seasons.

<table>
<thead>
<tr>
<th>Season</th>
<th>Correlation Coefficient (acr)</th>
<th>Correlation Coefficient (wac)</th>
</tr>
</thead>
<tbody>
<tr>
<td>JF</td>
<td>-0.339</td>
<td>0.435</td>
</tr>
<tr>
<td>MAM</td>
<td>-0.437</td>
<td>0.386</td>
</tr>
<tr>
<td>JJAS</td>
<td>-0.664</td>
<td>-0.743</td>
</tr>
<tr>
<td>OND</td>
<td>0.408</td>
<td>-0.022</td>
</tr>
</tbody>
</table>

NINO Index was also correlated to the time series of the first principal components (EOF1) of aridity index over East Africa. Table 5.1 shows the correlation coefficients thus obtained. In all the seasons the NINO Index correlates well with EOF1 of aridity index, except for EOF1 (wac) for OND. When the annual cycle in EOF1 is removed, the EOF1 time series and NINO Index are strongly negatively correlated, meaning that higher (lower) aridity index (aridity) occur during positive (negative) NINO
Index in JF, MAM, and JJAS, with the opposite relationship in OND. The strongest correlation is in JJAS.

5.3.2 Indian Ocean Dipole Mode (IODM) Index and Aridity

5.3.2.1 Aridity Index, Precipitation and PET

Indian Ocean SSTs are the important factor controlling the interannual climate variability over East Africa (Goddard and Graham, 1999; Mutai et al., 1998; Nicholson 1997; Ogallo, 1998). Positive (negative) IODM means that the SSTs over the western Indian Ocean are higher (lower) than SSTs over the eastern Indian Ocean. Figures 5.41, 5.42, and 5.43 show the correlations of aridity index, precipitation, and PET with the IODM Index. The correlation maps for aridity index and precipitation with the IODM Index are very similar for all seasons. PET over East Africa correlated positively with the IODM Index in all seasons.

In JF, low positive correlations are observed over central East Africa with significant positive correlations between the latitudes of 7°S and 12°S. Most of the Horn of Africa showed weak negative correlations. In MAM, significant positive correlations are concentrated in a very small region centered at 5°N and 40°E. The IODM does not seem to affect precipitation and aridity index much in JF and MAM. Indian Ocean SSTs seem to affect PET over southern central East Africa (Figure 5.43) where significant negative correlations were observed in MAM. This means that during positive (negative) mode of the dipole, PET (precipitation) is decreased (increased) over the southern inland region of East Africa. The PET is reduced by a higher rate compared to the increase in precipitation, which could cause higher aridity over southern East Africa during JF.
In JJAS, significant positive correlations are observed over the Horn of Africa, the coastal region and a few small regions over southern East Africa for aridity index (Figure 5.41) and precipitation (Figure 5.42). Low negative correlations were observed over northern East Africa. PET also correlated significantly positively over the coastal region and over central equatorial East Africa (Figure 5.43). The southwest monsoonal winds with the well-developed Somali jet penetrate inland over the Horn of Africa and coastal east Africa, bringing moisture and precipitation. This means that during the positive (negative) mode of the dipole, precipitation and PET increases (decreases), but the higher aridity could result as the rate of increase in PET is higher compared to that in precipitation. In JJAS, significant negative correlations with precipitation and the aridity index were observed over northwest East Africa and over Ethiopia. The inland penetrating Somali jet may not bring moisture far inland as it turns northwestwards. The dry Sahara winds prevail over northwest part of East Africa, which may explain the negative correlations over the region. PET did not show much relation with the IODM Index over northwest East Africa.

In OND, significant positive correlations with aridity index and precipitation are observed over equatorial Africa, while significant negative correlations are observed over southern and northwest East Africa. PET did not show much correlation with IODM Index in OND. During the positive IODM, the anomalous easterly winds cause above normal rainfall over equatorial East Africa, which may explain the lower aridity over the region of positive correlations.

5.3.2.2 Hydrology Variables

The IODM Index was also correlated with hydrology variables. Figure 5.44
Figure 5.41: Spatial distribution of correlation of aridity index with IODM Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.42: Spatial distribution of correlation of mid-monthly precipitation with IODM Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.43: Spatial distribution of correlation of PET with IODM Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
shows the correlations of IODM Index with surplus. Negative correlations are observed over most of the region in MAM and JJAS and positive correlations over most of the region in JF; but the correlations are not significant. In OND, Significant positive correlations are over Equatorial East Africa, the Horn of Africa, and over coastal East Africa, while the northwestern and southern regions of East Africa have negative correlations with the IODM Index. A positive correlation means that positive (negative) IODM Index corresponds to higher (lower) surplus, while a negative correlation means that positive (negative) IODM Index corresponds to lower (higher) surplus. IODM Index is not strongly associated with surplus much from January to September, but is associated strongly with surplus in OND.

Figure 5.45 shows the correlations of IODM Index with deficit over East Africa. In JF, the correlations are positive over most of the region, except over northern East Africa to the north of 10°N and over Somalia, where correlations are negative. In JF, correlations are not significant. In MAM, positive correlations are over the equatorial East Africa, while negative correlations are over the southern and northern East Africa including the Horn of Africa. Significant positive correlations are over a small region in central equatorial East Africa, along the equator. In JJAS, negative correlations are over Horn of Africa and coastal East Africa, where significant correlations are over equatorial coastal East Africa. In OND, significant positive correlations are over northwestern and southern (to the south of 15°S) East Africa, while remaining regions have negative correlations. Significant at 95 percent level negative correlations are over southern equatorial East Africa. Positive correlations mean that a positive (negative) IODM Index corresponds to the higher (lower) deficit, while negative correlations mean that positive
Figure 5.44: Spatial distribution of correlation of surplus with IODM Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
(negative) IODM Index corresponds to lower (higher) deficit. The IODM Index influences deficit over East Africa significantly in OND.

Figure 5.46 shows the correlations of IODM Index with mid-monthly soil moisture. Similar to the correlation pattern with surplus and deficit, IODM Index influences soil moisture significantly in OND only. In OND significant negative correlations are over northwestern and southern East Africa, while significant positive correlations are over equatorial East Africa and the Horn of Africa. In other seasons, correlations are not significant and are negative over most of the regions of East Africa.

5.3.2.3 EOF1

The IODM Index was also correlated to the time series of EOF1 over East Africa. Table 5.2 shows that correlation coefficients are negative in all seasons. The highest correlation is in JJAS.

Table 5.2: Correlation coefficients of time series of first principal components of the aridity index over East Africa and the IODM Index time series by season.

<table>
<thead>
<tr>
<th>Season</th>
<th>Correlation Coefficient (acr)</th>
<th>Correlation Coefficient (wac)</th>
</tr>
</thead>
<tbody>
<tr>
<td>JF</td>
<td>-0.006</td>
<td>-0.031</td>
</tr>
<tr>
<td>MAM</td>
<td>0.099</td>
<td>-0.004</td>
</tr>
<tr>
<td>JJAS</td>
<td>0.058</td>
<td>-0.248</td>
</tr>
<tr>
<td>OND</td>
<td>0.289</td>
<td>-0.027</td>
</tr>
</tbody>
</table>

5.3.3 NAO Index

5.3.3.1 Aridity Index, Precipitation and PET

Atlantic Ocean SSTs are shown to be the important factor controlling the interannual climate variability over East Africa. Figures 5.47, 5.48 and 5.49 show the correlations of aridity index, precipitation and PET respectively; with the NAO index.
Figure 5.45: Spatial distribution of correlation of deficit with IODM Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.46: Spatial distribution of correlation of mid-monthly soil moisture with IODM Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
The correlations maps are very similar for all seasons. PET over East Africa correlated negatively with the NAO index in all seasons over most of East Africa.

In JF, significant negative correlations with the aridity index and precipitation are observed over northern East Africa, while positive correlations were observed over southern East Africa. PET is significantly negatively correlated over the Horn of Africa and also over western East Africa, and positively correlated over northeastern (Ethiopia, southern Sudan, and Kenya) East Africa. Over Ethiopia, southern Sudan, and Kenya during the positive (negative) phase of NAO, precipitation and PET are decreased, when aridity is increased. During JF, western East Africa is out of phase with Eastern East Africa in terms of the correlation of NAO index with aridity index, precipitation and especially with PET. McHugh and Rogers (2001) have shown that precipitation over East Africa between the equator and 20oS decreases during the negative phase of NAO during Austral summer. Figure 5.50 agrees with this over most of the central and eastern region between equator and 20oS during JF.

In MAM, negative correlations are observed over central equatorial East Africa while positive correlations are observed over northern and southern East Africa. PET correlates negatively with NAO index over most of the East Africa during MAM.

NAO does not correlate with aridity index and precipitation over most of northern (southern) East Africa during JJAS (OND), while PET correlated negatively over most of East Africa during these two seasons. Aridity index and precipitation show low negative correlation with the NAO index over southern East Africa in JJAS and over equatorial Africa in OND.

The time series of area-averaged precipitation and PET are plotted against IODM
Figure 5.47: Spatial distribution of correlation of aridity index with NAO Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
**Figure 5.48**: Spatial distribution of correlation of precipitation with IODM Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.49: Spatial distribution of correlation of PET with IODM Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Index in Figures 5.50 and 5.51, respectively. The correlation coefficient of area-averaged precipitation time series with IODM index time series is -0.11 and is significant at 99.41 percent significance level. Precipitation range of -60 mm – 90 mm is associated with NAO index. The correlation coefficient of area-averaged PET correlated with time series of NAO index is -0.1 and is significant at 98.76 percent significance level. The range of -16 mm – 12 mm of PET is associated with NAO index. Anti-phase relationship could be seen in many years in the two time series.

5.3.3.2 Hydrology Variables

Correlation analysis was performed over NAO index and hydrology variables. Figure 5.52 shows the correlations of NAO index with surplus over East Africa. The NAO correlates negatively with surplus over most of the East Africa in all seasons.

Figure 5.50: Time series of area-averaged precipitation anomalies and NAO Index. The first number above the figure indicates the correlation coefficient and the number in the parenthesis indicates the significance.
Overall, the NAO does not influence surplus, as not many of the significant correlations could be seen. In JF, correlations significant at the 95 percent significance level are over southern East Africa. In MAM, significant correlations are over a small region in western equatorial East Africa. In OND, significant correlations are over small regions scattered across East Africa.

Figure 5.53 shows that the NAO Index does not seem to be strongly related to deficit. Positive correlations at significant at the 95 percent level are over equatorial coastal East Africa in JF and JJAS. Figure 5.54 shows the correlations of the NAO index with mid-monthly soil moisture. Significant negative correlations are seen over small pockets scattered across East Africa. Central East Africa correlated relatively significantly with the NAO index in JF, JJAS, and OND.

5.3.3.3 EOF1

The NAO index was also correlated to the time series of the first principal
Figure 5.52: Spatial distribution of correlation of surplus with NAO Index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.53: Spatial distribution of correlation of deficit with NAO index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.54: Spatial distribution of correlation of mid-monthly soil moisture with NAO index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
components (EOF1) of aridity index over East Africa (Table 5.3). The two variables did not show much of a correlation, with positive relationship during JJAS (EOF1-acr) and MAM (EOF1-wac). The highest correlation was found in JF.

**Table 5.3:** Correlation coefficients of time series of first principal components of aridity index over East Africa and NAO index time series for four seasons.

<table>
<thead>
<tr>
<th>Season</th>
<th>Correlation Coefficient (acr)</th>
<th>Correlation Coefficient (wac)</th>
</tr>
</thead>
<tbody>
<tr>
<td>JF</td>
<td>-0.162</td>
<td>-0.135</td>
</tr>
<tr>
<td>MAM</td>
<td>-0.038</td>
<td>0.064</td>
</tr>
<tr>
<td>JIAS</td>
<td>0.079</td>
<td>-0.002</td>
</tr>
<tr>
<td>OND</td>
<td>-0.085</td>
<td>-0.088</td>
</tr>
</tbody>
</table>

### 5.3.4 Correlating TASD Index with Aridity Index:

#### 5.3.4.1 Aridity Index, Precipitation and PET

A positive (negative) TASD index means that the SSTs over northern tropical Atlantic are higher (lower) compared to southern Tropical Atlantic Ocean. Figures 5.55, 5.56, and 5.57 show the correlations of aridity index, precipitation, and PET respectively, with the TASD index. The maps of correlations for aridity index and precipitation with TASD index are very similar for all seasons. PET over coastal Mozambique correlated significantly positively and positively over Congo with the TASD index in all seasons.

In JF, aridity index and precipitation are weakly positively correlated over most of northern East Africa except over Somalia and the eastern Central African Republic. Significant positive correlations were observed over Ethiopia and Uganda, meaning that a positive (negative) TASD index is associated with higher (lower) precipitation and lower (higher) aridity over this region. PET does not show much of an association with the TASD index over this region. PET is significantly positively correlated with the TASD
Figure 5.55: Spatial distribution of correlation of aridity index with TASD index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Index over eastern Ethiopia, Congo and coastal Mozambique, where aridity index and precipitation correlated weakly negatively with the TASD index.

In MAM, Aridity index and precipitation do not show much of an association with the TASD index. PET correlated positively with the TASD index over East Africa with significant positive correlations over Congo, southern Sudan, coastal Kenya, and coastal Mozambique.

In JJAS, the aridity index and precipitation is significantly positively correlated over northwestern East Africa, where PET correlated negative with TASD index. Over this region aridity could have an influence of northern tropical Atlantic SSTs through higher precipitation and lower PET. Aridity over other regions of East Africa does not show much of an association with the TASD index.

In OND, aridity index and precipitation correlated negatively with TASD over equatorial East Africa and Horn of Africa, while northern and southern East Africa correlated positively.

5.3.4.2 Hydrology Variables

The TASD index does not correlate well with surplus (Figure 5.58), but the significant correlations strongly mostly positive. In JF, small regions of significant correlations are over the central East Africa. In MAM, southern coastal East Africa show weak negative correlations, while the rest of East Africa correlates weakly positively with surplus. In JJAS, positive correlations are over northern East Africa, while central East Africa and the Horn of Africa region correlated negatively with surplus.

Figure 5.59 shows that the TASD index correlates mostly negatively with deficit in JF, except northwestern East Africa and the Horn of Africa, where correlations are
**Figure 5.56**: Spatial distribution of correlation of precipitation with TASD index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.57: Spatial distribution of correlation of PET with TASD index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
positive. A small region over the Horn of Africa correlates significantly positively at the 95 percent significance level with deficit, while small regions are scattered over north-central East Africa show significant (at 95 percent significance level) negative correlations. In MAM and JJAS, significant correlations are scattered over small regions over East Africa. In OND, negative correlations are over northwestern East Africa and southern East Africa, while positive correlations are over central East Africa and the Horn of Africa. Significant negative correlations at the 95 percent significance level are over north-central East Africa, while significant positive correlations are over west-central East Africa.

Figure 5.60 shows the correlations of the TASD index with mid-monthly soil moisture. The TASD index correlates negatively over most of East Africa in all seasons. Correlations are largely insignificant much in JF, MAM, and JJAS. Significant positive correlations, at 95 percent significance level are over equatorial western and southwestern East Africa in JF, while significant positive correlations (at 95 percent significance level) are over northern East Africa. In OND, significant positive correlations (at 95 percent significance level) are over northern East Africa and small regions over southern East Africa, while significant negative correlations (at 95 percent significance level) are seen over equatorial East Africa.

5.3.4.3 EOF1

The TASD index is also correlated to the time series of the first principal components (EOF1) of the aridity index over East Africa. Table 5.4 shows that the two variables do not show a strong association. The highest correlation is found in JJAS.
Figure 5.58: Spatial distribution of correlation of surplus with TASD index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.59: Spatial distribution of correlation of deficit with TASD index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.60: Spatial distribution of correlation of mid-monthly soil moisture with TASD index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Table 5.4: Correlation coefficients of time series of first principal components of aridity index over East Africa and NAO index time series for four seasons.

<table>
<thead>
<tr>
<th>Season</th>
<th>Correlation Coefficient (acr)</th>
<th>Correlation Coefficient (wac)</th>
</tr>
</thead>
<tbody>
<tr>
<td>JF</td>
<td>0.174</td>
<td>-0.016</td>
</tr>
<tr>
<td>MAM</td>
<td>-0.028</td>
<td>0.065</td>
</tr>
<tr>
<td>JJAS</td>
<td>0.233</td>
<td>0.219</td>
</tr>
<tr>
<td>OND</td>
<td>-0.189</td>
<td>0.120</td>
</tr>
</tbody>
</table>

5.3.5 QBO

5.3.5.1 Aridity Index, Precipitation, and PET

QBO index is correlated with aridity index (Figure 5.61), precipitation (Figure 5.62) and with PET (Figure 5.63). The influence of QBO on precipitation receipt over East Africa is shown in many studies (Rodhe and Virji, 1976; Ogallo, 1982; Nicholson and Entekhabi, 1986; Ogallo et al., 1994, Indeje and Semazzi, 2000). QBO is shown to have influence on the NINO Index (Nicholson and Entekhabi, 1986) and the IODM Index (Indeje and Semazzi, 2000), which influence precipitation receipt over East Africa. However, the QBO does not correlate well with aridity index, precipitation, and PET. Correlation patterns with aridity index and with precipitation are very similar. The significant correlations (significant at 95 percent and 99 percent significance level) with the aridity index and precipitation are observed in MAM over southern East Africa. The QBO does not correlate with PET over East Africa (Figure 5.63).

5.3.5.2 EOF1

The QBO Index is correlated with EOF1 (acr) and EOF1 (wac). Table 5.5 shows the correlation coefficient of QBO index and EOF1. QBO does not correlate well either with EOF1 (acr) or with EOF1 (wac) in any season. The strongest correlation coefficient is 0.138 with EOF1 (acr) and is -0.261 with EOF1 (wac).
Figure 5.61: Spatial distribution of correlation of aridity index with QBO index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
Figure 5.62: Spatial distribution of correlation of precipitation with QBO index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90percent, 95percent and 99.9percent significance levels, respectively.
**Figure 5.63**: Spatial distribution of correlation of PET with QBO index for four seasons. Contours of equal correlation coefficient are drawn at the interval of 0.1. Lighter to darker gray shades indicate correlations at 90 percent, 95 percent and 99.9 percent significance levels, respectively.
Table 5.5: Correlation coefficients of QBO index calculated with EOF1 (annual cycle removed) and with EOF1 (with annual cycle).

<table>
<thead>
<tr>
<th>Season</th>
<th>Correlation Coefficient (acr)</th>
<th>Correlation Coefficient (wac)</th>
</tr>
</thead>
<tbody>
<tr>
<td>JF</td>
<td>0.141</td>
<td>-0.156</td>
</tr>
<tr>
<td>MAM</td>
<td>0.029</td>
<td>-0.261</td>
</tr>
<tr>
<td>JJAS</td>
<td>-0.023</td>
<td>-0.076</td>
</tr>
<tr>
<td>OND</td>
<td>0.138</td>
<td>0.101</td>
</tr>
</tbody>
</table>

5.3.5.3 The Other Climatic Indices

Because of the influence of the QBO index over the NINO and IODM Index, QBO index was correlated with the other climatic indices. The correlation coefficients are shown in Table 5.6, but these correlations are not very significant.

Table 5.6: Correlation coefficients of QBO index with NINO Index, IODM Index, NAO index and TASD index.

<table>
<thead>
<tr>
<th>Seasons</th>
<th>NINO Index</th>
<th>IODM Index</th>
<th>NAO Index</th>
<th>TASD Index</th>
</tr>
</thead>
<tbody>
<tr>
<td>JF</td>
<td>-0.03</td>
<td>-0.021</td>
<td>0.084</td>
<td>0.042</td>
</tr>
<tr>
<td>MAM</td>
<td>-0.096</td>
<td>-0.059</td>
<td>-0.2</td>
<td>0.069</td>
</tr>
<tr>
<td>JJAS</td>
<td>-0.061</td>
<td>0.039</td>
<td>0.02</td>
<td>0.237</td>
</tr>
<tr>
<td>OND</td>
<td>0.032</td>
<td>0.276</td>
<td>0.006</td>
<td>0.067</td>
</tr>
</tbody>
</table>

5.4 Time Series Analysis For NDVI and PDSI

Time series of area-averaged NDVI and PDSI over East Africa are correlated with the time series of area-averaged hydrology variables over East Africa. Both NDVI and PDSI time series correlated significantly with hydrology variables at a significance level of 99 percent and above.

5.4.1 NDVI Time Series

Time series of area-averaged NDVI correlates well with time series of
precipitation anomaly over East Africa (Figure 5.64). The correlation coefficient 0.19 is significant at 99.49 percent significance level. Precipitation range of -40 mm – 90 mm is associated with NDVI over East Africa.

![Figure 5.64](image)

**Figure 5.64**: Time series of area-averaged precipitation anomalies over East Africa plotted against time series of area-averaged NDVI over East Africa. The first number above the figure indicates the correlation coefficient and the number in the parenthesis indicates the significance.

NDVI time series did not correlate well with the time series of area-averaged PET over East Africa (Figure 5.65). The correlation coefficient 0.09 is significant only at a 81.14 percent significance level. However the time series of NDVI correlated well with time series of area-averaged surplus (Figure 5.66). Correlation coefficient of 0.21 is significant at 99.82 percent significance level. Surplus range of -40mm – 60 mm is associated with NDVI.

The time series of NDVI correlates well with the time series of area-averaged deficit over East Africa (Figure 5.67). The correlation coefficient of -0.17 is significant at
**Figure 5.65:** Time series of area-averaged PET anomalies over East Africa plotted against time series of area-averaged NDVI over East Africa. The first number above the figure indicates the correlation coefficient and the number in the parenthesis indicates the significance.

**Figure 5.66:** Time series of area-averaged surplus anomalies over East Africa plotted against time series of area-averaged NDVI over East Africa. The first number above the figure indicates the correlation coefficient and the number in the parenthesis indicates the significance.
Figure 5.67: Time series of area-averaged deficit anomalies over East Africa plotted against time series of area-averaged NDVI over East Africa. The first number above the figure indicates the correlation coefficient and the number in the parenthesis indicates the significance.

Figure 5.68: Time series of area-averaged mid-monthly soil moisture anomalies over East Africa plotted against time series of area-averaged NDVI over East Africa. The first number above the figure indicates the correlation coefficient and the number in the parenthesis indicates the significance.
a 98.81 percent significance level. The deficit range of -30 mm – 30 mm is associated with NDVI. Deficit and NDVI are out of phase in most of the time periods.

The time series of NDVI correlates with time series of area-averaged mid-monthly soil moisture has correlation coefficient of 0.21, significant at a 99.82 percent significance level (Figure 5.68). Soil moisture depth in the range of -40 mm – 60 mm is associated with NDVI.

5.4.2 PDSI Time Series

PDSI correlates well with the time series of area-averaged precipitation over East Africa (Figure 5.69). The correlation coefficient of 0.42 is significant at a nearly 100 percent significance level. The time series of PDSI and precipitation are in phase for most of the time period. The precipitation anomaly in the range of -60 mm – 80 mm is associated with the PDSI.

![Figure 5.69](image_url)

**Figure 5.69:** Time series of area-averaged precipitation anomalies over East Africa plotted against time series of area-averaged PDSI over East Africa. The first number above the figure indicates the correlation coefficient and the number in the parenthesis indicates the significance.
PDSI correlates well with time series of area-averaged PET over East Africa (Figure 5.70). The correlation coefficient of -0.39 is significant at nearly 100 percent significance level. PET and PDSI are out of phase in most of the time period. PET anomaly in the range of -16 mm – 12 mm is associated with PDSI.

![Figure 5.70](image)

**Figure 5.70**: Time series of area-averaged PET anomalies over East Africa plotted against time series of area-averaged PDSI over East Africa. The first number above the figure indicates the correlation coefficient and the number in the parenthesis indicates the significance.

Time series of area-averaged surplus correlated well with time series of PDSI (Figure 5.71). Correlation coefficient of 0.5 is significant at 100 percent significance level. PDSI and surplus is in phase over most of the time periods. Surplus range of -40 mm – 60 mm is associated with the PDSI.

The time series of PDSI correlates significantly with that of area-averaged deficit over East Africa (Figure 5.72). The correlation coefficient of -0.57 is significant at a nearly 100 percent significance level. An out of phase relationship between the two time
Figure 5.71: Time series of area-averaged surplus anomalies over East Africa plotted against time series of area-averaged PDSI over East Africa. The first number above the figure indicates the correlation coefficient and the number in the parenthesis indicates the significance.

Figure 5.72: Time series of area-averaged deficit anomalies over East Africa plotted against time series of area-averaged PDSI over East Africa. The first number above the figure indicates the correlation coefficient and the number in the parenthesis indicates the significance.
series is evident over most of the time period. Deficit anomaly in the range of -30 mm – 20 mm is associated with the PDSI.

Finally, the time series of PDSI correlated well also with that of area-averaged mid-monthly soil moisture over East Africa (Figure 5.73). The correlation coefficient of 0.69 is significant at nearly a 100 percent significance level. The in phase relationship of the two variables is evident in Figure 5.73. Soil moisture depth anomaly in the range of -20 mm – 30 mm is associated with the PDSI.

**Figure 5.73**: Time series of area-averaged mid-monthly soil moisture anomalies over East Africa plotted against time series of area-averaged PDSI over East Africa. The first number above the figure indicates the correlation coefficient and the number in the parenthesis indicates the significance.
Chapter 6: Conclusions

Precipitation variability, both intra-annual and interannual is well-studied. Dryness and desertification over African regions such as northern Africa, Saharan Africa, and sub-Saharan Africa regions are also well studied. Such studies are lacking over East Africa. East Africa includes semiarid regions with high population, population growth rate, and poverty with high risk of desertification. East Africa has experienced frequent droughts and floods and the study of East African aridity is thus important.

Interannual variability of hydrological variables over East Africa such as precipitation, potential evapotranspiration (PET), water surplus, deficit, and soil moisture were studied with reference to interannual variability of the aridity index, where aridity index is calculated as the ratio of mean seasonal precipitation and mean seasonal PET. Principal components analysis was performed on the aridity index. Correlations of aridity index, first principal component of aridity index with hydrology variables, selected meteorological variables, and selected climatic indices were analyzed.

The aridity index over East Africa is driven by precipitation more than PET. The spatial distribution of correlation coefficients of the aridity index and precipitation are greater than 0.6 at 99.99 percent significance level over most of the East Africa in all seasons, which is not the case with such correlations with PET. The regions over which such significant (at 99.99 percent significance level) correlations are observed, follows the ITCZ, indicating that precipitation plays a significant role in determining PET over East Africa. It was also observed that over East Africa, the magnitude of change in area-averaged precipitation is much greater than the magnitude of change in area-averaged evapotranspiration.
The first principal component explained 47.46 percent of total variance when annual cycle in the aridity index was kept intact before performing PCA (EOF (wac)), while the first principal component explaining 14.82 percent of total variance was yielded when annual cycle in the aridity index was removed before performing PCA (EOF (acr)). These EOF1 (acr) and EOF1 (wac) were used for further analysis. Spatial patterns of correlations of EOF1 (acr) with hydroclimatic variables represent the association of those variables with interannual variability of the aridity index well compared to that using EOF1 (wac). Use of EOF1 (wac) in analysis can not be ignored because of its ability to explain a high portion of total variance.

EOF1 correlated well with all hydrological variables. High positive correlations with surplus were observed over the regions receiving rainfall in each season, while high negative correlations were observed over dry regions in each season. EOF1 correlated well with global SSTs. Indian Ocean SSTs correlated well especially in JJAS. SSTs in southern Atlantic and in southeastern Atlantic also correlated significantly positively in JJAS. Surface air temperature did not correlate well with EOF1. This could be because of the higher elevation of most of the East Africa regions. The surface temperatures are in the range of 20° – 30°C over East Africa in a typical year. The precipitation influence on PET is more compared to that of temperature.

The Palmer Drought Severity Index (PDSI) is an important measure of dryness. Available PDSI data East Africa has 2.5° x 2.5° latitude/longitude resolution. Because of the coarse resolution of PDSI data, aridity index was used as a measure of dryness. Available PDSI data was correlated with EOF1 and hydrology variables over East Africa and it correlated well with EOF1 as well as with hydrology variables. Time series plotted
against area averaged hydrology variables indicated that PDSI is in phase with precipitation, surplus, and soil moisture, while is out of phase with PET and deficit. Normalized difference vegetation index (NDVI) is also well correlated with EOF1. NDVI is associated with precipitation variations and thus time series of NDVI plotted against hydrology variables indicated in phase association with precipitation, surplus and soil moisture and out of phase relation with PET and deficit.

The influence of ENSO, IODM, and Atlantic Ocean SSTs on the interannual variability of precipitation over East Africa is also well-studied. Climatic indices such as NINO (3.4 region) Index, IODM index, NAO index and TASD index, and QBO index were correlated with aridity index, EOF1 and with hydrology variables. Spatial distribution of correlations of these indices with the aridity index, EOF1, and hydrology variables indicated the influence of the climatic indices on these variables in each season. Among these climatic indices the most influential index is the NINO index. NINO index influenced aridity index and hydrology variables in each season. The NINO index has significant positive correlation with precipitation and PET. El Niño is associated with increase in precipitation and PET, while La Niña is associated with decrease in precipitation and PET and with increase in dryness and aridity for East Africa.

IODM is shown to influence precipitation over East Africa especially in JJAS. IOD index correlated well with aridity and precipitation in JJAS and OND seasons, while correlated well with PET in JF and JJAS seasons and correlated well with remaining hydrology variables only in OND seasons. Association of NAO index with East Africa precipitation is shown in previous studies. Correlation of NAO index on aridity is not well-defined. NAO index is well-correlated with aridity index and precipitation in JF and
JJAS. NAO index did not show much of an association with other hydrology variables. Influence of tropical Atlantic Ocean on East Africa precipitation variability is not well-studied. TASD index correlates with aridity index and precipitation in JF and JJAS, and with PET in MAM. Influence of TASD index on other hydrology variables is not well defined. TASD index did not correlate with surplus, but correlated well with deficit and soil moisture in OND. QBO index did not correlate much with aridity and hydrology variables. QBO has been shown to influence ENSO and IOD through Pacific and Indian Ocean branches of the Walker Circulation, and thus it influences interannual variability of East African rainfall. The QBO index thus was expected to influence interannual variability of aridity over East Africa. QBO index is not well-correlated with EOF1 or with other climatic indices.

In summary, aridity index is driven by precipitation rather than PET and thus teleconnections influencing interannual variability in East Africa rainfall also influence interannual variability of aridity.
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Vita

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