CHAPTER ONE
INTRODUCTION

1.1 Background to the study

The sub-Saharan Africa in general and West Africa in particular had been experiencing pronounced anomalies in hydrometeorological parameters, particularly rainfall, since the beginning of the 20th centuary. The pattern of the rainfall variability since 1901 in the region shows that the 20th century begins with relatively long period of dry climatic conditions which are fairly persistent until about 1926 with occasional break of normal or wet condition. This is followed by a relatively wet period which lasted until about 1960. From about 1964, droughts have been relatively persistent in West Africa and particularly since 1969 reaching a peak in 1973, leading to a widespread crop failure and its attendant consequences across the region. However, more detailed analysis shows a lot of spatial and temporal variations in the characteristics of rainfall since 1901, so that it becomes difficult to generalize for relatively large areas, using information from a relatively small area.

Several studies have been carried out on the characteristics of climatic change and climatic variability in Africa (Grove, 1972; Lamb, 1980, 1985; Nicholson, 1981, 1983; Keer, 1985; Ojo, 1986; Nicholson and Grist, 2001; Rowell, 2001; Grist and Nicholson, 2001; Le-Barbe *et al*, 2002; Taleb and Druyan, 2003; Odekunle, 2004; Nicholson, 2005; Adejuwon and Adekunle, 2006; Sivakumar, 2007; Nicholson and Webster, 2007; Omotosho and Abiodun, 2007; Odekunle and Eludoyin, 2008; Mortimore, 2010; Druyan, 2011; Ekpoh and Nsa, 2011; Pu and Cook, 2012; Munemoto and Tachibana, 2012; Navarra *et al*, 2013; Nicholson, 2013). However, the results of these studies show divergence of opinion about the nature and characteristics of climatic variability and climatic change in the continent. Thus, many different controversial statements on the status of the characteristics and the possible trends of climate in the region are made by several scientists. For example, some scientists conclude that the Sahelian drought of West Africa persists since 1969 and would even persist into the next century, thus indicating a phenomenon of climatic change to drought conditions (Lamb, 1973; Winstanley, 1973a; Shukla, 1995; Janicot, 1996; Hulme, 2000; Nicholson, 2000; Dai and Wigley, 2000).

According to Lamb (1973), the Sahelian droughts “have already a long history. It is not likely to disappear in the near future”. Winstanley (1973b) argued that the downward trend in rainfall in West Africa would continue for 50 years. In contrast to these conclusions that the Sahelian drought would persist for some time, some scientists regard the drought condition as being somewhat unusual in terms of the recent past but not necessarily deviating from the longer term probabilities. In fact, to these scientists, the recent Sahelian droughts are part of the normal climate rather than an indication of climatic change. For example, Landsberg (1975), Bunting *et al* (1976) and Ogallo (1979) conclude that the Sahelian drought of the 1970s has to be accepted as part of the normal climate of the region. The available data do not indicate a trend and there is no indication of climatic change. These divergent opinions on drought characteristics in African continent and other parts of the tropics underscore the need for further research on the dynamics of tropical rainfall anomalies in order to better understand the pattern of rainfall anomalies in the tropical areas.

1.2 Aim and objectives of the study.

The study aims at examining the pattern of rainfall anomalies in Nigeria from 1901-2000 with a view to understanding the possible causal factors.

Specifically, the objectives of the study are stated as follows:

1. To examine the frequency and spatial pattern of drought in Nigeria from 1901-2000.
2. To determine the relationships among El-Nino/Southern Oscillation (ENSO), Sea Surface Temperatures (SSTs) anomalies and Nigerian rainfall.
3. To assess the influence of the Inter-tropical Discontinuity on the pattern of rainfall distribution in Nigeria.
4. To determine the trends and variability in annual rainfall and rain days in Nigeria over the period 1901-2000.

1.3 Hypotheses.

The hypotheses are stated as follows:

1. There has been an increasing intensity of drought in Nigeria over the period of the study.
2. There is significant relationship between rainfall anomalies in Nigeria and El-Nino/Southern Oscillation (ENSO) and Sea Surface Temperature anomalies (SSTAs).
3. The observed space-time variations in rainfall over Nigeria could be explained as a function of the seasonal movements of the Inter-Tropical Discontinuity over Nigeria.

4. There is significant upward trend in annual rainfall and rain days in Nigeria over the period 1901-2000.

1.4 Statement of the research problem

Despite the global concern for increasing frequency and intensity of extreme weather and climate events, their impacts are most devastating in developing countries than in the developed countries partly due to the weaker adaptation and mitigation options in the former. The loss of lives, displacement of persons and destruction of infrastructure resulting from rainfall anomalies in Nigeria in the last 100 years or so are unprecedented; hence the need to examine these phenomena. This work addressed the following questions:

1. What are the causal factors of annual rainfall anomalies in Nigeria?
2. To what extent does drought incidence vary in space and time over Nigeria?
3. How does annual rainfall pattern changed in Nigeria over the period 1901-2000?

1.5 Significance of the study.

The work examines the dynamics of Nigerian rainfall climatology with its attendant problems of rainfall anomalies particularly droughts, in order to aid decision making in policy analysis and implementation with a view to combating or mitigating the effect of drought if it reoccurs. This is through a better understanding of the nature and causal factors of rainfall anomalies in Nigeria. It is hoped that in this work, greater insight and better understanding will be gained in to the fact that the frequency and spatial pattern of drought phenomenon are important components of hydrometeorological disasters due to rainfall anomalies. The findings can be useful not only to the peasant farmers but also the agro-allied industries and agricultural planners and administrators in Nigeria for identifying related problems and devising policies and strategies that would help in avoiding further risk of crop failure associated with rainfall anomalies visa-vis droughts.

1.6 The Study Area: Nigeria.

Nigeria is approximately 923, 768 square kilometers territory that occupies the south-east corner of West Africa extending from Lat. 4^0N – Lat. 14^0N and Long. 2.5^0E – 14.5^0E. The greatest distance from east to West is approximately 1,300 kilometers and from north to south is about 1,046 kilometers. The country shares boundaries with the Republic of Benin in the West, Cameroon in the east, Niger and Chad Republic in the north and north- east respectively and the Gulf of Guinea in the South (see Fig. 1.1).

This space is for figure 1. 1

1.6.1 Relief

Nigeria can be divided into seven physiographic regions based on land units and common landform assemblages (Faniran, 1982; Akintola, 1982). The relief regions include the Creeks and Lagoons, the Niger Delta, the Coastal plains, the River basin troughs, the Inselberg landscapes, the Chad basin and the Eastern highland (Fig. 1.2). From a height of 1,500 meters to 1800 meters above sea level on the Jos Plateau, the land drops to under 15 meters above sea level over much of the delta area.

The Nigerian coastline is relatively straight except for the broad indented delta region that separates the eastern and western segments. From the coastline to about 10 kilometers inland, the mean altitude is about 30 meters above sea level. The Niger Delta consists of creeks and numerous tributaries through which the Niger empties into the sea. It extends from Forcados in the West to Bonny River in the east, a distance of about 350 kilometers, and from the apex of the delta at Abo to the coastline is about 150 kilometers (Fig 1.2). To the north of this region lies the gently undulating coastal plain of about 75 kilometers wide in the west but wider towards the east.

The elevation increases northwards with an average of about 150 meters above sea level. Nigeria's inselberg landscapes develop over the Basement Complex and extends from the northern part of Abeokuta – Ibadan–Ondo axis as far as the Niger valley with summits ranging from 300 meters to 600 meters above sea level. The Jos plateau itself is a much eroded volcanic relic that covers about 8,000 square kilometers with an average height of 1, 286 meters above sea level. The Biu plateau is found to the east of Jos Plateau reaching heights of over 900 meters above sea level. The entire landscape which stretches from Gumel through Nguru and Maiduguri as far as to Lake Chad, popularly called the Chad Basin, ranges from 300 meters to 340 meters above sea level.

Another notable region is the chain of hills along the eastern border known as the Bamenda highlands in the south and the Alantika mountains further north just south of the Benue River. In the northern part of the river Benue, there is a similar but much smaller range that culminates in the Mandara Mountain (Fig 1.2). There are many rivers which flow within the boundaries of Nigeria but the rivers with the broadest valleys are the Niger, Benue and Gongola rivers which flow in structural troughs. The Niger flows through a depression which in certain parts is more than 150 meters deep. In places where the river flows over sedimentary rocks, the trough is narrow.

However, the Benue trough is generally wide with the river flowing over sedimentary rocks for most of its course (Fig. 1.2)

This space is for figure 1.2

1.6.2 Geology

The geology of Nigeria reveals the rocks of various units of the geological succession range in age from the Precambrian to the Quaternary. The pre-Cambrian rocks are partly overlain by cretaceous sediments in some areas but lie directly on the pre-Cambrian in others. The pre- Cambrian rocks are referred to as the Basement Complex which constitute the oldest exposed metamorphic rock in Nigeria.

The basement complex comprises the remnants of an ancient sedimentary series –the meta- sediments –which transformed into anatetic, migmatites and granites. The meta-sediments include schist, quartzites, marbles, meta-conglomerates, amphibolites, and calci-silicate rocks. The older granites include basic and intermediate plutonic rocks, fine-grained granite and syntectonic granites. The younger granites, which consist of granites, granites porphyry Syenite, gabbro, rylite and others are now regarded as Jurrassic in age. These are found in Jos plateau, Kano, Zaria, Bauchi and Benue provinces where they form rugged hills (Fig. 1.3).

Furthermore, the Basement complex rocks are exposed from the Republic of Benin to Lokoja, from Ogoja to the Cameroon Highlands, and over more than half of the country north of the Niger and Benue Rivers. They are underlain by the sedimentary area to the south, north, east and north-west. The ground mass of undifferentiated metamorphic rocks consist of the migmatite complex found in Abeokuta, Ibadan, Ife and to the north of Minna. Part of the meta-sediments and metamorphic iron beds are found in Ibadan/ Ile-Ife area, and also between Kontagora and Zaria. The miscellaneous rock types including charnokites, pyroxones, syenites, pegmatite and dolerite dykes are found in the Basement complex areas north of Ibadan and among the metamorphic rocks of the Jos plateau. Ancient metamorphic rocks-the Basement complex-out crop very widely, intruded by various igneous masses and overlain in the north by continental sedimentaries of Tertiary age-the Chad Formation in the north-east and the Gundumi and Gwandu Formations in the north-west (Fig. 1.3). In the south, and also along the Niger –Benue –Gongola trough (Fig. 1.3) earlier marine sediments of lower and upper cretaceous age are deposited, constituting in the vicinity of the Niger Delta, the economically important structures wherein petroleum is formed and preserved (see Barbour *et al*, 1982). Beside the main rivers and also in the Delta and coastal areas, extensive alluvial deposits mask the underlying geological structure.

This space is for figure 1.3

1.6.3. Climate

The climate of Nigeria is more varied than those of any other country in West Africa. This is due to the fact that the distance from the south to the north of the country is very great (1150km) and thus covers many (virtually, all) of the climatic belts of West Africa (Udo, 1971; Iloeje, 1981). Nigeria receives rainfall from the south westerlies which invade the country from the Gulf of Guinea coast, i.e. the tropical Atlantic. This moist air stream is overlain by the northeast trades which originate from above the Sahara and therefore are dry and dust laden. The zone of contact of the two air masses at the surface is a zone of moisture discontinuity and it is known as the Inter-Tropical Discontinuity. The ITD advances inland as far as 22-25^0 N in August at the margin of the Sahara i.e. considerably beyond Nigeria's northern border (Adejokun 1964; Adedokun, 1978) while it does not retreat equatorward beyond 4^0N latitude during the 'Harmattan' dry season (see Adefolalu, 1983).

Five weather zones are associated with the dynamics of the ITD in Nigeria. These weather zones include the dry harmattan (Weather zone A), the the dry but humid weather (Weather zone B), the linesqualls (Weather zone C), the Monsoon weather (Weather zone D) and the little dry season (Weather zone E). Zone A weather lies to the north of the ITD and hence is rainless as well as zone B to the immediate south because they did not contain rain-producing clouds. Rainfall in the ITD occurs in zones C and D where conditions favor the development of clouds of great vertical extent. Thunderstorms and squall lines are associated with zone C weather and monsoon rains with zone D weather. Consequently, rainfall is spatially discontinuous when zone C weather prevails. On the other hand, the monsoon system gives continuous rains which may last 12 hours or more (see Olaniran, 1986, 1995). Overall, rainfall occurs at a distance of about 500km south of the surface location of the ITD, 4-6 weeks behind it in its annual cycle. When the fifth weather type associated with the ITD i.e. zone E, prevails over an area, light rainfall usually occur because Zone E weather is dominated by layered stratiform clouds.

Rainfall usually commences at the beginning of the rainy season from the coast (in the south), spreads through the middle belt, to eventually reach the northern part very much later. The converse of the situation also holds for the rainfall retreat period (see Ojo, 1977; Olaniran, 1986; Umar, 2010a). Generally, the northward advance of the rains is fairly gradual and steady over the whole country, whereas, the southward retreat is more rapid.

In Nigeria, based on the average monthly totals of rainfall, stations north of latitude 9^0N experience one rainfall peak in August while those places south of this latitude experience double maximum in July and September respectively with ‘little dry season’ in August generally called the ‘August break’. Rainfall decreases from the coast to the interior both in amount and in duration. Hence, the coastal areas are more humid all the year round and have the highest annual rainfall (Olaniran, 1987), while the northern part of the country has a shorter period of rain resulting in lower rainfall amounts (Fig. 1.4) and a longer dry season.

Along the coast, the annual total rainfall ranges from 4,295mm at Bonny in the east to about 1,775mm at Lagos in the west. Northwards, the mean annual rainfall decreases inland to less than 700mm in Lake Chad Basin (Maiduguri, 625mm). The latitudinal decrease in rainfall is interrupted by the effects of the relief in the interior. For instance, the Niger/Benue trough lying generally on the leeward side of the southern uplands receives less than average rainfall of what would be expected for its latitude. North of it is underlain by the Jos Plateau whose southern windward slopes receive much more rainfall than the latitudinal mean (Fig. 1.4).

This space is for figure 1.4

In general, Nigeria experiences high temperatures all the year round due to its latitudinal location within the tropics which have the highest amounts of net radiation on both the ocean and the land surfaces. Ayoade (2004) shows that net radiation values are higher over oceans than those over land surfaces within the low latitudes. This is because of the relatively higher albedo of the land surface in this area and the fact that land surfaces in this zone are predominantly deserts with little or no cloud cover. The absorbed solar radiation in this zone is therefore less and the effective outgoing radiation greater over land than over oceans. Poleward of 50^0 of latitude in both hemispheres, net radiation values over land and ocean surfaces are about the same (Ayoade, 2004).

The mean temperature for most stations is about 27^0C (Fig. 1.5). The highest air temperatures (34^0 - 40^0C) are normally in April and May in the northern part of the country. In the south the mean daily temperatures have the highest values (28^0 –30^0C) around March and their lowest values (23^0 –24^0C) around August. Generally, all over the country, the months of March, April and May show higher values of temperature than the rest of the year. The values of mean annual maximum temperatures are lower in the south (29^0 –32^0C) and increases northwards to 32^0C. However, the lowest values of mean annual maximum temperatures are found in Jos (27^0-30^0C). The mean annual minimum temperature has its lowest value of 18^0 C –19^0 C around Jos. The highest values (22^0 –23^0C) are found in the riverine areas of Edo, Delta, Bayelsa, Rivers, Akwa Ibom and cross River states and around the central and eastern parts of the Niger/Benue Valley (Adebayo, 1999).

This space is for figure 1.5

1.6.4. Vegetation.

The latitudinal extent of Nigeria from latitude 4^0N to 14^0N offers a wide variety of vegetation types representative of Africa between the equator and the southern fringes to the Sahara. There are two major types of vegetation in Nigeria: forest belt and savanna region. Each of these has many variants, affecting the floristic diversity and the structural appearance of the plant communities.

In broad terms, a zonal pattern of vegetation is observed in Nigeria changing from mangrove swamp in the south through forest-savanna mosaic in the middle belt to the Sahel savanna in the extreme north-east (Fig. 1.6). One can say categorically that the distribution of rainfall is the most important factor influencing the vegetation of Nigeria. The forest communities are subdivided into coastal/ mangrove forest, deltaic forest and moist Lowland/ tropical rain forest. Immediately after these we have the forest-savanna mosaic characteristic of the derived savanna zone. The savanna communities consist of the guinea savanna (mixed woodland), Sudan or grass savanna and the Sahel savanna. Another major type of vegetation is the montane forest found around Jos, Obudu and Mambila Plateaux (Fig. 1.6).

However, it is pertinent to submit that the present day vegetation types in Nigeria, are all anthropogenic derivatives of climax communities. For instance, the forest and derived savanna communities of the humid south are derivatives of the tropical rainforest. Likewise, the southern and northern guinea savanna are derived from tropical deciduous forest which developed in a climatic region characterized by a dominance of the humid over arid tropical conditions. Also the Sudan and Sahel savanna are derived from forest tropical xerophytic woodlands developed in a sub-humid to semi-arid climatic environment

This space is for figure 1.6

1. 6. 5. Land use.

The land use pattern in Nigeria is very complex, so that classification can only be at a very general level. Although some areas show distinct orientation towards particular types of agricultural enterprise, for most of the country there is no such clear-cut towards a single land use system (Areola, 1982). Thus, in every part of Nigeria, the people have adopted multiple land use system, growing variety of crops within the same locality.

In general, the land use systems in Nigeria can be correlated with the major eco-climate zones. Thus, the dominant land use systems in the humid forest region to the south are lumbering and root- and tree-crop production, while the sub-humid Sudan savanna belt to the north specialize in leguminous, cotton and grain crops. Between these zones, the derived and guinea savanna zones combine the grain and root crop economics, while they are seasonally used for grazing by the far-ranging Fulani cattle (Areola, 1982). Within each of these broad ecological zones, subdivisions have been recognized based on the intensity of land use, the mixture of crops and the land use system. These are in turn related to differences in soil, topography and the population distribution patterns.

The land use systems include those that are used for construction of transportation routes, building of cities, siting of industries among others. The most spectacular way in which man modifies local climate is by the construction of large cities. In Nigeria , such cities include the state capitals such as Lagos, Ibadan, Port Harcourt, Enugu, Calabar, Lokoja, Makudi, Ilorin, Abuja, Kaduna, Jos, Kano Katsina, Sokoto, Maiduguri to mention a few.

The process of city construction generally involves the modification of the natural environment by the clearance of natural vegetation, the reclamation of swampy areas, the construction of buildings, roads and concrete surfaces as well as open spaces like parks and pools, generally referred to as ‘urban surfaces’. These structures, coupled with the characteristic human activities in industry, transportation and waste disposal, together control the climate of the immediate environment.

Temperature is the most commonly measured climatic element which symbolizes changes in the heat generated by the space heating of buildings, motor vehicles and a mass of people located in small area (Oguntoyinbo, 1981). This is referred to as ‘the heat island effect’. The heat island phenomenon is best developed under conditions of low wind speed on calm night and it is associated with the highest density of urban dwelling. Building materials such as bricks, stones, steel, tarmac and concrete absorb more heat during the daytime than the vegetation in the countryside.

Thus, the thermal contrasts of any city depend on the nature of its landscape. The rapid and unplanned growth of Nigeria's major urban centers e.g Ibadan city has had great repercussions on the environment (Areola, 1994). The overall effects of all these factors are that; the city center experiences a less rapid drop in temperature than the countryside in the evening, high night temperature occurs in the city with the result that notably higher minimum temperatures are recorded in the city compared with the rural areas. The slash and burn system of agriculture that indiscriminates lumbering activities, incessant annual bush burning and land clearing for various developmental purposes leads to reduction in vegetal cover. This in turn leads to significant changes in land use systems. All these have effects on the Nigerian microclimatic environment.

CHAPTER TWO

CONCEPTUAL FRAMEWORK AND LITERATURE REVIEW

2.1. Southern Oscillation, El-Nino and La-Nina

In recent decades studies of the Southern Oscillation (SO), the El-Niño and La-Niña phenomena, and their associated effects over the entire atmosphere, have assumed greater economic and social importance and have been given considerable scientific attention. The term El-Nino is coined more than 100 years ago to describe the unusually warm waters that would occasionally form along the coast of Peru and Equador. This phenomenon typically occurs late in the calendar year near Christmas, hence the name El-Nino (Spanish for "the boy child" referring to the Christ child). Today the term El-Nino is used to refer to a much broader scale phenomenon associated with unusually warm water that occasionally form across much of the tropical eastern and central Pacific. The time between successive El-Nino events is irregular but they typically tend to recur every 3 to 7 years.

La-Nina is the counterpart to El-Nino and is characterized by cooler than normal SSTs across much of the equatorial eastern and central Pacific. A La-Nina event often, not always, follows an El-Nino and vice-versa. Once developed, both El-Nino and La-Nina events tend to last for roughly a year although occasionally they may persist for 18 months or more. El-Nino and La-Nina are both normal part of the earth's climate and there are recorded evidence of their having occurred for hundreds of years (IRTCS, 2010).

Although El-Nino and La-Nina events are characterized by warmer or cooler than average sea surface temperatures in the tropical pacific, they are also associated with changes in wind, pressure and rainfall patterns. In the tropics where El-Nino and La-Nina form, rainfall tends to occur over areas having the warmest sea surface temperature (Palmer, 1986). During normal conditions, the warmest water is found in the western pacific, as the region with the highest amount of rainfall. Winds near the ocean surface travel from east to west across the Pacific (these winds are called easterlies) but during El-Nino, the easterlies weakened, warmer than average sea surface temperature covers the central and eastern tropical Pacific, and the region of heaviest rainfall moves eastward as well. However, during La-Nina, the easterlies strengthened, colder than average ocean water extends westward to the central Pacific, and the warmer than average sea surface temperatures in the western Pacific are accompanied by heavier than usual rainfall (Folland et al., 1986; Nicholson and

Kim, 1997). While the tropical ocean affects the atmosphere above it, so too does the atmosphere influences the ocean below it. In fact, the interaction of the atmosphere and ocean is an essential part of El-Nino and La-Nina events (the term coupled system is often used to describe the mutual interaction between the ocean and the atmosphere). During an El-Nino, sea level pressure tends to be lower in the eastern Pacific and higher in the western Pacific while the opposite tends to occur during a La-Nina. This see-saw in atmospheric pressure between the eastern and western tropical Pacific is called the *Southern Oscillation,* often abbreviated as simply the SO. A standard measure of the Southern Oscillation is the difference in sea level pressure between Tahiti and Darwin, Australia. Since El-Nino and the Southern Oscillation are related, the two terms are often combined into a single phrase, the El-Nino/Southern Oscillation, or ENSO for short. Often the term ENSO Warm Phase is used to describe El-Nino and ENSO Cold Phase to describe La-Nina.

Once developed, the El-Nino and La-Nina events typically persist for about a year and so the shifted rainfall patterns associated with them typically persist for several seasons as well. This can have significant impact on people living in areas of tropical pacific since the usual precipitation patterns can be greatly disrupted by either excessively wet or dry conditions. In addition, the shifted tropical rainfall patterns during El-Nino and La-Nina not only affect the tropical Pacific region but also areas away from the tropical Pacific as well. This include many tropical locations as well as some regions outside the tropics in both the Northern and Southern hemispheres (Popelewiski and Halpert (1987).

Hamilton and Allingham (1988) reported an investigation into chronology of El-Niño /Southern Oscillation (ENSO) events between 1531 and 1841 in Peru. They identified aspects of the global teleconnection patterns usually associated with the tropical ENSO events. Van loom and Madden (1981) described the global correlations between a measure of Southern Oscillation and sea level pressure and surface air temperature in the northern winter. They found that the most stable correlation coefficients were over India; the North Pacific Ocean, the Rocky Mountains and the Central (and Western North Atlantic Ocean). The characteristics of the Southern Oscillation (SO) in the tropics and sub-tropics of the southern hemisphere have been presented by many workers, for example, Walker and Bliss (1932), Popelewiski and Halpert (1987), Dai and Wigley (2000) to mention a few.

While the chain of events that constitute an El-Nino or La-Nina is well organized, it provides some statistical predictability in regions around the globe. Once the phase of ENSO is established, actual understanding of the mechanism which starts the chain of events or stops once started remains elusive and the prediction of the onset of El- Niños by both dynamical and physical models are yet to display skill relative to a simple climatology and persistence model of ENSO onset (Landsea and Knaff, 2000). One theory for ENSO is the delayed oscillator theory (Suarez and Schopf, 1988; Battisti and Hirst, 1989) which posits an unstable atmosphere-ocean system where oceanic Rossby waves generated from previous El- Niños or La- Niñas act as the excitation for the next ENSO event. Other mechanisms such as monsoonal activity (Webster and Yang, 1992) and the Madden Julian Oscillation (MJO) have also been proposed as modulators of ENSO; though, such theories only lengthen the chain of causality as these phenomena are themselves even less understood than ENSO. Jin and Neelin (1993) hypothesize that the tropical pacific atmosphere /ocean system is not an unstable system waiting for triggering mechanisms but that the ENSO variability is best thought of as response of the tropical pacific system to stochastic climate noise.

2.2. The ITD model and the weather zones

Nigeria receives rainfall from the south westerlies which invade the country from the Gulf of Guinea coast, i.e. the tropical Atlantic. This moist air stream is overlain by the northeast trades which originate from above the Sahara and therefore are dry and dust laden. The zone of contact of the two air masses at the surface is a zone of moisture discontinuity and it is known as the Inter-Tropical Discontinuity. The ITD advances inland as far as $22\text{-}25^0$ N in August at the margin of the Sahara i.e. considerably beyond Nigeria's northern border (Adejokun 1966; Adedokun, 1978) while it does not retreat equatorward beyond 4^0N latitude during the 'Harmattan' dry season (see Adefolalu, 1983).

Five weather zones are associated with the dynaimics of the ITD in Nigeria. From figure 2.1, it could be observed that Zone A lies to the north of the ITD and hence is rainless as well as zone B to the immediate south because they did not contain rain-producing clouds. Rainfall in the ITD occurs in zones C and D where conditions favor the development of clouds of great vertical extent. Thunderstorms and squall lines are associated with zone C weather and monsoon rains with zone D weather. Consequently, rainfall is spatially discontinuous when zone C weather prevails. On the other hand, the monsoon system gives continuous rains which may last 12 hours or more (see Olaniran, 1986, 1991a, 1995). Overall, rainfall occurs at a distance of about 500km south of the surface location of the ITD, 4-6 weeks behind it in its annual cycle. When the fifth weather type associated with the ITD i.e. zone E, prevails over an area, light rainfall usually occur because Zone E weather is dominated by layered stratiform clouds. A summary of the major characteristics of five weather types associated with the ITD in Nigeria is given by Ayoade (1995) and presented in table 2.1.

This space is for figure 2.1

Table 2.1. Major characteristics of the five weather types over Nigeria

Weather Type	Location	Major Characteristics
A- Harmattan Weather	North of surface position of ITD	Dry, North-easterly winds, Maximum temperature 30-35^{0} C, Minimum temperature 12-18^{0}C, Relative humidity is less than 40%, Visibility is poor because of dust haze.
B- Dry but Humid weather	Up to 300km South of the surface position of ITD	Little rainfall from isolated thunderstorms about 25-50mm per month. Winds change from Southwesterlies in the mornings to Northeasterlies in the afternoons. Maximum temperature 30-35^{0}C, Minimum temperature 18-24^{0}C. Relative humidity is 60-90%.
C- Disturbance Lines Weather	300-1300km South of the surface position of ITD	Moderate to high rainfall from organised lines of thunderstorms known as Disturbance lines/Line squalls about 120-200mm per month. Maximum temperature 27-32^{0}C, Minimum temperature 18-24^{0}C, Relative humidity is 65-95%.
D- Monsoon Rain Weather	1300-1600km South of the surface position of ITD	Heavy monsoon rainfall from monsoon depression about 300-350mm per month. Small diurnal range of temperature. High humidity of more than 80%
E- Little Dry Season Weather	More than 1600km South of the surface position of the ITD	Low to moderate rainfall from stratus and stratocumulus clouds about 125-175mm per month. Maximum temperature 21-27^{0} C, Minimum temperature 18-21^{0}C. Relative humidity is 65-95%.

Source: Ayoade (1995)

The position of the ITD fluctuates seasonally over West Africa and the different ITD zones affect different areas of Nigeria at various times (see Fig. 2.2). Between January/February and August, the ITD migrates northward and there is a corresponding shift northward of the area of rainfall activity, and from the end of August when the ITD is at its most northerly position, zone E weather migrates a short distance inland and caused a period of reduced rainfall in the coastal area, a phenomenon known as the 'little dry season' or the 'July/August break'. During this period the south-westerlies becomes deflected in to westerlies which brings little or no rain. This causes rainfall to increase eastward over southern Nigeria during the July-August period (Olaniran, 1988*a*, *b*, 1991). The account of the rainfall-producing systems presented above for Nigeria, depicts rainfall activity over the country as a function of the migration pattern of the ITD (see Kowal and Knabe, 1972; Ayoade, 1974; Olaniran. 1985; 1988*a*, *b*; Adefolalu, 1986). Accordingly, droughts in Nigeria, and indeed over West Africa, are associated with a restricted northward advance of the ITD.

Different from this simplistic picture, the ITD itself is erratic in its south-north advance and north-south retreat. It moves in a series of surges, retreats and stagnations. Data presented by Walker (1958) showed that along longitude 3^0E in that year the ITD advanced up to 11^0N latitude in January but retreated southward to 6^0N latitude in February i.e. the following month, a retreat of 500km. Oguntoyinbo and Richards (1977) also reported a similar situation for southern Nigeria during 1972/73. Such irregular movements of the ITD have implications for the location of the area of rainfall activity over the country. Often, they cause a false start of the rainy season i.e. early onset of rainfall at a location which is subsequently followed by a prolong dry spell.

This space is for figure 2.2

In Nigeria, as in most of West Africa, the Inter-Tropical Discontinuity is located by the surface difference in dew point or vapor pressure between the northern and southern air streams. The dew point in the northern air is very much below 15^0 C while that in the southern air is usually over 15^0 C (Odekunle, 2004). The ITD is, therefore, frequently located at the surface by a threshold dew point of 15^0 C or 15mb vapor pressure. When the surface dew point at a station falls below 15^0 C, the ITD is said to have passed the station from a northerly direction. When the dew point, on the other hand, rises above 15^0 C, the discontinuity is said to have passed from a southerly direction (Kraus, 1977a, b; Odekunle, 2004). The diurnal variation in the dew points of both northern and southern air streams, established by Adejokun (1966), which points to a morning maximum at 0800hr (all times GMT) for the northern air and a double maxima at 1000hr and 1800hr for the southern air exercises no disruptive effect on the daily location of this discontinuity. The location of this discontinuity by the use of the 15^0 C surface dew point isopleth is, however valid only insofar as the northern and southern air stream dew points coincide with the zone of the maximum dew point gradient (Adejokun, 1966).

2.3. The General Circulation of the Atmosphere (GCA) Model

According to Allan *et al* (1996), the first usage of the word ‘teleconnection’ in a climate context is in a paper by Bjerknes (1966), although the focus of that study is on climatic changes in the north Atlantic sector and Europe. Namias (1969) also used the terminology early on, in discussing the connection between weather phenomena in different parts of the world and he recognized the importance of Bjerken’s conceptual model of large-scale air- sea interactions in a paper (Namias, 1969) in which he proposed that feedback mechanisms between ocean and atmosphere in the North pacific leads to the development of decadal –length climatic regimes.

It is generally acknowledged, however, that it is Bjerknes who organized and interwined elements of the Southern Oscillation, global-scale teleconnections and the large-scale air-sea interactions associated with pacific warming into a new conceptual and thermodynamic reasoning (Rasmusson, 1984). Certainly, the principal teleconnection associated with anomalous rainfall in the equatorial pacific, known as the Pacific North America (PNA) pattern was suggested in his 1966 paper, a study of the effects of the 1957-1958 El- Niño episode (Bjerknes, 1966). However, it was in Bjerknes’s (1969) that a request link between the northern Hemisphere (NH) winter circulation and the warm phase of the southern oscillation (SO), through change in the strength of the Hadley and Walker circulations was established.

A pacific basin-wide view of the sea surface temperature (SST) changes associated with the changing phase of the SO was given by Weare et al (1976). The recognition of the strong modulation of the thermal field of the entire pacific ocean associated with the SO stimulated theoretical (e.g. Hoskin and Karoly, 1981; Webster, 1981) observational (e.g. Horel and Wallace, 1981; Van Loon and Madden, 1981) and general circulation modeling experiments (e.g. Rowntree, 1972), which sets the stage for a burst of activity at all levels in the aftermath of the major El- Niño event of 1982 – 1983

According to Oguntoyinbo (1986) the atmosphere does much of its work over large geographical scales so that climatic anomalies tend to be extensive in space. Thus, it is common to find the variation of one element in one area correlated with its variation in another area sometimes quite remote or correlations may exist between different elements over such distance. Linkages such as these are called teleconnections. It is used to describe relationships between anomalous surface boundary conditions at one location on the earth's surface and meteorological variables (e.g. upper-atmospheric flow patterns, temperature, precipitation) at other locations (Namias, 1981; Wallace and Gutzler, 1981).

The atmospheric and oceanic teleconnections govern the variability in our climate system on a broad range of time and spatial scales, in both the tropics and extra-tropics. On inter-annual time scales, the connection between El-Nino/Southern Oscillation and the Asian monsoon system influences rainfall amounts in regions particularly sensitive to floods/droughts (Popelewski and Halpert, 1987). On inter-annual and decadal time scales, rainfall variability in the Sahel region of West Africa appears to be governed to a large extent by teleconnection patterns related to Pacific Ocean, the Indian Ocean and the Atlantic Ocean (Folland et al., 1986). The decadal behavior of the North Atlantic Oscillation (NAO) that influences climate in Europe, Asia and north Africa, is also likely to be connected to both tropical and extra-tropical sea surface temperatures in the Indo-Pacific and Atlantic regions (Popelewski and Halpert, 1987, 1989).

The subject of teleconnections in the atmosphere and oceans has motivated a great deal of vigorous research in the last 30 years or so (see Gill and Rasmusson, 1983; Nicholson, 1986, 1996; Lindesay et al., 1986; Wolter, 1989; Rasmusson and

Mo, 1993). The main research areas have been coupled ocean-atmosphere processes that lead to better predictions of the ENSO phenomenon (Hoskins and Perace, 1983). The circumglobal waveguide and its use as a modeling agent for large-scale hemispheric wave patterns near the core of subtropical jet are regarded as another example of an important outcome of teleconnection research. As a consequence, enormous progress has been made in seasonal forecasting, which allows the estimation of the climate anomalies for several months ahead and might have important implications for agriculture and economies (Tourre and White, 1995). However, progress in seasonal forecasting is far from saturated and further improvements may be expected from better predictions of sea surface temperature outside of the ENSO region.

Decadal predictions, as an emerging research field, are beginning to bridge the gap between seasonal forecast and climate change predictions. Such predictions aimed at investigating climate anomalies for the next few decades and may therefore be of more immediate relevance than climate change predictions that aimed at a time horizon of several generations a head (Tourre and White, 1995). The separation of human-induced climate change and natural climate variations are some of the most critical challenges in this timeframe. Very complex Earth syatem models are needed for this purpose. However, teleconnection research may provide crucial understanding for progress in this field.

However, the general circulation of the atmosphere (GCA) broadly determines the global weather and climate. The general circulation of the atmosphere can be described as the large scale patterns of wind and pressure which persisted through the year or recurred seasonally (Barry and Chorley, 1976). It is the large-scale motion of the atmosphere averaged in time and space. The general circulation of the atmosphere (GCA) owes its existence primarily to the inequalities in the net radiation between the earth's surface and the atmosphere and between the tropics and the extra-tropical areas. However, the general circulation of the atmosphere redistributes not only the energy in the earth-atmosphere but also moisture and momentum which are also unequally distributed. Pole ward of the latitudes 40^0 North and South of the Equator and in the Equatorial zone between 10^0 North and South of the Equator, precipitation exceedes evaporation. There is therefore a net transfer of water vapor into the equatorial zone via the General Circulation of the Atmosphere (Ayoade, 1983, 2004).

Similarly, there is excess of angular momentum in the low latitudes while there is deficiency in the middle and high latitudes. The atmosphere possesses angular momentum because it rotates with the earth as the earth rotates. The angular momentum per unit mass of a body rotating about a fixed axis is proportional to its velocity and its distance from the axis of rotation. The distance from the axis of rotation in case of the earth and its atmosphere is highest at the equator and decreases pole ward to reach zero at the poles where the axis of rotation itself passes. Accordingly, angular momentum decreases from the equator poleward. With a uniformly rotating earth and the atmosphere, there is conservation of angular momentum i.e. the total angular momentum remains constant. The friction between the tropical easterlies and the earth rotating from west to east generates easterly angular momentum in the low latitudes. This excess momentum is transferred to the sink areas in the middle latitudes where the westerlies continually imparted westerly momentum to the earth by friction. It has been estimated that were this momentum not continually replenished from the low latitudes via the general circulation of the atmosphere, the westerlies would die off within 10 days due to frictional dissipation of energy. If we assumed a homogeneous non-rotating earth, the global wind system would look simpler than it was shown in figure 2.3.

This space is for figure 2.3

With a homogeneous rotating earth, the winds will be subjected not only to the pressure gradient force but also to the coriolis force. Winds moving from the areas of higher pressure to those of low pressure are deflected to the right of their path in the northern hemisphere and to the left of their path in the southern hemisphere owing to the effect of coriolis force arising from the rotation of the earth. For a heterogeneous rotating earth, the pressure distribution patterns are more cellular than zonal owing to the differential heating of land and water surfaces. The global wind systems will then be as shown in figure 2.3. If we imposed on this the fact that the earth surface has varied topography, the pattern of wind systems will even be more complex than shown in figure 2.3. Relief (i.e. topography) could influence the wind systems in two ways. First, they are capable of obstructing air flow by posing as a barrier to air flow (barrier effect). Relief could have a channeling effect on air flow by forcing it to flow through valleys and canyon in hilly terrain (channeling effect). Relief also encourages frictional drag on air-flow (frictional effects) thus reducing wind speed at low levels (Barry and Chorley, 1976; Ayoade, 1983, 2004).

Ayoade (2005) discussed several factors that could determine the pattern of the general circulation of the atmosphere among which include the distribution of land and water surfaces, the rotation of the earth and variation in the earth's topography. A change in any of these factors may lead to a change in the pattern of the general circulation of the atmosphere and hence a change in global weather and climate (Nieuwolt, 1977; Ayoade, 1983, 2005).

West Africa consist of two different regions with different climates. Rainfall increases away from the Sahara both southward and northward. The region north of the Sahara, often referred to as North Africa, is influenced by the Mediterranean climate and mid-latitudes cyclones. West Africa is the region between the Sahara desert (in the north) and the Atlantic coast (in the south). West Africa is located within the tropics, where the atmospheric circulation featured the meridional overturning known as the 'Hadley circulation'. Driven by the differential heating in the atmospheric surface layer, the Hadley circulation occurs with the rising branch near the equator and descending branch over the subtropics. This sinking motion is the main cause for the formation of the Sahara desert over West Africa (Tuckler *et al*, 1991). Another important circulation in West Africa is the ‘monsoon circulation’, which is driven by the differential heating between the land and the ocean (Webster, 1981).

During the summer, wind blow from the ocean to the land at the low levels, with the returning wind towards the ocean at the high levels. The rising branch of the monsoon circulation is located over land. The strong monsoon wind brings moist air from the ocean surface to the continent. The rising motion of the air over land generates summer rainfall in most of West Africa, Nigeria inclusive. During the winter, this circulation is reversed. Except along the coast where winds are towards the land all the time, the low-level wind in most of West Africa blows from the interior towards the ocean in winter, bringing with it the dry and dusty air from the desert. The subsiding branch of the winter circulation is over lands, which suppresses rainfall. This seasonal pattern of the monsoon wind is shown in figure 2.4, using August and February as examples.

This space is for figure 2.4

The seasonality of the monsoon circulation causes strong seasonality of climate over most of West Africa, marked by a wet summer and a dry winter. Moving away from the coast, the strength of the monsoon wind decreases, and so does the length of the rainy season. Figure 2.5 shows the rainfall seasonal cycle at 0^0 Longitude. The length of the rainy season decreases from more than 10 months in the coastal area to 1-2 months on the desert margin.

This space is for figure 2.5

Topography over West Africa is remarkably flat. Most of the region lies between the sea level and 400 meters. Therefore, the topography-induced modification to the large-scale atmospheric circulation is negligible. Since the coast line in the south is also parallel to the equator, the West Africa monsoon circulation is primarily a meridional phenomenon. Together with the meridional overturning associated with the Hadley circulation, this favors a high degree of zonal symmetry in the climate of West Africa.

Figure 2.6 presents the spatial distribution of the annual precipitation, air temperature and specific humidity at the 1000-mb level, as well as the downward solar radiation at the surface. Clearly, the contour lines of these key climate variables are oriented parallel to the coast lines. Except for the coastal region in the west and southwest, the zonal variability of climate over West Africa is negligible. This justified the use of zonally symmetric models (e.g Zhang, 1997) and zonally averaged models (e.g., Xue and Liou, 1990) in simulating the climate of West Africa.

This space is for figure 2.6

2.4. Physical mechanisms of ENSO events

The amplitudes of ENSO events can vary significantly, however, their phases are remarkably similar. As a result, a composite or typical event can be described. For El-Nino, this composite is presented in figure 2.7 (Rasmusson and Carpenter 1982; Ramage, 1986; Popelewski and Halpert, 1987, 1989; Kiladis and Van Loon, 1988; Kiladis and Diaz, 1989). Near the end of the year preceding an event (i.e. October or November), Sea Level Pressure (SLP) in the eastern tropical Pacific decreases and the southeast trades weaken. No longer supported by the trade winds, warm water in the western tropical Pacific (which has a higher sea level than the east) begins to flow eastward. The flow occurs via sub-surface Kelvin wave which flow along the equator and reaches the western coast of South America in two or three months (Wyrtki, 1975; Ramage, 1986; Cane, 1991). Kelvin waves tend to warm surface waters by physically transporting warm water from the west but more importantly, they suppress the thermocline and thus, prevent upwelling of cold water along the west coast of South America.

SSTs in the eastern tropical Pacific begins to warm around December or January when the first Kelvin waves reach the South American coast (Wyrtki, 1975; Ramage, 1986; Cane, 1991). This is known as the onset of the event and often referred to as year (0). High SST anomalies then begin to spread westward along the equator (see Fig. 2.7*a* and *b*. Since the ITCZ remains over the warmest waters, its normal seasonal northward migration is inhibited (Rasmusson and Carpenter,1982, 1990; Ramage, 1986).

As the El-Nino develops, trade winds continue to relax and eventually replaced by surface westerlies. These westerlies trigger more Kelvin waves which further depress the thermocline when they reach the South American coast. The South Pacific Convergence Zone (SPCZ) moves northward and eastward toward the higher SSTs and eventually merge with the ITCZ (Ramusson and Carpenter, 1982; Philander, 1985, 1990; Ramage, 1986). These anomalous conditions continue through year (0) and reach their peak early in year +1 when SSTs and negative SOI values also reach their maximum anomalies (see Fig. 2.7*c*. The peak is referred to as the mature stage of the El-Nino. The appearance of cold surface waters in the eastern tropical Pacific near the middle of the year +1 mark the end of the event (see Fig. 2.7*d*) (Rasmusson and Carpenter, 1982; Popelewski and Halpert, 1987, 1989; Kiladis and Van Loon, 1988; kiladis and Diaz, 1989).

This space is for figure 2.7

La-Nina events are not fully recognized until the mid 1980s and therefore, have not been extensively researched (Van Loon and Shea, 1985; Philander, 1990). However, typical La-Nina events are generally opposite to those of El-Nino. The onset of La-Nina is essentially caused by strengthened south-east trades. As a result, the upwelling of cold water along the west coast of South America increases and the thermocline becomes shallower and the South equatorial current intensified. The negative SST anomalies in the eastern tropical Pacific increases the SST gradient along the equatorial Pacific and enhance the Walker circulation (Trenberth, 1991). The SST anomaly pattern is further amplified because the strengthened South Pacific high cause more upwelling of cold water which is physically transported westward by the strengthened South Equatorial current. As a result, sea levels are unusually high in the west and low in the east. Colder region of water, which normally only extend to approximately 150^0W, often extend to 170^0E causing the ITCZ to be displaced northward and the SPCZ south and westward (Philander, 1985, 1990; Van Loon and shea, 1985; Trenberth, 1991).

Unlike El-Nino, the timing of a typical La-Nina event is difficult to define because each La-Nina differs greatly in terms of onset and duration. However, several studies which analyzed the teleconnections associated with both El-Nino and La-Nina, proposed that typical La-Nina events also tend to develop near the beginning of the calendar year, mature during the winter following their onset and last around 18 months (Philander, 1985; Van Loon and Shea, 1985; Yarnal and Diaz, 1986; Bradley *et al*., 1987; Lau and Sheu, 1988; Kiladis and Diaz, 1989; Popelewski and Halpert, 1989; Deser and Wallace, 1990).

El-Nino and La-Nina are two extremes of the ENSO cycle. There is a tendency for the system to go from one extreme to the other in adjacent years. Therefore, El-Nino and La-Nina frequently follow one another (Van Loon and shea, 1985; Meehl, 1987; Kiladis and Diaz, 1989; Rasmusson *et al*., 1990). Each ENSO event (both El-Nino and La-Nina) can also vary significantly in terms of timing, evolution, amplitude, duration and spatial extent. These differences can be key factors in determining teleconnection patterns associated with individual ENSO events (Diaz and Kiladis, 1992).

The importance of Hadley and Walker circulations to ENSO phenomenon could be understood better if the mean atmospheric and oceanic conditions are described. The northeast and southeast trade winds move toward the equator along the coasts of North and South America and converged at the Intertropical Convergence Zone (ITCZ) (Figures 2.8 and 2.9). Moist air rises from the ITCZ, diverge poleward in the upper troposphere and descend over both sub-tropical high pressure cells. This completes the meridional (north-south) Hadley circulation (Philander, 1990) (see also Fig. 2.9)

Superimposed on the Hadley circulation is the zonal (east-west) Walker Circulation. It involves ascending motion over the western tropical Pacific in the area associated with the South Pacific Convergence Zone (SPCZ). The SPCZ resultes from convergence of easterly trade winds from the South Pacific high and southwesterly winds generate by migratory surface anticyclones in the Australian region (Kiladis *et al*., 1989) (see Fig. 2.8). The rising air from the SPCZ diverges eastward and westward in the upper troposphere. The eastward component descend over the South Pacific sub-tropical high. From this region, surface easterly winds move along the equator and complete the circulation (Philander, 1990) (see Figs. 2.8 and 2.9). The Walker Circulation leads to trans-Pacific SLP gradient with lower pressure over Australia and Indonesia and higher pressure off the west coast of South America (Brigg, 1990).

This space is for figure 2.8 and figure 2.9

The atmospheric circulation patterns directly influence ocean currents and SSTs (see Fig. 2.10). Along the western coast of South America, counter-clockwise circulation associated with the South Pacific high generate the Peru Current. Due to the alignment of the coast, this current pushes surface water offshore and leads to upwelling of cold water (Ramage, 1986). Surface easterly winds associated with the Walker Circulation lead to the strong westward South Equatorial Current (Philander, 1990). The Current transports cold water (that upwells along the South American coast) westward to the central tropical Pacific. Therefore, along the equator, there is an increase in SST from east to west (Fig. 2.10). This current causes an accumulation of water in the western tropical Pacific which deepens the warm surface layer of the ocean and depresses the thermocline (the boundary between the well mixed surface and colder deeper layers) (Ramage, 1986; Brigg, 1990). North of the South Equatorial Current is the eastward flowing Equatorial Counter current. It results in warm surface waters north of the equator that extend along the entire equatorial Pacific, and are associated with the ITCZ (Philander, 1990) (see Fig. 2.10).

The distribution of strong convection and subsidence in the tropical Pacific is largely determined by the SST pattern and the two convergence zones (which typically lay over the regions of highest SST). High SSTs (which normally heat the atmosphere from below) coupled with moisture convergence from the trade winds, lead to strong convection over the ITCZ and SPCZ. Conversely, due to cooling from below, lower SSTs in the eastern tropical Pacific stabilizes the atmosphere. This, along with an absence of moisture convergence, leads to large-scale subsidence in this region (Trenberth, 1981).

This space is for figure 2.10

2.5. El-Nino/Southern Oscillation (ENSO) and rainfall anomalies

The El- Niño/Southern Oscillation (ENSO) phenomenon has been studied largely in the context of the Pacific Ocean and adjacent regions. Although research has long established that it was a global-scale phenomenon (Horel and Wallace, 1981; Arkin, 1982; Barnett, 1983; Yasunari, 1985). World-wide teleconnections in rainfall have been established (e.g. Ropelewski and Halpert, 1987, 1989; Kiladis and Diaz, 1989). These include areas of eastern Africa, where most ENSO events produce abnormally high rainfall, and much of Southern Africa, where the likely consequence of ENSO is drought (Lindesay *et al.,* 1986; Nicholson and Entekhabi, 1986,1987; Van Heerden *et al.,* 1988). These regions are also strongly influenced by the Atlantic and Indian Oceans (Nicholson and Entekhabi, 1987; Walker, 1990), and the question arises as to whether the ENSO influence in this region is actually manifested through its effects in the Atlantic and Indian Oceans

A number of observations suggest that ENSO episodes trigger large-scale changes in the tropical Atlantic and western Indians Oceans. The dominant time-scale of sea- surface temperature (SST) fluctuations in tropical sectors of these oceans is 4 to 6 years, the same as for the Southern Oscillation (Nicholson and Nyenzi, 1990). Warming is apparent in most of these during ENSO years. Large-scale Atlantic anomalies in association with ENSO have been demonstrated by Hastenrath and Heller (1977), Covey and Hastenrath 1978), Wolter (1987) and Aceituno (1988); Semazzi *et al* (1988) also found an ENSO mode in Atlantic variability. Cader (1985) and Yasunari (1987a.b) have demonstrated strong ENSO signals in the Indian Oceans.

Areas of high SST variability in the Atlantic include the Gulf of Guinea and equatorial Atlantic to the south (Merle *et al.,* 1979) and the region of the Benguela current in the south-eastern Atlantic (Walker *et al.,* 1984; Shannon, 1985, 1990). Warming in upwelling regions in the Gulf of Guinea and off north-west Africa appears to be analogous to the Pacific ENSO and roughly correspond to it temporally (Hisard, 1980; Michelchen, 1985a,b). Anomalous warming occurs along the Banguela coast during most ENSO years (Gilloly and Walker, 1984; Nicholson and Entekhabi, 1987).

Inter-annual variability of SSTs appears to be remotely forced in these regions (Bakun, 1978; Micheclen, 1985b; Merle and Amault, 1985; Shannon *et al.,* 1986; Walker, 1990). For example, the strength of the trades off north-east Brazil is correlated with SSTs and thermocline structure in the Gulf of Guinea (Servain *et al.,* 1982: Sevain and Legler, 1986). A relaxation of the trades apparently induce a Kelvin-type wave to propagate along the Equator and elevate SSTs in that region and further south (Moore *et al.,* 1978; O' Brien *et al.,* 1978). However, Zebiak (1993) has recently demonstrated a strong relationship between pseudo-wind stress and SSTs in the equatorial Atlantic and suggest a dynamic ocean-atmosphere coupling akin to ENSO.

Elsewhere in the tropical Atlantic, warm episodes have been demonstrated (e.g. Lamb *et al.*, 1986), but they seem to be somewhat out of phase with the pacific ENSO. One such episode in 1984 follows the 1982-1983 pacific ENSO (Horel *et al.,* 1986; katz *et al*., 1986; Philander, 1986; Hisard and Henin, 1987; Mechoso and Lyons 1988). Maximum warming appears to have achieved towards the end of the onset of ENSO (Hamilton and Allingham, 1988), i.e. during the 'mature phase' of ENSO (Rasmusson and Carpenter, 1982), and the mechanisms of warming are quite complex.

Although ENSO has been shown to be one of the primary determinants of the inter- annual variability in the low-latitudes, its influence over Africa remains controversial. A number of studies have confirmed a relationship between rainfall and ENSO in parts of eastern and southern Africa (e.g Lindesay *et al*., 1986; Farmer, 1988; Janowiak, 1988; Van Heerden et al., 1988;Nicholson, 1996). However, there is disagreement about its influence in the Sahel and elsewhere (e.g. Wolter, 1989; Semazzi *et al*., 1988; Nicholson and Palao 1992).

As a distant forcing, the role of El- Niño/Southern Oscillation (ENSO) in the variability of Sahel rainfall is also investigated (Nicholson and Entekhabi, 1986), but thes correlation between ENSO and rainfall in most of West Africa is relatively weak. Similarly, a study of rainfall patterns in all Sahelian stations in ENSO years does not show any significant correlation between El- Niño events and total Sahelian rainfall (Adedoyin 1989a, b, c). For example, 1958 is very wet and 1983 is very dry in spite of the fact that the El- Niño events of these years are of comparable intensity. There is no pacific El-Nino in 1984 but in summer months (June, July, August) in the Northern Hemisphere of that year, the south Atlantic warms up considerably off the coast of Angola (WMO, 1985). The implication of this event described by some researchers as the 'Atlantic El- Niño' on Sahelian rainfall need to be examined. The effect of the 1984 extra ordinary warming in south Atlantic is easily discernible. This phenomenon is, no doubt, caused by extra-ordinary weakened Hadley circulation. Since this aberration occurs in August, the main rainy month in the Sahel, 1984 has been identified as the driest year in the Sahel, at least since 1901 (Palmer, 1986).

The study of meteorological teleconnections, or relationship between temporal fluctuations of meteorological parameters at widely distant locations, has received much recent attention because of the insight such relationships provide to the understanding of climate and weather variability. Observational studies, commencing with the work of Walker and Bliss (1932), and numerical models provide evidence of such teleconnections in the tropics, in mid-and high-latitude regions, and nearly globally (Wallace and Gutzler, 1981). One example is the Southern Oscillation, a quasi-periodic change in the sea level pressure patterns in the Pacific. Associated with the pressure fluctuations is a distinct teleconnection pattern which is strongest in the tropical and subtropical latitudes but which can be linked to phenomena on a planetary scale (Horel and Wallace, 1981; Rasmusson and Carpenter, 1982). Several other consistent teleconnections have been documented, such as the North Atlantic seesaw (Van Loon and Rogers, 1979) and recent theoretical studies provide a dynamic interpretation of many of the global linkages (Egger, 1977; Webster, 1981).

From theoretical and diagnostic studies of both teleconnections and individual weather anomalies, several points relevant to understanding tropical weather anomalies emerged. One is the spatial coherence of anomalies throughout much of the tropical region. Fleer (1981) and Kraus (1977a, b) demonstrated such coherence on a multiyear time scale and Krueger and Winston (1975) and Kanamitsu and Krishnamurti (1978), among others, documented individual years of anomalous weather throughout the global tropics. Clear linkages between the tropics and the extra-tropics also emerged. Examples of mid-latitude response to tropical forcing are numerous, but many studies also suggest that the mid-latitudes likewise influence tropical systems (Krueger and Winston, 1975; Lim and Chang, 1981; Gray and Love, 1981).

Over Africa, large-scale teleconnections are indicated by both the extreme spatial coherence of rainfall anomalies in the low latitudes and the interrelationship between tropical and extra-tropical regions. Recent surveys of African rainfall (Kraus, 1977a, b; Nicholson, 1980, 1981) indicates a marked coherence of variation over large portions of the continent; this characteristic was especially strong in the semi-arid regions south of the Sahara. As a result, a small number of departure patterns, or anomaly types, suffice to describe much of the rainfall variability in Northern

Hemispheric areas of the continent. These patterns illustrate a strong tendency for synchronous fluctuations along the tropical and temperate margins of the Sahara, resulting in expansion and contraction of the desert zone on inter-annual and longer time scales (Nicholson, 1980b; Nicholson and Chervin, 1983).

Both historical studies and a preliminary analysis of recent rainfall fluctuations suggest that the coherence described for northern Africa is continental and extend throughout much of southern Africa. This implies strong inter-hemispheric teleconnections (Nicholson and Chervin, 1983). Inter-hemispheric linkages are postulated long ago by Namias (1963), base on diagnostic studies of rainfall and temperature over South America, and are evident in the linkages with the Southern Oscillation (e.g. Rasmusson and Carpenter, 1982; Trenberth, 1976).

An example of the difficulties in associating an observed teleconnection pattern with physical mechanisms is the Madden-Julian Oscillation (MJO). First identified in the 1970's, the MJO is characterized by an observed, eastward moving atmospheric circulation anomaly and associated convection anomalies which can be identified in the wind, cloud and outgoing long wave radiation (OLR) fields along the equator with an approximate 40-50 day time period. The convective anomalies are strongest over the Indian Ocean and eastward over the west pacific warm pool to the date line. Little sign of convective anomalies appears from the central through eastern pacific.

The MJO differs from the relatively spatially fixed teleconnection patterns such as the Arctic Oscillation in that it travels across the pacific with a speed of approximately 5-10 m/s. The exact nature of the forcing of the MJO and its methods of eastward propagation have eluded theoretical explanation as yet (Waliser *et al*., 1999) and the oscillation is not well represented in model simulations (Slingo *et al* 1996). This is a problem in that the MJO dominates tropical climate variability at intra-annual timescale (while ENSO dominates inter-annual variability). Theories for the initiation and propagation of the MJO have centered on the wave-CISK mechanism (Lau and Peng, 1987) and the wind- evaporation feedback (Emmanuel, 1987; Neelin *et al*., 1987). Both have failed to produce adequate representations of the MJO, however. Wave CISK theory typically produces oscillations with phase speeds of 15m/s or greater which is significantly faster than observations. Wind-evaporative feedback mechanisms requires easterly winds at the surface. While surface winds are easterly in much of the tropics, regions where the oscillation is most noticeable has climatologically westerly winds at the surface.

The idea that the MJO is part of a coupled atmosphere-ocean oscillation, similar to ENSO, is the subject of active research (e.g. Waliser *et al*. 1999; Woolnough *et al*. 2000). Recent research (Seo and Kim, 2003) concludes that the MJO is a coupled oscillation of the ocean-atmosphere system and represents an interaction between two classes of waves, Rossby and Kelvin waves leading to a self-generating and self-propagating disturbances. The MJO is associated with the timing of the active and break period of both the Indian and Australian monsoons (Madden and Julia, 1994) and might have some role in triggering ENSO events (Kessler *et al*., 1996; Zhang and Gottschalk, 2002) further complicates ENSO prediction. There does not appear to be a strong signal of the MJO in the extra-tropics (Madden and Julian, 1994).

2. 6. Sea surface temperatures (SSTs) and rainfall anomalies

Different from the large-scale southward displacement of the ITD, drought occurrence can also be attributed to factors which inhibit rainfall activity behind (i.e. to the south of) the Inter-tropical Discontinuity (ITD). This may take the form of a weakening of the rainy season "intensity". According to Palmer (1986), warming of the tropical Atlantic Ocean reduces the meridional gradient of sea-surface temperature (SST) south of the ITD and results to a weakened Hadley meridional circulation (i.e. the pattern of circulation of the atmosphere over the tropics). The weakened circulation reduces the intensity of the southwest monsoon flow into west and central Africa and consequently reduce rainfall in the region.

Significant climate anomalies over the globe in the past few years can be illustrated in terms of deviations of sea surface temperature and precipitation from normal. This is particularly so in African countries spanning the Sahel where there has been a downward trend in annual rainfall over the past 20 years. Available facts from these countries, as given by the World Meteorological Organization (1983), are representative of the observed deficits in moisture supply to the Sahelian region of Africa in the last quarter of a country. The immediate causes of these deficits, according to Obasi (1984), appear simple. These are:

(a) Lack of adequate moisture in the affected areas;

(b) Absence of (or weakened) organized atmospheric rain-generating systems during the rainy season;

(c) Persistent widespread subsidence which often results from large-scale global atmospheric circulations.

Hypotheses proposed for the persistence of the decline in rainfall in some parts of Africa are, however, less simple and mainly unvalidated. Some of these are:

(1) soil moisture deficit leading to reduced evaporation;

(2) lack of (or weakened) atmospheric disturbances necessary for lifting and cooling the air to produce rain; and/or

(3) the lack of layers of dust which reduce downward radiative flux into the Sahara while increasing the heating of the atmosphere thereby creating a stabilizing effect which inhibits convection (World Meteorological Organization, 1983)

Some recent studies, however, indicate that rainfall trends in virtually the whole of Africa are influenced by global –scale SST anomalies. Rainfall trends in Africa between 1901 and 1985 have been shown by Hastenrath (1984), Lough (1986) and Folland et al (1986) to be directly influenced by contrasting patterns of SST anomalies on a global scale. Also by extending the global circulation model experiments of Folland *et al* (1986), Palmer (1986) studied the model's response to SST anomalies in individual oceans and found that, over the western Sahel, the Atlantic and Pacific SST anomalies tend to reduce total rainfall amount whereas the Indian Ocean anomalies produce slight enhancement.

It is likely that these global-scale SST anomalies directly influence rainfall in Africa by altering tropical circulations. For instance, Pacific El-Niño event alters the concept of the Walker Circulation by causing a shift, to the east, of the Pacific ascending branch and create subsidence over some parts of Africa (World Meteorological Organization, 1985). On the other hand, warming of the South Atlantic SST reduces the meridional gradient of SST south of the Inter-tropical Discontinuity (ITD) (Palmer, 1986), and as a result, there is a weakening of the Hadley meridional circulation. The weakened circulation reduces the intensity of the south-west monsoon flow into west and central Africa.

A limited analysis of atmospheric conditions that accompanied dry and wet Sahel years by Newell and Kidson (1984) also showed that the reason for drier years might be due to weaker moisture-flux convergence into the West African sector of the

Sahel. They found that the lower tropospheric south-westerly winds are shallower and weaker in dry years. The persistence of drought in the Sahel over the last 20 years can therefore be understood easily in terms of the warmer-than-normal SST in the South Atlantic (Hastentrath, 1984; Lough, 1986) during this period. The observation of Motha *et al* (1980) that the 'little dry season' disappears along the Nigerian coast in Sahelian dry years can also be explained in terms of the weakened Hadley circulation; zone E of Hamilton and Archbold (1945) is entirely over the ocean in those years.

Sea surface temperature anomalies have been found to be closely associated with rainfall in many areas, and increasingly are being found useful as predictors of seasonal rainfall, most notably in tropical regions (e.g Hastertath, 1984; Bah, 1987; Janowiak, 1988; Mechoso and Lyons, 1988; Adedoyin, 1989a, b, c; Ward and Folland, 1991; Chuand He, 1992; Phillips, 1992; Rowell *et al*... 1992; Fontaine and Bigot, 1993) but also in the mid-latitudes (Phillips, 1992; Zorita *et al.*, 1992; Peng Mysak, 1993). In the southern African region association between sea-surface temperatures and rainfall is evident but a more thorough investigation is required before a seasonal forecasting program can be established.

It is thought that sea-surface temperature anomalies in the south-west Indian ocean have an important influence upon the formation of Southern Africa's tropical-temperate rain-bearing synoptic system (Walker, 1989, 1990; Jury and Pathack, 1991). High sea-surface temperature to the north of Madagascar encourages the development of tropical easterly disturbances over the western equatorial Indian ocean rather than over the sub-continental interior (Jury and Pathack, 1991), hence resulting in dry conditions over the land. In addition, high sea-surface temperatures throughout most of the Agulhas Current region and the Mozambique Channel are associated with wet conditions over the summer rainfall region, as a result of anomalously high sensible and latent heat fluxes into both the tropical easterly inflow and the temperate system forming to the south of the country. Important feedback mechanisms between the atmosphere and the ocean helps to maintain the anomalous conditions (Walker, 1989; Jury *et al.*, 1993).

Although the inter-annual sea-surface temperature variability in the South Atlantic ocean is less than in any of the other oceans in the Southern Hemisphere (Streten, 1981), association with southern African rainfall variability does occur (Walker, 1987, 1990). Anomalously high sea-surface temperature in a broad zonal band between the approximate latitudes of 15^0 and 35^0 S and in an area immediately to the south of the sub-continent accompanied wetter conditions over much of the summer rainfall region. High sea-surface temperature in this region would enhance atmospheric baroclinicity and hence cyclogenesis, which would have implication for southern Africa rainfall. In the temperate latitudes of the South Atlantic oceans, cyclonic activity has been found to be enhanced over both warm water anomalies and along area of strengthened sea-surface temperature gradients (Brundrit and Shannons, 1989: Walker and Lindesay, 1989). High sea-surface temperature between the latitudes of 15^0 and 35^0 S of the south Atlantic ocean implies a poleward shift in the zone of surface westerlies also (Walker, 1989), a phenomenon known to be associated with anomalously wet conditions over the interior (Tyson, 1986).

Lough (1986) presented a strong correlation between the Atlantic SST and the Sahel rainfall during the period 1948-1972. The importance of regional SST forcing are also indicated by modeling studies which simulate the contrast of the Sahel rainfall between individual years based on only the regional SST forcing in those years (e.g. Druyan and Hastenrath, 1991; Druyan, 1991). Recently, a modeling study of Zheng *et al* (1999) document that a spring warming and/or a summer cooling in the tropical Atlantic favors a positive anomaly of the summer rainfall over the Sahel, which may have implications regarding the predictability of the Sahel rainfall.

Other studies extend the region of the focus beyond the Atlantic ocean. Folland *et al* (1986) found that the persistent wet and dry periods in the Sahel region are strongly correlated to contrasting patterns of SST anomalies at the global scale. Using an Atmospheric General Circulation Model (AGCM), they showed that worldwide SST anomalies could influence the summer rainfall over the Sahel through changes in the tropical atmospheric circulation. Using the same AGCM, Palmer (1986) showed that the SST patterns in the Pacific and Atlantic oceans are equally important in reducing the rainfall over the western Sahel while the Indian Ocean plays the dominant role in rainfall reduction over the eastern Sahel. The importance of the global SST forcing found by Folland *et al* (1986) and Palmer (1986) was confirmed by studies that forced the general circulation models with observed SST data for individual years and successfully reproduced the observed anomalies of the seasonal Sahel rainfall for those years (Folland *et al*, 1991; Palmer, 1986). Despite the improved capabilities of these models, a proper evaluation of the respective roles of the continental surface conditions and of the tropical oceans (and of their interactions as well) remains to be achieved. This would constitute a major challenge for the coming years, as underlined by Nicholson (2000) and in the recent report produced by the CLIVAR Africa Task Team (2000, hereafter CATT).

2. 7. Seasonal movements of ITD and rainfall anomalies

Analyses of rainfall over West Africa have however largely been confined to the estimation by multivariate statistics of the relative contributions of the geographic factors of altitude and continentality, for example, to rainfall amounts (Gregory, 1965; Oyebande and Oguntoyinbo, 1970; Ayoade, 1974). Another approach which has been adopted by only a few workers is the examination of various atmospheric parameters as they relate to the observed rainfall amounts.

Earlier studies (Bryson, 1973) and Wistanley (1973a) linked the rainfall variability in Africa with the position of the Inter-Tropical Convergence Zone (ITCZ) and suggest that the Sahel droughts might have been caused by a southward shift or displacement of the ITCZ. However, further studies (e.g. Newell and Kidson, 1984; Nicholson, 1981) showed that there is no systematic southward displacement of ITCZ over West Africa.

A comparison of sub-Saharan droughts and wetter years suggest that a northward displacement of the ITCZ might account for wetter years, but that a weakened intensity of the rainy season, independent of ITCZ position is the most likely cause of drought in the Sub-Saharan region. This and the tendency for synchronous fluctuations north and south of Sahara (i.e. tropical and extra-tropical regimes) suggest that changes in the intensity of Hadley circulation may be an important factor in West African rainfall fluctuation (Bunting *et al*, 1976; Ogallo, 1979).

Adedokun (1978) in his study of West African precipitation and the dominant atmospheric mechanisms found out that the ITD influences the northern part of the region while the Walker circulation is more important in the southern part. Lamb (1978) suggests that the influence of the rain-hearing summer monsoon has been less since 1950's. Lamb's analyses are however limited to the Atlantic Ocean and have been refuted by Nicholson (1981) and Newell and Kidson (1984) who showed that over African continent itself, there is no systematic southward displacement of the Inter-Tropical Convergence zone (ITCZ). Namias (1974) and Miles and Folland (1976) also produced additional evidence refuting Lamb's hypothesis. Lamb (1983) however points out that there is systematic difference in the depth of the monsoon layer, its humidity or advective moisture flux over West Africa between wet and dry years.

Olaniran (1983) assessed the influence of the monsoon factor on the seasonality of rainfall distribution in Nigeria and showed that only two elements of the monsoon system determined the magnitude of the rainfall seasonality in Nigeria. These include the seasonal reversal in the wind system which accounts for 73.4% of the variations and frequency of the prevailing wind in January which accounts for 8.3% of the variation. He also noted that monsoonal factor represents the combined influence of different factors that affect rainfall seasonality in the tropics such as distance inland from the coast, variation in latitude and seasonal variation in air mass source.

Limouzin (1969) also studied the contribution of maritime winds to rainfall amount in Dakar while Obasi (1965) and Anyadike (1979) related precipitable water to rainfall amounts. Anyadike using monthly ASECNA radiosonde data for 1968-76, increased the number of contributory atmospheric parameters to five-depth of maritime winds, precipitable water content, depth of convectively unstable layers and Zonal and meriodinal water vapor fluxes for Abidjan, Bamako, Dakar, Niamey and Nouadhibon. Result of this study showed that a large proportion of the variation in rainfall amounts is explained by the five factors only for the two stations of Bamako and Niamey. The central Plateau and eastern region have the factor depth of maritime winds that explained the largest proportion of the rainfall variance while the precipitable water content of the atmosphere is only a secondary factor.

The dominant influence of the ITD over rainfall is related to the varying degree of vertical motion associated with the northern and southern air masses found north and south of the discontinuity. Flohn (1960), studied the structure of the Inter-tropical Convergence Zone (ITCZ/ITD) in the tropical Atlantic between 10^0 N and 10^0 S and estimated the varying degrees of vertical motion on either side of the boundary. In the region of the NE trades near the ITD, Flohn observed that the low-level trades are slightly convergent up to the trade wind inversion close to 1.5km and divergent between 1.5 and 5.0km. He obtained a maximum upward motion of about 0.34cm sec – 1 (300 m day – 1) at the 1.5km level, with decreasing values down to the surface and up to the 3-km level above which subsidence prevails. Flohn then deduced that the shallow and relatively weak lift in the NE trades is responsible for the dryness that occurred on the northern flank of the ITD. South of the ITD, Flohn found much stronger upward motions extending to 3 – 5km. A typical maximum value is about 1.5cm sec^{-1} (1300 m day^{-1}) 500 – 600 km south of the ITCZ/ITD. He ascribed the great convection activity and unstable conditions South of the ITD to the influence of this strong vertical motion.

Flohn's study, together with the very low rainfall to the northern to the progressively higher rainfall south of the ITD, seem to confirm the presence of spatial variations in vertical motion on either side of the moisture boundary. Moreover, since the ITD defines the northern boundary of the moist southern air, its motions determine the space-time variations in the northward penetration of this moist layer, and invariably, the corresponding variations in the depth of the convective layer. In West Africa, and Nigeria in particular, aerological evidence suggests that the depth of the moist layer ranges from 5,000 – 10,000 ft during the rainy season months of April – October, and is less than 5,000 ft in the dry season months (Hamilton and Archhold, 1945). The depth of this moist layer normally decreases as one approaches the surface position of the ITD; observations which support this variation include 1) low cloud bands and little or no rainfall associated with surface boundary limit and 2) progressively higher rainfall south of the discontinuity (Kraus, 1977). There is thus an inherent association of rainfall patterns with space-time variations in the depth of the moist layer and in vertical motion, which, in turn, depends on the motion of the ITD.

2. 8. Tropical Easterly Jet (TEJ)/African Easterly Jet(AEJ) and rainfall anomalies

The Tropical Easterly Jet (TEJ) is part of the Indian Monsoon system and it extends from India over Africa in the Northern Hemisphere Summer months, generally at a height of around 12-15km. The west-east axis of TEJ is located between 4-10^0 N. On the southern side of the axis, conditions are conducive to the ascent of the air and consequently rainfall occurrence whilst the northern side is marked by subsidence. The TEJ, therefore, reinforce aridity over the extreme northern part of the country but cause a belt of above average rainfall in the central part.

The African Easterly Jet (AEJ) is another important dynamic feature of the circulation over West Africa. It provides the instability required to produce the African wave disturbances that characterized the rainy season (Burpee, 1972). Characteristics of the AEJ (particularly horizontal and vertical shear) influence the inter-annual variability of rainfall (Fontaine et al., 1995). Most previous studies (e.g.

Kidson, 1977; Kanamitsu and Krisnarmurti, 1978; Newell and Kidson, 1984; Fontaine and Janicot, 1992; Fontaine et al., 1995) showed a consistent association between the AEJ and rainfall, with the AEJ being anomalously strong (higher core speeds, stronger shear) in dry years and anomalously weak in wet years. It is generally assumed that the stronger jet is a factor in anomalously dry conditions. Most prior studies of the link between the AEJ and rainfall have focused on the jet intensity because the location of the core could not be evaluated with the available data (Newell and kidson, 1984).

An association between a weaker (stronger) AEJ and higher (lower) Sahel rainfall has been consistently found. However, the intensity at a given station reflects a combination of jet intensity and latitudinal position. Nicholson and Grist (2001) have examined the time series of the intensity of the AEJ and the location of its core during the months of June-September for the years 1958-1997 and found that the core of the AEJ is generally displaced equatorward during the dry decades, compared with earlier years, and the jet becomes substantially stronger in June. They further showed that changes in intensity are less apparent in the other months, but the jet tends to be stronger in the dry years, consistent with the change noted in June. An equatorward shift of the AEJ core occurred in 1967, 1 year prior to the onset of the dry episode in the Sahel. Prior to 1967, the jet core is at ca. 15^0 N in July and September, but generally around 17.5^0 N in August and 12.5^0N in June. An equatorward shift of ca. 2.5^0 of latitude in June, July and September, and ca. 3^0 in August, or about one grid point in all cases.

The westward propagating synoptic disturbances, termed easterly or African waves, occur throughout the West African rainy season. The waves appear to originate as a consequence of a joint baroclinic-barotropic instability mechanism described by Charney and Stern (1962). The instability criterion for this mechanism is dependent on both the horizontal and vertical wind shear. Burpee (1972) showed that the mean shears associated with the AEJ satisfied the Charney-Stern criterion. The African waves modify the frequency of cloud clusters and the amount of rainfall associated with them (Thompson et al., 1979; Houze and Betts, 1981) and seem to influence the inter-annual variability of rainfall (see Reed, 1988; Nicholson and Grist, 2001). This implies that the shear associated with the AEJ plays a major role in Sahel climate. Observations and models have shown that both the horizontal and vertical shear influence the characteristics of the African waves (Burpee, 1974; Rennick, 1976; Simmons, 1977). Both the location of the zone of maximum shear over West Africa and its intensity appear to modulate rainfall variability in the region (Norquist *et al.*, 1977; Druyan, 1998).

More than three-quarters of the total annual rainfall in Sahelian Africa is attributable to squall lines (Eldridge, 1957: Omotosho, 1985, 1990). Adedoyin (1989b, 1989c) has shown that the process of the initiation of these tropical disturbances is linked with the instability of waves generated along the surface of discontinuity between the two tropospheric air masses (i.e the moist south-westerlies and the dry north- easterlies) in the sub-region. These squall-inducing vertical transverse waves are likely initiated by the passage of the trough of the dominant horizontal transverse waves in tropical North Africa namely, the Africa Easterly Waves (AEW). It may therefore seem that the more the number of AEW in any year, the more the number of squally activities and the better the rainfall regime in Sahalian Africa.

In pursuance of this hypothesis, Reed (1988) and Landsea and Gray (1992) have suggested that AEW are less numerous in Sahelian dry years but Pasch and Avila (1994) have shown that the number of waves per year is relatively constant (around 60) no matter if the Sahel is wet or dry. Therefore, the reduction in the number of squall lines in the Sahel in dry years, as found by Adedoyin in (1989a), cannot be conclusively ascribed to a reduction in the number of AEW.

However, one synoptic feature which seems to distinguish Sahelian dry years from the wet is the strength of the African Easterly Jet (AEJ). AEJ is stronger in dry years (Newell and Kidson, 1984: Janicot, 1992). There is therefore a need to investigate whether or not there is any relationship between enhanced horizontal shear (caused by stronger AEJ) in the lower troposphere of West Africa and the evolution of wave-like perturbations along the surface of discontinuity between the two tropospheric air masses in the sub-region since, the amplification of these perturbations has been shown to be fundamental to the development of squall lines (Adedoyin, 1989c).

In addition, other attempts have been made to link African rainfall fluctuations with sea surface temperatures, moisture transport, upper level pressure, wind patterns and circulations in the Hadley circulation (Hastenrath, 1984; Lough, 1986; Palmer, 1986 and Adedoyin, 1989a). Pearce (1988) noted that there is little variation from year to year of the sea surface temperatures off West Africa so that the observed inter-annual variability of the northward penetration of the West African rains must be related primarily to the variability of the strength and position of the African Easterly Jet (AEJ).

2. 9. Biogeophysical feedback mechanism and rainfall anomalies

In the 1970s, internal forcing provides the only explanation of droughts. Biogeophysical feedback mechanisms between land surface and precipitation are studied by Otterman (1974) and Charney *et al*. (1975). Their hypotheses argued that desertification actually contributes to drought, and not vice versa. Otterman (1974) is perhaps the first to propose the idea that modification of land cover characteristics in dry land regions may have climatic effects, citing the example of the Sinai-Negev region, where the denudation of bright sandy soil by grazing on the Egyptian side increases albedo and decreases surface temperature compared to the more densely vegetated Negev side.

Following a similar line of reasoning, Charney *et al*. (1975) used a global circulation model to show a positive feedback mechanism between a decrease in plant cover and corresponding decrease in precipitation via increasing albedo, radiative cooling of the air column above and thereby an enhancement of large-scale atmospheric subsidence and desiccation. In the years following the introduction of Charney's hypothesis, there has been substantial effort made to examine the sensitivity of regional rainfall to large-scale changes in land cover through climate modeling experiments, the results of which support Charney's basic hypothesis that sufficient changes in albedo could, at least potentially, produce droughts. However, satellite measurements of actual sub-Saharan albedo showed no evidence for the persistent increase in albedo necessary to produce significant differences in rainfall (Folland *et al*., 1991; Hulme, 2001). Observed changes in albedo due to conversions of land surface characteristics have been localized in extent and often short in duration, in contrast to the widespread and sustained changes assumed in the modeling studies. On another note, a decrease in vegetation density does not necessarily lead to an increase in albedo, but could in certain cases decrease albedo by reducing the number of geometric elements that reflect incoming solar radiation (Balling *et al*., 2001). Despite the absence of supporting empirical evidence for the Charney hypothesis, modeling studies are valuable as simulation experiments for understanding the inter-relationships between land surface and atmospheric processes (Hulme and Kelly, 1993).

More recently, Balling (1991) puts forward a contrasting but not widely supported hypothesis that desertification produces a warming trend at regional scales, which could be mistaken for a signal of greenhouse warming. Subsequent studies do not support this idea because in the Sahel region, where desertification is most prevalent, the warming trend has actually been smallest of all dry land regions, possibly due to the increased atmospheric dust loading resulting again from human-induced land cover changes (Hulme, 1996).

The role of dust in affecting precipitation is equally controversial. Contrary to what some theoretical models predicted (Yin *et al*., 2000), Rosenfeld *et al*. (2001) suggested that mineral dust in the atmosphere actually reduce precipitation efficiency of clouds due to the coalescence-suppressing effects of large concentrations of dust particles. In addition, dust could inhibit the formation of convective clouds because of radiative cooling and increased subsidence. That would mean that higher dust frequency might be the cause rather than the result of the decreased rainfall. Thus, we found that dust emissions from anthropogenic sources could provide a mechanism for initiating a desertification feedback cycle (Rosenfeld *et al*., 2001). On the other hand, the dust loading over the Sahel has clearly followed, rather than preceded, the trends in precipitation (Nicholson *et al*.,1998). The impact of dust on warming is a complex issue, because it modifies both the incoming shortwave solar radiation and outgoing long wave radiation, either a cooling or heating effect could occur, depending on cloud cover and the albedo of the underlying surface (Nicholson, 2001).

The above arguments have clearly shown that there has been considerable interest in the relationship between land surface effects and the Sahel drought. As noted earlier that the Charney's (1975) pioneered work on the effects of albedo on the African climate showed that increases of albedo could cause a reduction in precipitation. This discovery is confirmed by a number of studies with different models (see Chervin, 1979; Sud and Fennessy, 1982; Laval and Picon, 1986). The effects of soil moisture and evaporation have also been investigated in the modeling experiments (Walker and Rontree, 1977; Shukla and Mintz, 1982; Sud and Fennessy, 1984; Yeh *et al*, 1984; Cuttington and rowntree, 1986 and Rowell and Blondin, 1990. Most of these studies have shown that less initial soil moisture would lead to less precipitation. Furthermore, the combined effects of surface albedo and soil moisture have also been studied (Sud and Molod, 1988; Kitoh *et al*, 1988). Kitoh *et al* (1988) found that the combined effect of surface albedo and soil moisture is nearly equal to the sum of the albedo and soil moisture effects. These modeling studies consistently demonstrate that the land surface has a significant impact on the Sahel climate. However, in most of these sensitivity studies, the area associated with land surface changes and the magnitude of the changes in the surface characteristics, such as albedo and soil moisture, are somewhat arbitrary. All the studies showed a positive feedback between initial soil moisture and rainfall with the exception of one study by Sud and Fennessy (1984). Conflicting opinions have also been expressed on the magnitude of the changes in albedo.

Analyzing satellite observations, Norton *et al* (1979) found that the albedo in West Africa increased during the period 1967-1973. During the wet seasons, albedo changed from 0.23 in 1969 to 0.33 in 1973. Courel *et al* (1984), also using satellite observations, found that the albedo is reduced by 0.1 in the Sahel (mainly in the Ferlo and Goudo regions) from 1973 to 1979. The variation range of 0.1 in both studies is far less than the dramatic changes that are made in some of the modeling studies.

Despite the divergent views on the impact of land-atmosphere interactions on rainfall anomalies, Farmer and Wigley (1985) have provided an excellent summary of the studies of Otterman (1974), Charney (1975) and Rassol (1984) on biogeophysical feedback mechanism (BFM). All these researchers agree that reduced rainfall, combined with human and animal activities (such as over grazing) could reduce the vegetation cover and increase reflectivity, or albedo, of the land surface. Higher albedo changes the heat balance of the surface-atmosphere system and leads to increased divergence in the lower atmosphere and reduce uplift over the higher albedo region. These changes, in turn, lead to less rainfall and maintenance of drought condition. Undoubtedly, biogeophysical feedback mechanism as described above has served to reinforce drought conditions in Nigeria particularly over the northern part of Nigeria. This is due to large-scale depletion of the vegetation for fuel wood in this part of the country (Oguntoyinbo, 1982 and Adefolalu, 1990).

2. 10. Sunspots Cycle and rainfall anomalies

Sunspots are darkened portions of the solar surface and locally they may affect the output of radiation. Lockwood (1979) gives an excellent explanation of the pattern of occurrence of sunspots. They exhibit cycles of 11 years and the double sunspots cycle 20-22 years. Cycles of 9 and 14 years have also been reported for sunspots (Lockwood, 1979).

Generally, increases in sunspots cycles have been associated with weather conditions while decreases in the cycles have been associated with other conditions. Evidence of sunspots cycles has not been found in Nigerian rainfall data (Ayoade, 1973). This implies that the occurrence of dry and wet conditions in the country could not be explained by reference to the sunspots-drought theory.

2. 11. Concept of Drought

One of the major problems of an objective drought investigation is its definition. What is drought? Drought is a complex phenomenon of widespread significance, but despite all the problems it has caused, and still cause man, it has proved formidable to define. The simplest starting point is to regard drought as a relative meteorological phenomenon. In this respect, the term drought is usually associated with a sustained period of significantly below normal precipitation. A precise definition of a sustained period is, however, not clearly found in scientific literature. For instance, in a humid environment where precipitation is normally evenly distributed throughout the growing season and where irrigation is not widely practiced, a summer dry period of several weeks may constitute a drought. On the other hand, in persistently drier areas, such as the Sahelian zone of West Africa, droughts are recognized only after two or more rainy seasons without rain. Moreover, drought in relation to water use may be regarded as a shortage of food or water rather than precipitation. In this case drought definition is not purely meteorological, because the shortage of food or water will depend on how well water supplies are managed. Thus, there is no universally acceptable definition of drought. This is because (i) unlike flood, drought is not a distinct event, and (ii) it is often the result of many complex factors that interact with the environment. Complicating the problem of drought definition is the fact that it has neither a defined period of start nor end. It is recognizable only after a period of time, and because drought may be interrupted by one or more short wet periods, its end is often difficult to recognize.

There is only a distinction between the concept of drought and that of aridity. Aridity is a permanent climate feature of a region that results from low precipitation. Drought, on the other hand, is a temporary feature of the climate of a region, and occurs only when less than adequate precipitation occurs for a period of time. Because of the undramatic behavior of drought, the scientific literature is still full of definitions that reflect the area of interest of individual investigators or the purpose of study.

The climate of an area, the amounts of soil moisture, spatial and temporal distributions of precipitation series, water-table fluctuations, water quality and soil types are among the natural factors most widely used to designate drought. Human factors may include 'the degree of water storage, distribution, system, use per capita consumption, number, locations and depth of wells, and many more' (Mattai, 1979). Good expositions of the problems of drought definition are given by Subrahamanyam (1967), Hounam *et al.*(1975), and Matthi (1979). A recent review is given by Dracup *et al*. (1980) and Oladipo (1985), although they failed to offer any other concise definition. There are four commonly used definitions base on meteorological, hydrological, agricultural and economic considerations.

Agricultural and hydrological drought corresponds to the two main uses to which water is put. Agricultural drought occurs only when available soil moisture is inadequate to meet evaporative demand by crops. Hydrological drought, on the other hand, refers to periods of below normal stream-flow and depleted reservoir storage. Palmer (1965) attempts to combine these two types of drought into an overall measure of drought severity. An alternative overall measure is meteorological drought which depends on rainfall or rainfall-potential evapotranspiration (e.g. Tabony, 1977). The economist saw drought from an entirely different point of view-that of areas of human activity affected. Regarding drought as a supply and demand phenomenon, an index of economic drought may be defined as 'rainfall-induced shortage of some economic goods… brought about by an inadequate or badly timed rainfall' (Sandford, 1978). By this definition, it implies that the incidence of drought depends not only on rainfall but also on trends in requirements and on factors other than weather, such as water supply management.

The above summary of definitions demonstrates the apparent difficulties encountered in an attempt to give an objective definition of drought. A drought may exist in the agricultural sense before it is evident to the hydrologist. Conversely, an agricultural drought may have ended, at least temporarily, by rainfall that replenishes soil moisture, but which is not heavy enough to contribute to ground-water or stream-flow. An agricultural drought may also exist because of poor temporal distribution of precipitation for the year (season) even though, the year (season) would be statistically speaking normal or above normal without any meteorological drought in evidence.

In the quest for a precise definition of drought, many indices have been developed in order to come quantitatively to grips with drought's climatological aspects. Most of these indices are useful in providing a means of summarizing information about abnormally low water conditions and their potential effects. Ideally, a desirable property of any drought index is that it must be general enough so that drought occurrence in different climatic regions could be modeled. Thus drought indices vary in their degree of complexity. They range from a simple measure of meteorological drought such as the fractional deviation of rainfall from its average (e.g. Tabony, 1977) to the more elaborated crop-water parametric models (e.g.Bunting *et al* ., 1981) which incorporates soil moisture conditions and land use management for climatological areally homogeneous units. Because of the magnitude of effort and cost that would be required, a detailed crop-moisture-need approach is not very feasible for large-scale drought study, especially when data for many climatological stations are considered.

2. 12. Droughts in historical records

Interpretation and quantification of historical droughts which occur in the Sudano-Sahelian belt of Africa is useful in understanding the long-term climatic experience of the region. Here, the term "historical" referred to the period that predated the beginning of scientific measurements of climatic variables (Koslowski and Glaser, 1995). In the central and western Sahel, instrumental records for most climatic processes date back to only the beginning of the century. For large areas, synoptic weather stations are not established until the 1930's. Such series are not long enough, in the climatic sense, to provide information on long-term trends.

The usefulness of historical information and other proxy data in extending the length of available records has been demonstrated in many disciplines. In hydrology, paleo-flood records and other evidence of high water marks (physical, verbal, written archaeological and geomorphological} have been used to augment measured flow records (Guo, 1990; Sutcliffe, 1987). Despite inherent uncertainties, it is generally acknowledged that the information contained in such records, when used in conjunction with the systematic data, could increase the accuracy in estimating flood quantities at a given site (Cohn and Stedinger, 1987); Hosking and Wallis, 1986) Salas *et al*., 1994; Stedinger and Baker, 1987). In climatology, Koslowski and Glaser (1995) used various sources of documentary historical data and reconstructed the ice winter severity in the western Baltic from about 1700. Rodrigo *et al*., (1994, 1995) reconstructed the total annual rainfall in Andalusia during the 16^{th} and 17^{th} centuries using annals and chronicles of cities, ecclesiastic archives and other documentary evidence. Similarly, Oliver (1991) compared monthly precipitation distribution and wind direction during the period 100 BC-100 AD and presented day conditions for many cities base on translations from AL-Biruni's *The Chronology of Ancient Nations"*

The Sahel has a long history of droughts and information on many of these droughts exists in various forms from the days of the great empires and kingdoms. However, there has been no attempt to utilize historical information in articulating various attributes of drought such as their intensity or duration. This is somewhat surprising since drought analysis has often been constrained by shortness of record sample and could clearly benefit from extension of the effective record length (Woo and Tarhule, 1994). The situation is particularly acute in arid and semi-arid environments where for example, several years of record-may yield only one sample drought event.

In northern Nigeria for example where rain-fed agriculture is practiced, periods of severe famines often inflicted by droughts are given specific names by the people and preserved in songs, oral folklore and myth (Watts, 1983). Van Apeldoorn (1981) compiled such folklore events between 1835 and 1956. Eight drought periods are identified, six by their names. These events are subsequently corroborated by other historic evidence including reports of colonial administrators such as the Colonial Blue Books and Annual Report on Northern Nigeria. Unfortunately, apart from the names which hint at the approximate time when they occurred, not much else is known about such characteristics as their spatial extent, their severity or intensity, their durations or even what qualifies them for such honours. The parallel existence of record on historical or folklore droughts and measured scientific data since 1905 allows historical drought events to be defined from the rainfall series.

The perception of drought is inextricably inter-woven in the social and political fabric of societies. Hence, considerations other than climate frequently influence whether an event is recorded or not. For example, a drought which coincides with an important socio-political event such as the death or coronation of a new king is more likely to be remembered and therefore recorded and preserved in some way. In addition, drought concepts vary between cultures and evolve in response to changing environmental and social conditions. This implies that similar evidence or records of historical droughts in different cultures could potentially refer to quite different phenomena.

For an economy based on rain-fed agriculture, the reconstruction of historical drought is facilitated by its close relationship with famine. However, there is no inevitable predetermined relationship between the two phenomena (Watts, 1983). Several factors other than rainfall deficit may cause famine including wars, incidence of pests and diseases, failure of the economy and other administrative measures. Excessive rainfall might also lead to famine by destroying field crops through flooding. Torry (1986, p. 8) distinguished between underlying (ultimate) and catalytic (proximate) causes of famine. The latter are "situational and originated shortly prior to or during an emergency"; the former are "predisposing conditions transforming proximate causes into famine distresses …in fact proximate causes (e.g. drought) could land a household in the clutches of famine with or without the involvement of ultimate causes". Hence, "while a specific drought may be considered a proximate cause of famine, droughts (in general) can be considered an underlying cause" (Glantz, 1987, p. 56). Similarly, Watts (1983, p. 104) observed that in Hausaland (northern Nigeria), "the great hungers of the past are the almost inevitable outcome of excessively poor rainfall *(fari),* either because seasonal totals are greatly inadequate, or the rains terminate abruptly prior to maturity *(Kumshi)* of the upland grain crops". While such considerations allow famine chronologies to be used as proxy for droughts, they nevertheless required that both the input information and the ensuing results be interpreted with caution

2. 13. Previous studies on Droughts

Drought has no universal definition but it is generally characterized by a prolonged and abnormal moisture deficiency (Palmer, 1965). Wilhite and Glantz (1985) thoroughly reviewed dozens of drought definitions and they identified six major categories of drought definitions: meteorological, hydrologic, climatological, atmospheric, agricultural and water management. Dracup *et al* (1980) noticed several drought characteristics in all drought studies: Onset, retreat, intensity, duration, frequency and magnitude. Drought is a “creeping Phenomenon” so an accurate determination of the onset or retreat is difficult (Whilhite & Glantz, 1985). Drought onset could be identified based on precipitation statistics or by drought indices. Identifying a drought’s withdrawal could be measured by a departure of a climate index from its normal value, low flows in rivers or ground water levels or the effects on crops, flora & livestock (Girux, 2001). Drought duration is closely linked to its onset and withdrawal date and is sometimes expressed in terms of the number of consecutive days of no rain. The frequency of an extreme event is usually expressed by its return period or occurrence interval, which may be defined as the average time lag between two events of the considered magnitude or larger magnitude (Dalezois et al., 2000). The magnitude of a considered drought event corresponds to the cumulative water deficit over drought period (Thompson, 1999), and the average of this cumulative water deficit over the drought period is mean intensity.

Though, droughts and anomalous periods of dryness are regular features of the climate of the arid and semi-arid regions of West Africa, there has been increasing recognition of the impact of climatic variations on mankind’s economic and social conditions (Ayoade, 1995). In fact, during the past five centuries, period of extreme rainfall fluctuation ranging from a few years to several decades occurs over the area (Nicholson, 1978). More recently in 20th century, severe droughts which affected all or larger parts of Sahelian African occurred in 1913, 1942, 1973 and 1983. However, it is the droughts of 1968-73 with its maxima in 1972-73 that attracts the attention of the world and scientific community on the existence of the problem. The catastrophic drought, which struck the Sahel in the years 1968 to 1973, prompted numerous hypotheses explaining rainfall variation in the semi-arid sub-Saharan lands (Nicholson, 1981). The first of these (Bryson, 1973 a, b; Winstanley, 1973 a,b), which becomes well known in both scientific and popular literature, interprets changes mainly as a function of the position of the Intertropical Discontinuity (ITD) and subtropical high. Although several subsequent papers (Kraus, 1977 a, b; Greenhut, 1977; Beer *et al*, 1975; Lamb, 1978) support this idea, the hypothesis still remains controversial (Tanaka *et al*, 1975; Nicholson, 1979, 1980a; Miles and Folland, 1974; Namias, 1974). More recently researchers (Kidson, 1977; Schupelius, 1976;

Kanamitsu and krishnamurti, 1978) have found more plausible explanations for that drought by extending their analysis to additional factors determining rainfall intensity such as the contrasting sea surface temperature anomalies.

Since then, droughts as well as associated phenomenon of rainfall anomalies have been studied from a number of different perspectives. On the one hand are the physically based studies, which sought to explain the dynamic causal mechanism in terms of large-scale atmospheric circulation processes (e.g. Wolter, 1989; Druyan, 1991; El-tahir and Gong, 1996; Janicot, 1992; Lamb and Pebbler, 1992; Semazzi *et al*; 1996; Ward, 1992). The major theories and models in this category are reviewed in Druyan (1989), Hastenrath (1995) and Nicholson (1989). On the other hand are the empirical studies which emphasize the characteristics and manifestations of droughts and rainfall variability (e.g. Ayoade, 1974; Damaree and Nicholson, 1990; Hulme, 1992; Nicholson and Palao, 1993; Oladipo, 1995; Adelekan, 1998; Odekunle, 2004). Most of the investigations in Nigeria fall into this category. From the foregoing, it could be deduced that rainfall variability and droughts are part and parcel of the climate of arid and semi-arid environments of West Africa.

Drought has always been a problem associated with extreme rainfall anomalies in Nigeria and always pose a serious threat to mankind economically and other wise. For example, during the drought of 1973, approximately 250, 000 people died in the six worst affected Sahelian countries and agricultural productivity drops to 70% of the pre-drought levels (Oyebande, 1990). In northern Nigeria for example, 10.3million northern people faced starvations of grain, and estimated 600, 000 heads of cattle perished. In 1987, 5 million metric tones of grain, mainly millet and sorghum, valued at over US $ 400 million were lost to drought (Oladipo, 1995). In addition to devastating economic losses, recurrent droughts impose severe constraints on the biological productivity as well as the generative capacity of the ecological systems, mass migrations leading to social dislocations, refugee crises and a perpetration or reinforcement of rural poverty (e.g. Anyadike, 1989; Van Apeldorn, 1981; Watts, 1983, 1989). In fact, many researchers have argued that droughts and water scarcity constitute the major constraint to the attainment of self-sufficiency in food production and the development of the semi-arid regions of Sub-Saharan Africa (Falkenmark, 1987; Glantz, 1987, 1994). This view is not generally shared. The opposing argument is that drought prevails in other parts of the world and, in affluent societies, need not be more than a nuisance (Hulme, 1992; Morse, 1987). However due to over whelming poverty and the fact that subsistence agriculture is almost entirely rain fed, the effects of reduced rainfall are often the loss of agricultural productivity, decimation of livestock herds and other effects which are woven in the economic and social fabric of the region (Morgan and Solarz 1994; Sivakumar, 1992).

Among the researchers who attempt a reconstruction of past climates over the lake are Maley (1973, 1981) using the results of pollen analysis, Kuzbach (1980) base on water-and energy-balance equations and Tetzluff and Adams (1983) base on the comparison of evaporation pattern rates over the lake and early-Holocene Lake Chad. Nicholson and Flohu (1980) have also presented circulation patterns over Africa for three distinct periods between 20,000 and 4,500 BP. However, these results must be treated with caution because as noted by Ojo (1985) lake level fluctuations reflect the control of both climatic and non-climatic factors.

From the data summarized on lake level fluctuations by Street and Grove (1976), it could be inferred that the extreme northern part of Nigeria experiences arid condition 21,000-12,500 BP. Using biogeophysical evidence they show further that equatorial lowland rainforest is scanty in West Africa (including the southern part of Nigeria) during this period. Farmer and Wigley (1985) who summarize the data reported by several researchers notes that between 12,500 and 5,000 BP the climate becomes progressively wetter and a belt of expanded lakes in Africa developed. In particular, in the early Holocene (9,000 –6,000 BP), Lake Chad is reported to have extended over vast area known as Mega-Chad covering some 320,000km^2 (Schneider, 1969).

Ojo (1985) quoting data report in the literature noted that the shoreline of Mega-Chad could be traced to Bama Ridge, 128.7km in the southwest and to Koro Toro, 644m in the east. The water level of Mega-Chad has been put at between 50-53m above the lake level in the late 20th century (see Farmer and Wigley, 1985; Ojo, 1985). Mega-Chad prevails until about 700 AD while the levels of the lake show evidence of desiccation in the period between 700 and 1500 AD. In summary, the condition that prevails in Nigeria 20,000 BP-1500 AD is characterized by alternated dry and wet conditions which differs in severity and geographical extent

However, a survey of the existing literature on the pattern of rainfall variability in West Africa and Africa in general as presented in this work shows that, there are certain grey areas that are yet to be addressed especially when reference is made to the Nigerian rainfall climatology. For example, despite the large number of studies on Nigerian rainfall, several aspects of drought and rainfall characteristics remain unanalyzed or poorly understood. For example, there has been little or no attempt to examine the probability of occurrence of droughts of various intensities/magnitudes. Attempt is made in this work to determine the probability of receiving less than the least, as well as, of receiving more than the highest annual rainfall ever recorded in Nigeria over the period of the study.

Another problem is the uncertainty associated with the period on which to base the analysis of drought and rainfall characteristics. The base period controls the statistical parameters through which droughts are distinguished from other events as well as the magnitude of trends and rainfall fluctuations. Over short intervals, rainfall sequence may constitute significant trends (e.g. the 1950-61 annual rainfall in Sahel) but the same sequences might represent mere fluctuations when viewed from a longer time perspective such as a century. It is in consideration of these inadequacies associated with previous studies that this work examined the long-term trends and variability in annual rainfall and rain days in Nigeria for a century.

In a similar vein, most of the studies on droughts and rainfall characteristics in Nigeria have laid emphasis on mere the intensity of droughts in Nigeria (Aondover,1997) with little or no attempt to determine the spatial-temporal patterns of drought occurrence in Nigeria. In view of this, the present study had examined among other things the spatial-temporal aspect of drought in Nigeria within the study period.

CHAPTER THREE

3. 0. METHODOLOGY

3.1. Data base

The data required for this study were obtained from three different sources. The first set of data relating directly to Nigeria was obtained from twenty-seven selected synoptic stations managed and supervised by the Nigerian Meteorological Agency, Oshodi, Lagos. This set of data include the monthly rainfall and rain days of the available record over the period 1901-2000. Although the record length varies in nearly all cases, the records are long enough to produce stable and comparable statistics (see Table 3.3).The choice of this length of time (100 years) is in support of Guttmann (1999) and Dracup and Keyantash (2002) who encourage the use of longer time series, namely for the identification of multiple year drought events as previously referred to. Wu *et al* (2005) also investigate the effect of the length of precipitation record on standardized precipitation index (SPI) calculation and concluded that the longer length of record used in SPI calculation, the more reliable the SPI values would be.

The monthly dew point temperature data for a 40-year period (1961-2000) were also obtained from the same synoptic stations being supervised and managed by the Nigerian Meteorological Agency, Oshodi, Lagos.The daily dew point temperature data from five stations in Niger republic (Lat. 14^0N – 20^0N) were also obtained and utilized for this study. The daily dew point temperature data from these stations were aggregated into the monthly dew point temperature for the analysis. The monthly dew point temperature data from the twenty-seven selected Nigerian stations and those in Niger republic were used to determine the mean monthly surface positions of the Inter-tropical Discontinuity (ITD). The inclusion of the monthly dew-point temperature data of the five stations located in the Republic of Niger (Lat. 14^0N – Lat. 18^0N) is based on the fact that the ITD reaches its northernmost position in August between latitude 18^0N - 22^0N which is far beyond Nigeria. Therefore, the dew-point temperature data of those stations must be used in order to determine the northern most position of ITD in August. The use of dew point temperature to define the surface position of the ITD is justified on two grounds.

First, of all the measures of the contrast between the tropical continental and the tropical maritime air masses, the humidity content of the air in terms of its dew points is the most representative. Dew points not only show the greatest easily measurable contrast between the northern and the southern air, but they are also readily available. Second, dew points have been found to be closely associated with the occurrence and non-occurrence of rainfall (see Kraus, 1977). Sellick (1960) reported a study in Salisbury, Rhodesia, which showed a 75% rainfall occurrence frequency with dew points $>15^0$C and only 31% with dew points $< 15^0$ C. The relationship is based on dew points at 0800hr-0900hr and rainfall in the succeeding 24hr. The threshold for the dew-point temperature adopted for this study is 15^0 C. This is because the ITD is frequently located at the surface on the dew-point temperature gradient of 15^0 C.

Though, the length of the rainfall records varies with each station having at least 50-year record, data availability on the dew point temperature appears to be the main factor that constrained the length of the data used in this study. The available dew point temperature data during the period of data collection for this study are from 1961-2000. Thus, for the purpose of uniformity in the length of data, this study made use of 40 years data (1961-2000) in evaluating the relationship between the mean monthly surface positions of the ITD and mean monthly rainfall distribution patterns in Nigeria.

The second set of data is the monthly Southern Oscillation Index (SOI) relevant to El-Nino/Southern Oscillation (ENSO) events for a 50-year period (1951-2000) that was obtained from the official web sites of Climatic Analysis Centre of the United States Department of Commerce, Washington D.C in collaboration with National Oceanic and Atmospheric Administration (NOAA), National Weather Service (NWS) and National Centre for Environmental Prediction (NCEP), U. S. A.

Since the Southern Oscillation (SO) affects the entire ocean-atmosphere system in the tropical Pacific, a measure of this variable is often sufficient to quantify ENSO events (i.e. El-Nino and La-Nina) (Aceituno, 1992; Halpert and Popelewski, 1992). Several indices have been used to quantify ENSO event; however, the most common is the southern oscillation index (SOI). It is defined as the normalized difference in monthly mean Sea Level Pressure (SLP) between Tahiti, French Polynesia (18^o S, 150^0 W) (i.e. the Eastern Tropical Pacific) and Darwin, Australia (12^0 S, 131^0E) (i.e. Western Tropical Pacific) (Chem, 1982). The SOI is calculated as Tahiti sea level pressure (SLP) minus Darwin sea level pressure (SLP). Negative values (i.e. anomalously low pressure in the eastern and anomalously high pressure in the western tropical Pacific) represents El-Nino conditions, while positive values (i.e.

anomalously high pressure in the eastern and anomalously low pressure in the western tropical Pacific) represents La-Nina conditions. Numerous investigations have used the SOI to quantify ENSO (e.g. Fu *et al*., 1986; Kiladis and Diaz, 1989; Halpert and Popelewski, 1992) and therefore this study also incorporates this index. The SOI index is determined using the following formula:

$$\text{SOI} = \frac{10(Pdiff - Pdiffav)}{SD(Pdiff)} \quad \text{----------------------} \; 3.1$$

where:

SOI = Southern Oscillation Index

Pdiff = (average Tahiti mean sea level pressure for the month)-(average Darwin mean sea level pressure for the month)

Pdiffav = Long-term average of *Pdiff* for the month in question

SD (*Pdiff.*) = standard deviation of *Pdiff* for the month in question

Multiplying the answer obtained by 10 will give a range of values for SOI of between -35 and +35 (see ayoade, 2004). Sustained negative values of SOI over months often indicate El-Nino episodes as such negative values are usually accompanied by sustained warming of the central and eastern Tropical Pacific ocean and a decrease in the strength of Pacific trade winds (Walker Circulation). Positive values of SOI, on the other hand, are associated with stronger Pacific trade winds and cooling of the waters of the eastern and central tropical Pacific ocean. This episode is known as La-Nina episode which is the opposite of El-Nino. The changes in atmospheric pressure epitomised by the Southern Oscillation Index are thus related to the strength of the equatorial zonal east-west circulation. This is a zonal circulation with ascending air over Indonesia and the descending air in the east Pacific giving rise to surface easterlies and upper westerlies over most of the equatorial Pacific (See Popelewski and Halpert, 1987; Nicholson and Jeeyoung, 1997; Ayoade, 2004)

The third set of data is monthly Sea Surface Temperature (SST) anomalies for a 50-year period (1951-2000). It was obtained from the official web sites of Hadley Centre for Climate/Meteorological Office, United Kingdom and from the Comprehensive Ocean-Atmosphere Data Sets (COADS) global marine data in collaboration with National Oceanographic and Atmospheric Administration (NOAA) and the Climate Prediction Centre (CPC), U. S. A. While the SOI dataset is that of Western and Eastern Pacific Oceans, the SSTs dataset is that of South Atlantic SST anomalies, North Atlantic SST anomalies, NINO 1+2 SST anomalies, NINO 3 SST anomalies, NINO 3.4 SST anomalies, NINO 4 SST anomalies and the Global SST anomalies. It should be noted that both the SOI data relevant to ENSO and the SSTs data are reanalyzed data from the Climatic prediction Centre (CPC)/NOAA, U.S.A.

Though, the length of the rainfall records varies with each station having at least 50-year record, data availability on El-Nino/Southern Oscillation (ENSO) index and sea surface temperature (SST) anomalies appears to be the major factor that constrained the length of the data used in this study. The available El-Nino/Southern Oscillation (ENSO) index and sea surface temperature (SSTs) data during the period of data collection for this study are from 1951-2000. Thus, for the purpose of uniformity in the length of data, this study made use of 50 years data (1951-2000) in describing the nature of the interaction between the observed anomalies in rainfall over Nigeria and the coupled ocean-atmospheric phenomena of El-Nino/Southern Oscillation (ENSO) and Sea Surface Temperature (SST) anomalies.

Indices based on sea surface temperature or, more often, (its departure from the long-term average) are those obtained by simply taking the average value over some specified region of the ocean. There are several regions of the tropical Pacific ocean that have been highlighted as being important for monitoring and identifying El-Nino and La-Nina. The most common ones are the NINO regions. The boundaries of these oceanic regions have been shown by Folland *et al.* (1986).

NINO 1+2 (Extreme Eastern Tropical Pacific) (0-10^{0}S, 80-90^{0}W): This is region that typically warms first when an El-Nino event develops.

NINO 3 (Eastern Tropical Pacific) (5^{o}S-5^{0}N; 150^{0}W-90^{0}W): This is the region of the tropical Pacific that has the largest variability in sea surface temperature on El-Nino time scales.

NINO 3.4 (East Central Tropical Pacific) (5^{0}S-5^{0}N; 170^{0}W-120^{0}W): This is the region that has the largest variability on El-Nino time scales, and that is closer (than NINO 3) to the region where changes in local sea surface temperature are important for shifting the large region of rainfall typically located in the far western Pacific.

NINO 4 (Central Tropical Pacific) (5^{0}S-5^{0}N; 160^{0}E-150^{0}W): This is the region where changes of sea surface temperature lead to total values around 27.5^{0}C, which is thought to be an important threshold in producing rainfall.

However, If the concern regarding El-Nino and La-Nina is the subsequent effect of that tropical Pacific variability on the climate in a particular region, then one index may be useful than the others. For widespread global climate variability, NINO 3.4 is generally preferred, because the sea surface temperature variability in this region has the strongest effect on shifting rainfall in the western Pacific. And in turn, shifting the location of rainfall from western to central Pacific modifies greatly where the location of the heating that drives the majority of the global atmospheric circulation (Folland *et al*, 1986; Palmer,1986).

The reliability of the records, the continuity of the record and date of establishment coupled with respective location of each synoptic station are considered in selecting the stations for this study. The names of the synoptic stations from Nigeria are presented in table 3.1 (see also Fig. 3.1) while table 3. 2 presents the names of the synoptic stations from the republic of Niger whose monthly rainfall and dew point temperature data for the period 1961-2000 were also used in this study.

Table 3.1. The meteorological stations in Nigeria used for this study

Station Code	Station Name	Lat.^{0} N	Long.^{0}E	State	Altitude (m)	Year Established
65001	Yelwa	10.53	04.45	Kebbi	244.00	1926
65010	Sokoto	13.01	05.15	Sokoto	350.80	1904
65015	Gusau	12.10	06.42	Zamfara	463.90	1952
65019	Kaduna	10.36	07.27	Kaduna	645.40	1913
65028	Katsina	13.01	07.41	Katsina	517.60	1922
65030	Zaria	11.06	07.41	Kaduna	110.90	1968
65046	Kano	12.03	08.12	Kano	472.50	1905
65055	Bauchi	10.17	09.49	Bauchi	609.70	1906
65064	Nguru	12.53	10.28	Yobe	343.10	1942
65082	Maiduguri	11.51	13.05	Borno	353.80	1909
65101	Ilorin	08.29	04.35	Kwara	307.40	1905
65112	Bida	09.06	06.01	Niger	144.30	1928
65123	Minna	09.37	06.32	Niger	256.40	1914
65134	Jos	09.52	08.45	Plateau	NA	1921
65167	Yola	09.14	12.28	Adamawa	186.10	1904
65203	Lagos	06.27	03.24	Lagos	14.00	1892
65208	Ibadan	07.26	03.54	Oyo	227.2	1905
65215	Oshogbo	07.47	04.29	Osun	302.0	1935
65222	Ondo	07.06	04.50	Ondo	287.3	1906
65229	Benin	06.19	05.06	Edo	77.80	1906
65236	Warri	05.31	05.44	Delta	6.10	1907
65243	Lokoja	07.47	06.44	Kogi	62.50	1901
65250	Port harcourt	04.51	07.01	Rivers	19.50	1915
65257	Enugu	06.28	07.33	Enugu	141.8	1916
65264	Calabar	04.58	08.21	C/ river	61.90	1901
65271	Makurdi	07.44	08.32	Benue	112.90	1926
65273	Ikom	05.58	08.42	C/river	119.00	1972

Source: Nigerian Meteorological Agency, Oshodi, Lagos.

Table 3. 2. The meteorological stations in the Republic of Niger used in this study.

S/No.	Station Name	Lat.^{0}N	Long.^{0}E	Altitude	Year Established
1	Agadez Aéro	16.97	07.98	501.02m	1/1/1921
2	Bilma	18.68	12.92	355.37m	1/1/1923
3	N'Guigmi	14.25	13.12	286.50m	1/1/1921
4	Tahoua Aéro	14.90	05.25	385.81m	1/1/1922
5	Tillabéry	14.20	01.45	209.10m	1/1/1923

Source: National Meteorological Service, Niamey, Republic of Niger.

This space is for figure 3.1

Given the broad latitudinal expanse of Nigeria, it is considered necessary to divide Nigeria into at least three regions: The South, the Middle-belt (Cental) and the North. This is done on the basis of Olaniran's (1986) classification of tropical climates for the study of regional climatology using Nigeria as a case study. His work on climatic grouping of Nigeria into seven regions is used and re-group the climatic regions into three broad regions with particular reference to their prevailing air masses. The two principal air masses that control the climate of West Africa are the tropical maritime air mass (mT) and tropical continental air mass (cT). The southern region was further divided into two namely: the South-West and South-East (see Fig. 3.2)

This space is for figure 3.2

Olaniran's [1986] region i and ii are grouped as southern region in the present study because the region is dominated by the mT airmass.This southern region is aligned in the west-east direction and lies below latitude 7^{o} 30^{1}N.

The regions iii and iv of Olaniran's (1986) classification are taken to form the middle belt (central) region in this study. The climates of the middle belt are transitional between those of the southern and northern regions. The region experiences both the mT and the cT air masses.

The northern region of the present study comprises Olaniran's (1986) regions v, vi and vii. The region experiences the cT air throughout the year except between July and September when the mT air dominates this part of the country. Although, the climate of Jos Plateau (Olaniran's 1986 region v) and Kaduna (Olaniran's 1986 vii) areas are affected by increase in altitude and low annual rate of evaporation respectively, they are grouped as part of northern region in this study because of their latitudinal locations.

In Nigeria, the acquisition of long-term rainfall data from the 27 stations permits a regional analysis to examine the variation in rainfall patterns for a 100-year period (1901-2000). Olaniran's (1986) regions i and ii which are grouped as southern region in the present study, is further divided into south-west and south-eastern regions. Hence, the country is divided into four regions (the North, the Central, the South-West and the South-East) which are characterized by distinct rainfall patterns. The grouping of the stations into regions is shown in table 3.3.

Table 3.3. Nigeria rainfall stations grouped by region.

Region	Station	Duration of rainfall record (years)
North	Sokoto	83
	Kano	95
	Katsina	76
	Yelwa	62
	Yola	86
	Bauchi	89
	Gusau	48
	Maiduguri	85
	Nguru	59
Middle Belt (Central)	Kaduna	63
	Zaria	58
	Minna	85
	Bida	73
	Jos	79
	Lokoja	83
	Makurdi	72
	Ilorin	85
South West	Lagos	99
	Ibadan	96
	Ondo	95
	Oshogbo	57
	Benin	95
	Warri	93
South East	Port Harcourt	96
	Enugu	81
	Calabar	97
	Ikom	29

SOURCE: Nigerian Meteorological Agency, Oshodi, Lagos

The northern region consists of nine stations which are representative of the Sudano-Sahelian zone. The central region, which include eight stations, separates the semi-arid zone of northern Nigeria from the sub-humid tropical zone of the south. The climate of the central region is complicated by the topographical features of the highlands.

Six stations are grouped into the south-western region which experiences double-rainfall maxima in May/June and September/October with a particularly well-marked “little dry season” in July/August (Griffiths, 1972; Ayoade, 1973, 1974; Olaniran, 1988a, b; Adejuwon and Odekunle, 2006). The little dry season is associated with a temperature inversion above the surface layer which inhibits convective activity and rainfall (Nieuwolt, 1977; Adefolalu, 1986; Olaniran, 1991a). Local divergence is primarily responsible for the stability in the monsoon air masses during this period. In the south-east region, which is represented by four synoptic stations, there is a tendency for somewhat less rainfall in August than during July and September; however, rainfall amounts remains considerably high throughout the entire wet season. The high rainfall amounts within the south-eastern region is common features of the more humid equatorial zone. The influence and contributions of world’s ocean SST anomalies and El-Nino/Southern Oscillation (ENSO) events to the observed anomalies in rainfall over Nigeria are analyzed and the spatial-temporal effects of these factors on Nigerian rainfall records also examined.

3.2. Limitation(s) of the archival data and homogeneity testing

The length of period that the data record covers has some effects on the statistical measure, especially the measures of central tendency and measures of dispersion such as the mean, median, standard deviation e.t.c; because the longer the period of record, the more representative the true values of the measure would be. Jumps or occurrence of gaps in record, as a result of one problem or the other such as inadequate field officers, improper record keeping/storage, and the impact of the civil war in Nigeria between 1967 and 1970 are great factors that affect the reliability of the data. This is a major limitation of the study because of the discontinuity of the record. Also irregularity in the postage of records from various stations to the meteorological services central office is another limitation. Changes in instrumentation, calibration, and exposure and station relocation are additional problems associated with long climatic records (Jones *et al*, 1986).

Climate data could provide a great deal of information about the atmospheric environment that impact almost all aspects of human endeavor. For example, these data have been used to determine where to build homes by calculating the return periods of large floods, whether the length of a frost-free growing season in a region is increasing or decreasing, and the potential variability in demand for heating fuels. However, for these and other long-term climate analyses-particularly climate change analyses-to be accurate, the climate data used must be as homogeneous as possible. A homogeneous climate time series is defined as one where variations are caused only by variations in climate.

Unfortunately, as stated earlier, most long-term climatological time series have been affected by a number of non-climatic factors that made these data unrepresentative of the actual climate variations that occurred over time. These factors include changes in instruments, observing practices, station locations, formulae used to calculate means and station environment. Some changes caused sharp discontinuities while other changes, particularly change in the environment around the station caused gradual bias in the data. All of these inhomogeneities could bias a time series and lead to misinterpretations of studied climate. It is important, therefore, to remove the inhomogeneities or at least determine the possible error they might caused.

Many researchers have put a great deal of effort into developing ways to identify non-climatic inhomogeneities and then adjust the data to compensate for the biases these inhomogeneities produced. Several techniques have been developed to address a variety of factors that impact climate data homogenization such as the type of element (temperature versus precipitation), spatial and temporal variability depending on the part of the world where the stations are located, length and completeness of the data and double mass curve analysis technique (see Farmer and Wigley, 1984; Van and Francis, 1999; Ayoade, 2008).

Although the best way to ensure homogeneity is to keep the record homogeneous through appropriate management of the observation site and associated equipment, this is very difficult to achieve. Besides, because it is almost impossible to be 100% sure about the quality of the past data, a homogeneity assessment is always recommended. It is however, interesting to note that, the Nigerian meteorological stations whose climatic records were used in this study have not suffered from the relocation of sites as put forward by Adejuwon *et tal* (1988). This shows that factors like changes in calibration of instruments, observation practices and stations' site have not affected the stations' data used in this study, and the climatic records used are therefore homogeneous to a greater extent (see table 3.3).

In the present study however, two stations from each of the broad regions are randomly selected for homogeneity test using the double mass curve analysis. The stations selected and whose annual rainfall series are tested include Minna and Ilorin in the north, Lagos and Ibadan in the south. The double mass curve analysis is used to test for consistency of precipitation data used in this study. Lack of consistency or homogeneity in precipitation records may be due to environmental changes in the vicinity of the guage or changes in methods of observation (Ayoade, 1988). To test for consistency in rainfall records, the cumulative precipitation totals at Minna and Lagos stations respectively are plotted on the arithmatic graph paper against the cumulative totals at Ilorin and Ibadan stations respectively over the period 1921-2000 when all the stations had adequate rainfall records. The records tested are found to be consistent since the plotted points are linear (See Fig. 3.3*a* and 3.3*b*). However, if the records tested are not homogeneous, there will be one or more breaks in the slope of the double mass curve. The results of the homogeneity testing presented in figures 3.3*a* and 3.3*b* for the selected stations can serve as representative of the stations in the country. Hence, the annual rainfall series of Nigerian meteorological stations used in this study can be said to be homogeneous.

This space is for figure 3.3*a*

This space is for figure 3.3*b*

3. 3. 0. Methods of data analyses

Before carrying out any analysis, the consistency of rainfall records is checked for all the synoptic stations in this study for the period 1901-2000 (100 years) and records are found to be homogeneous. This is carried out using the double mass curve analysis. Thereafter, the monthly rainfall, rain days, dew point temperature, southern oscillation index and sea surface temperature anomalies are subjected to a preliminary analysis using the version 15 of the SPSS package and computes the annual and monthly mean values of all the variables for further in-depth analysis of the data.

The annual rainfall variability in Nigeria is explained in terms of their decadal statistics- Mean, Standard Deviation, Coefficient of variation, and difference from long-term average and decadal changes. This is because a useful way to express long-term changes in a climatic time series is to examine the frequency of the short-term anomalies in the series (Verma *et al*, 1985). Therefore, the period between 1901-2000 is divided on the decadal basis as follows: 1901-1910, 1911-1920., 1921-1930, 1931-1940, 1941-1950, 1951-1960, 1961-1970, 1971-1980, 1981-1990, 1991-2000. The relative changes in annual rainfall in Nigeria are then analyzed for the ten sub-periods.

The inter-annual and inter-decadal variability in annual rainfall in Nigeria over the period 1901-2000 are also analyzed using the coefficient of variation (C.V). The coefficient of variation is the ratio of standard deviation to the arithmetic mean and is usually expressed in percentage terms. It is given as:

$$CV = \frac{\sigma}{\overline{x}} X100\% \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots \; 3.2$$

where $\overline{X}$ is the mean of the entire series and σ is the standard deviation from the mean of the series.

One of the objectives of this study is to identify trends in the time series of annual rainfall and rain days in Nigeria over the period 1901-2000. The linear regression method and Mann-Kendall *tau* methods seem most appropriate because they can detect both linear and non-linear trends in climatic time series. All graphs in this study are created with SPSS version 15 and maps displayed in this study are generated using the Kriging interpolation method with ArcGIS spatial tool of analysis. The Kriging method has an advantage over other interpolation techniques because it incorporates the spatial correlation of data while other interpolation techniques do not.

The isopleth maps presented in this work have been objectively drawn using the map computer program. The program works on the basis of given coordinates of data points and the values associated with them. Continuous variation is assumed between two isopleths lines. The applied methods of analysis are detailed below

3.3. 1. Simple linear regression model

An experimental study of the relation between two variables is often motivated by a need to predict one from the other. Regression is a statistical technique that allows decision makers not only to establish quantitative relationships among such variables but also measure the "strength" of the relationship. In regression analysis, a mathematical equation is developed which relates an unknown variable to a known quantity of interest. The known variable (s) is (are) the independent (explanatory, predictor) variable represented by x, while the variable to be predicted is the dependent (or response) variable represented by y. The object is to find the nature of the relation between x and y from experimental data and use the relation to predict the response variable y from the input x. Naturally, the first step in such a study is to plot and examine the scatter diagram. If a linear relationship emerges, the calculation of the numerical value of r would confirm the strength of the linear relationship. Its value indicates how effectively y can be predicted from x by fitting a straight line to the data. The coefficient of determination, or r^2 , expresses the strength of the relationship between the X and Y variables. It is the proportion of the variation in the Y variable that is "explained" by the variation in the X variable. The coefficient r^2 could vary from 0 to 1; value near 1 means that the Y values fall almost right on the regression line, while value near 0 means that there is very little relationship between X and Y. A line is determined by two constants; its height above the origin (intercept) and the amount that y increases whenever x increases by the unit (slope).

(see Wigley, 1984; Van and Francis, 1999; Ayoade, 2008).

The linear regression equation is given as

$$Y = a + bx \text{ --------------- } \quad 3.3$$

Where a = intercept of the regression, b = regression coefficient or slope of the regression.

Annual rainfall and rain days totals of the respective stations constitute the dependent variable (y) while Time, (here years) constitutes the independent variable (x). Trends are determined using version 15 of SPSS package.

Trends in time-series data (total annual rainfall and rain days) are therefore analyzed using simple linear regression. The slope b indicates the average rate of change in the climatic parameter over the time period. One advantage of this method is that it is easy to apply to a large number of sites. Its disadvantage is that it does not detect trends that are non-linear but still monotonic (generally in one direction). Other methods, such as the Mann-Kendall test, could be used to detect trends that are monotonic but not necessarily linear, but these only indicate the direction, and not the magnitude of trends. The linear regression model is also used to assess the influence of the Inter-tropical Discontinuity (ITD) on rainfall distribution patterns in Nigeria. The mean monthly surface positions of the ITD constitute the independent variable while the mean monthly rainfall constitute the dependent variable. The detailed procedure of the linear regression model has already been given above.

3. 3. 2. The Mann-Kendall's *tau* method.

The most likely alternative to randomness in a climatic time series is some form of trend, which may be linear or non-linear. It is, therefore, necessary to use a test of randomness to check the trend. The Mann-Kendall rank method has been suggested as powerful test by Kendall and Stuart (1961), and hence adopted for this study, in addition to the linear regression method discussed earlier. In this case, the statistic t, also known as *tau* statistic is computed using the formula:

$$t = \frac{4\varepsilon ni}{N(N-1)} - 1 \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots \quad 3.4$$

Where ni is the number of values larger than the ith value in the series subsequent to its position in the time series. The value of t was tested for significance by the statistic $(t)_t$, which was given by:

$$(t) \;=\; tg\left(\frac{4N+10}{9N(N-1)}\right)^{½} \quad \text{-----------------} \quad 3.5$$

Where t_g is the value of t at the appropriate probability point in Gaussian distribution corresponding to the desired level of significance.

3. 3. 3. Step wise multiple regression and correlation analyses.

The second hypothesis which states that the observed anomalies in rainfall over Nigeria could be explained as a function of El-Nino/Southern Oscillation

(ENSO) index and Sea Surface Temperature (SSTs) anomalies is tested using a series of regression models.The statistical procedure employed in the selection of the significant explanatory variables among the hypothesized set of explanatory variables is stepwise multiple regression procedure. Diraper and Smith (1966) and Vans and Francis (1999) have evaluated this procedure of regression analysis as the best among others. The most important feature of the stepwise multiple regression model is that a significant variable, which has been added at an earlier stage, may later be considered insignificant and thus deleted. At the end of the search, it is only the most significant variables which account for the largest portion of the total variance in the dependent variable (y) that are retained in the regression equation (Ayeni, 1979; Vans and Francis, 1999; Ayoade, 2008).

The independent variables used in the stepwise multiple regression models include: (1) The ENSO index, (2) Global SST anomalies (3) NINO 1+2 SST anomalies (4) NINO 3 SST anomalies (5) NINO 3.4 SST anomalies (6) NINO 4 SST anomalies (7) South Atlantic SST anomalies (8) North Atlantic SST anomalies. The dependent variables include: Annual Rainfall anomalies for Nigeria; Annual rainfall anomalies for the North; Annual rainfall anomalies for the Central; Annual rainfall anomalies for the South-West and Annual rainfall anomalies for the South-Eastern region. Direct regressions are made between the stated independent variables and each of the stated dependent variables for the period 1951-2000 when all the 27 synoptic stations considered in this study have adequate records.

3. 3. 4. Standardized Precipitation Index (SPI)

The Standardized Precipitation Index (SPI) is employed in this study for analyzing droughts in Nigeria. Different indices have been developed through the years to quantify drought intensity. Such indices are usually based on the precipitation deviation from the mean for a given period. The most commonly used indices include: Palmer Drought Severity Index (PDSI), Crop Moisture Index (CMI), and Surface Water Supply Index (SWSI). The PDSI, developed by Palmer (1965) requires observed precipitation, temperature and available soil water content of the soil region. In addition to those four main parameters, there are other complicated numerical computation steps which are required to get 25 parameters in order to reach the final PDSI estimate. Despite its complexity, the PDSI becomes the most prominent index for determining meteorological drought around the world. However, Palmer’s (1965)

Drought Severity Index suffers criticisms due to its excess data requirement and for being mathematically complex (Alley, 1984; Oladipo, 1985).

After nearly three decades, a new drought index is introduced by McKee *et al* (1993) called the Standardized Precipitation Index (SPI) to quantify water deficits on different time scales. The SPI is based on the standardized precipitation. Standardized precipitation is simply the difference of precipitation from the mean for a specified time period divided by the standard deviation where the mean and the standard deviation are determined from the past records. This same method could also be used to evaluate variations in any of the five usable water sources of any area with long precipitation records. The basic approach is to use standardized precipitation for a set of time scales which together represents water sources of several types.

The SPI is calculated through the following sequence. A monthly precipitation data set is prepared for a period of *m* months, ideally a continuous period of at least 30 years. A set of averaging periods are selected to determine a set of time scales of period *i months* where *i* is 3, 6, 12, 24, or 48 months. These represent arbitrary but typical time scales for precipitation deficits that affect the five types of usable water sources: snow pack, stream flow, reservoir storage, soil moisture and ground water.

The SPI is given below:

$$SPI = \frac{x_{ik} - \bar{x}}{\sigma_i} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots \; 3.6$$

Where:

σi =standardized deviation for the ith station

X_{ik}=is the precipitation for the ith station and kth observation.

$\bar{X}$ = mean precipitation for the ith station.

When the time periods are small (3 or 6 months), the SPI moves frequently above and below zero. As the time period is lengthened to 12, 24, and 48 months, the SPI responds more slowly to changes in precipitation. Periods with the SPI negative and positive become fewer in number but longer in duration. A drought event for time scale *i* is defined as a period in which the SPI is continuously negative and the SPI reaches a value of –1.0 or less. The drought begins when the SPI first falls below zero and ends with the positive value of SPI following a value of -1.0 or less. A 12-month SPI scale is adopted for the present study because of its suitability for agricultural/meteorological droughts analyses. Further details of the method are given and intensively discussed in the literature (McKee et al, 1993, 1995; Bornaccorsa et al, 2003; Sonmez et al, 2005).

Drought intensity is derived as for values of the SPI in the following categories as shown in table 3.4.

Table 3.4. SPI drought category (After McKee *et al.*, 1993)

SPI VALUES	DROUGHT CATEGORY.
0 to -0.99	Mild drought
-1.00 to –1.49	Moderate drought
-1.50 to –1.99	Severe drought
≤ -2.00	Extreme drought

The SPI has found a wider application area in different scientific disciplines due to its simplicity and user friendliness, and its great success in describing drought severity and spatial extent. Its major strength stemmed from the possibility to compare drought events in regions and areas with different climates, taking into account the different time scales at which the component of the hydrological cycle were affected by precipitation deficits. It did, however, assumed that the data were normally distributed, and this could introduce complication for shorter time periods. Hence the SPI technique was adopted for this study.

3. 3. 5. The Percentage Deviation below the Mean (PDBM) index

This is another index used to quantify the severity or intensity of drought in climatological studies (see Ayoade, 2008). The PDBM index is found to have a wider application in the study of climatological droughts (Oladipo, 1985, 1995). It is used in this study for analyzing droughts in Nigeria coupled with the SPI index and their performances compared. This is because climatological droughts can also be analyzed using the percentage deviation below the mean. The index is calculated using the following formula:

$$\frac{xi - \bar{x}}{\bar{x}} \times 100\% \text{ ------------------- } 3.7$$

where xi is observed precipitation series

$\bar{x}$ is the mean of the entire series.

Then droughts of various intensities could be read using the scales presented in table 3.5.

Table 3.5. Drought category based on the PDBM index

Scale	**Drought Category**
11-25%	Slight Drought
26-45%	Moderate Drought
46-60%	Severe Drought
More than 60%	Disastrous Drought

Source: Ayoade (2008)

3. 3. 6. Normal frequency distribution function

With a large and normally distributed rainfall data, the normal frequency distribution function is used to calculate the percentage probability that some critical rainfall amount will be exceeded or will not be reached. The procedure followed involves the computation of the Z-Score (z) using the formula:

$$Z = \frac{Xc - \overline{X}}{\sigma} \quad \text{---------------------} \ 3.8$$

Where Xc is the critical rainfall amount (hereafter referred as the least/highest annual rainfall ever recorded as indicated in table 10.4 and table 11.4 respectively). The $\overline{X}$ and σ are the mean and standard deviation respectively. The Z-score indicates the extent to which the critical value differs from the mean in terms of 'so many' standard deviations. The percentage probability that the rainfall will be more or less than the Z-score corresponding to the critical value is obtained from the table of the normal distribution function.

When the Z-score is negative, it indicates the number of standard deviations that the critical value is below the mean. The percentage probability is that the occurrence will be less than the corresponding value of z. The probability that it will be more than this value is (100 - %). Where the z is positive, it represents the number of standard deviations that the critical value is above the mean. The corresponding percentage probability will be that the occurrence will be more than the critical rainfall amount. The probability that it will be less than the critical value will be 100 - % probability corresponding to the z-score. A guiding principle is that the mean has a 50% probability of being received in a given year. So, any rainfall amount higher than the mean will have lower probability of occurrence and vice-versa (see Gregory, 1965; Ayoade, 2008).

3. 4. 0. Significance and hypothesis testing

All trends considered in this study are tested for 'significance'. In statistics, a result is called significant if it is unlikely to have occurred by chance. In statistical testing, the significance level of a test is a maximum probability, assuming the null hypothesis, that the statistic would be observed. Hence, the significance level is the probability that the null hypothesis would be rejected in error when it is true (a decision known as type 1 error, or "false positive"). The significance of a result is also called its p-value; the smaller the p-value, the more significant the result is.

Significance is usually represented by the Greek symbol, a (alpha). Popular levels of significance include 5%, 1% and 0.1%. If a test of significance gives a p-value lower than the a-level, the null hypothesis is rejected. Such results are informally referred to as ‘statistically significant’. The 0.05% and 0.01% levels of significance are adopted throughout this study.

CHAPTER FOUR

FREQUENCY AND SPATIAL PATTERNS OF DROUGHTS IN NIGERIA

Several indices exist for quantifying drought in a given region amongst which include Palmer's (1965), Roy's (1974), Bhalme and Mooley's (1986). The commonly used index is the Palmer's index. It has, however, suffered criticisms due to its excess data requirements (Girux, 2000; Hilman, 2001). MCKeee *et al* (1993) developed a new technique for quantifying droughts known as Standardized Precipitation Index (SPI). This technique has been adjudged as the best technique for determining droughts of various intensities in the sense that it requires only rainfall data which is readily available (Thomson, 1999; Thrix, 2001). It allows a regional comparison of drought events and hence employed in this study coupled with the percentage deviation below the mean (PDBM) index. For detailed computational procedure of these drought indices (see MCKeee *et al*, 1993; Ayoade, 2008).

4. 1. Rainfall fluctuations in Nigeria

Rainfall as a spatially discontinuous variable has inherent oscillations which can mar any meaningful trend determined in the rainfall data for drought analysis. Several methods can be used to filter or smooth out the inherent fluctuations in rainfall data. A review of such methods is provided by Vans and Francis (1999) and Ayoade (2008). However, a 5-year moving average is used in the present study to smooth out the inherent fluctuations in annual rainfall amounts for all the selected stations in this work. The results are shown in figure 4.1.

This space is for figure 4.1

This space is for figure 4.1

This space is for figure 4.1

This space is for figure 4.1

This space is for figure 4.1

The mean annual rainfall at Bauchi from 1908 to 2000 is 1066mm, with the highest value of 1493mm recorded in 1946. The least value of 725mm is recorded in 1985. The 5-year moving average for Bauchi shows a decline in total annual rainfall in the late 1940s, with an improvement in the 1950s. However, rainfall continuous to decline in Bauchi from late 1960s through 1980s. The decline in rainfall from late 1960s through 1980s coincides with the great Sudano-Sahelian droughts of 1970s and 1980s which ravaged the West African Sahel (Wistanley, 1973a; Landsberg, 1975; Bunting *et al*, 1976; Raymond, 1986 and Nicholson, 1989).

The results for Jos show an appreciable increase in annual rainfall from 1932 to 1939. Thereafter, rainfall decreases in the 1940s and alternated with the wet years in the 1950s reaching an annual magnitude of 167mm in 1957. Rainfall continuous to decrease from late 1960s through 1980s, reaching another peak of 1176mm in 1973. The mean annual rainfall at Jos from 1922 to 2000 is found to be 1332mm. At Kano, it indicates an increase in annual rainfall from 1908 to 1912 reaching a peak of 1200mm in 1911. There is, however, a continuous decline in rainfall from the 1940s to 1990s, reaching a minimum of 416mm in 1973. Thereafter, the annual rainfall continuous to increase reaching a maximum of 1139 in 2000. The mean annual rainfall at Kano from 1905-2000 is 844mm.

The results for katsina are marked with oscillations in mean annual rainfall with a general decline in annual rainfall reaching a peak of 398mm in 1987. The mean annual rainfall at Katsina from 1918-2000 is 632mm. At Kaduna, it is also marked by oscillations in annual rainfall. There is, however, a remarkable increase in rainfall in 1950s reaching a maximum in 1955, with an annual rainfall of 1692mm. Thereafter, rainfall continuous to decline from late 1960s through 1980s reaching a peak in 1989, with an annual rainfall of only 996mm and mean annual rainfall of 1246mm over the period 1930-2000.

The highest annual rainfall amount recorded at Sokoto from 1916-2000 is 1025mm in 1936; marked with oscillations in annual rainfall amounts with a general decline in rainfall from 1940s through the 1980s, with the exception of the wet period of 1950s. The least annual rainfall of 373mm is recorded in 1987; the mean annual rainfall is 705mm. At Maiduguri, the results show a general decline in rainfall from 1920s through the 1980s, with the exception of the 1930s, which show an appreciable rise in annual rainfall reaching a peak of 804mm in 1936. The lowest annual rainfall of 263mm is recorded in 1983. The mean annual rainfall at Maiduguri from 1915-2000 is 622mm.

At Yola, it is also marked by fluctuations with the highest annual rainfall of 1369mm recorded in 1941. The least annual rainfall of 336mm is recorded in 1968, which also coincides with the onset of Sudano-Sahelian droughts between 1968-73. There is, however, an appreciable rise in annual rainfall in the 1990s reaching a peak 1042mm in 1998. The mean annual rainfall at Yola from 1914-2000 is 926mm. The annual rainfall amount at Minna decreases between the 1920s and 1940s. There is, however, a sharp rise in the annual rainfall in the 1950s reaching a peak of 1562mm in 1957. Thereafter, rainfall continuous to decline from late 1960s through the 1980s reaching a maximum in 1983, with an annual rainfall of only 846mm. The mean annual rainfall at Minna from 1916-2000 is 1284mm. The results for the other northern stations also show similar pattern of rainfall fluctuations and a general decline in rainfall as shown in figure 4.1.

Lagos is also marked by oscillations in annual rainfall amounts. The highest annual rainfall amount recorded from 1901-2000 is 3265mm in 1968. However, annual rainfall amount decreases through 1970s and 1980s reaching a peak of 736mm in 1982. The mean annual rainfall at Lagos over the study period is 1813mm. The highest annual rainfall amount recorded at Calabar from 1902-2000 is 4860mm in 1927, with the least annual rainfall amount of 2110mm recorded in 1973. The mean annual rainfall at Calabar from 1902-2000 is 2995mm. The results of the moving average show fluctuations in the annual rainfall amount at Calabar, with a general decline in rainfall from the 1970s through the 1980s.

At PortHarcourt, the results show periodic oscillations in annual rainfall amounts, with a sharp decline in rainfall from the late 1950s through the 1980s. The highest annual rainfall amount recorded at PortHarcourt from 1902-2000 is 4252mm in 1905 and the least annual rainfall amount of 1683mm is recorded in 1958. PortHarcourt has a mean annual rainfall of 2696mm from 1902-2000. Enugu is also marked by fluctuations in annual rainfall, with a general decline in annual rainfall amounts from the late 1960s through the 1980s, reaching a peak in 1983. The highest annual rainfall amount recorded at Enugu from 1916-2000 is 4206mm in 1918, with the least annual rainfall amount of 913mm recorded in 1983. Enugu has a mean annual rainfall of 1792mm from 1916-2000. The results for the other southern stations also indicate a similar pattern of rainfall fluctuations as revealed in Fig. 4.1.

The results of a 5-year moving average of annual rainfall (Fig. 4.1) show that nearly all the stations experience a decline in total annual rainfall towards the beginning of the 1970s. The declines are prominent among the northern stations such as Sokoto, Maiduguri, Bauchi, Kano, Minna, Katsina and Kaduna. In the early 1970s, rainfall amount exhibits significant downward trend in all these stations. This period corresponds to the drought period in Nigeria (Kowal and Kassam, 1973; Mortimore, 1973; Ayoade, 1977). Consistent decrease in rainfall amount towards the later part of the 1970s and 1980s is another common characteristic of the pattern observed in the study. Rainfall decreases in virtually all the stations in Nigeria within the period 1970s through 1980s. There are some similarities in the pattern of trend in rainfall fluctuations in some stations while some others are entirely different. For instance, the pattern observed in Jos and Kaduna is similar. The prominent irregularity in their rainfall pattern can only be explained by the presence of highlands (Ayoade, 1974).

A comparison of the rainfall pattern between the south-western stations and south-eastern stations shows that rainfall tends to increase from the south-west to the south-eastern part of the country due to the occurrence of little dry season phenomenon in the west. This is as a result of a strong inversion layer developed over the south-western Nigeria, thereby causing rainfall to increase from the west to the eastern part over southern Nigeria during 'the July/August break' also known as 'little dry season' in south-western Nigeria (see Ayoade, 1974; Olaniran, 1988a,). Other factors such as upwelling of cold waters at the Gulf of Guinea coast or stabilization of the south-westerlies caused by the southern Africa sub-tropical high pressure may have acted to reinforce the dryness of this period (Ireland, 1962; Ojo, 1977). On the other hand, the rainfall pattern in Calabar could be attributed to the influence of Cameroun Mountains (Adefolalu, 1986). The relief factor provides a trigger action for convection (Ayoade, 1974) and helps to bring an increase in rainfall when the meteorological conditions are favourable in the region.

The observed pattern of rainfall fluctuations as shown by the 5-year moving average (Fig. 4.1) is consistent with the result of the averaged regional annual rainfall fluctuations in Nigeria (Fig. 4.2*a*). Figures 4.2*a* and *b* present separately the regional averaged annual rainfall and raindays fluctuations for the four regions in Nigeria (i.e the North, Middle belt, South-West and South-East) over a period 1901-2000. The annual rainfall amounts demonstrates differences in magnitude between regions. However, a cautionary note is necessary before discussing these results. The station data are screened for quality and consistency, and some stations (Gusau and Ikom) are excluded from the analysis due to inadequate data. Inconsistency in rainfall record results from either a change of method of observation or relocation of site (Fasheun, 1984), though Adejuwon (1988) shows that the Nigerian meteorological stations, whose data were used in this study, have not suffered from the relocation of site. An examination of the time series in figure 4.2*a* suggests different patterns (i.e., trends, variability) before and after the beginning of the wet period of the 1950s (i.e., 1951).

The northern region receives the least amount of annual rainfall and is, therefore, the most drought-prone. Figure 4.2*a* (northern region) shows a relatively long period of below mean annual rainfall in the 1940s, which is interrupted only by two years of relatively high rainfall in 1945 and 1946; and another low rainfall period beginning in 1963 and lasting through 1987. (These two below-normal rainfall periods will hereafter be referred to as the drought periods). These rainfall series are similar to those reported by Ayoade (1973, 1974), Bunting *et al* (1976), Adefolalu (1986), Adejuwon *et al* (1990) and Hess *et al* (1995).

The corresponding rainfall time series for the central region in figure 4.2*a* shows similar tendencies towards below-normal rainfall during these two periods but to a lesser degree. Examination of the rainfall records for the southern regions shows no consistent year-to-year pattern compared to the rainfall time series for the northern region. There are years within this drought period, however, which show below-normal rainfall amount throughout most of Nigeria (i.e., 1926-28, 1942, 1943, 1944, 1948, 1971, 1972, 1973, 1976, 1982, 1983, 1984, 1987 and 1988). Similarly, the analysis of annual rainfall time series for the four regions shows a downward trend in the north ($\beta = -0.17$, $p<0.001$), middle belt ($\beta = -0.27$, $p<0.001$), south-west ($\beta = -0.001$, $p<0.97$) and south-east ($\beta = -0.37$, $p<0.001$) while, annual rain days anomalies displays a different pattern with a declining trend in the north ($\beta = -0.22$, $p<0.001$), middle belt ($\beta = -0.04$, $p<0.001$), south-west ($\beta = -0.49$, $p<0.001$) and upward trend in the south-east ($\beta = 0.25$, $p<0.001$) (see figures 4.2*a* and *b*). These results suggest a strong association between abnormal rainfall and anomalous large-scale circulation pattern in this sub-tropical region during these years, especially since rainfall data for the 1971-1973 period indicate widespread drought conditions throughout the entire West and Central African regions (Burning *et al*, 1976; Kidson, 1977; Nicholson, 1978). In fact, Kidson (1977) shows that the low rainfall in the Sahel during the peak wet season is associated with the virtual disappearance of the 850mb trough near 8^0 N and the weakening of the easterly jet at 200mb.

Nicholson (1981) examines 37 century-long regional rainfall departure series in sub-Saharan Africa and suggests that a northward displacement of the Intertropical Convergence Zone (ITCZ) may account for wetter years, but that a weakened "intensity" of the rainy season, independent of ITCZ position, is the most likely cause of drought in the sub-Saharan region. This and the tendency for synchronous fluctuations north and south of the Sahara (i.e., tropical and extratropical regimes) suggest that changes in intensity of the Hadley circulation may be an important factor in West Africa rainfall fluctuations.

This space is for figure 4.2*a*

This space is for figure 4.2*b*

The annual rainfall data for the four regions of Nigeria, expressed in terms of departures from the long-term mean of the respective regions, i.e, 100-year normals (1901-2000) were presented in table 4.1 and figure 4.3, with the drought periods of the 1920s, 1940s, 1968-76 and 1980s identified. Table 4. 1 and figure 4.3 show an interesting feature of Nigerian climate with respect to comparisons between the northern and southern regions. Notably, during the years 1912, 1914, 1915, 1916, 1936, 1938, 1945, 1945, 1946, 1950, 1958 and 1964, the northern region experiences above-normal rainfall while southern regions experience below-normal rainfall. Conversely, during the years 1941, 1947, 1949, 1968 and 1969 the opposite pattern prevails in the southern regions. These conditions are believed to be associated with the ITD as discussed below.

Trewartha (1961), Ayoade (1974), Nieuwolt (1977), Nicholson (1981) and Olaniran (1991) have discussed the association between seasonal and yearly rainfall variations in West Africa with the movements of the ITD. Typically, the ITD moves far enough North to cause a pronounced little dry season in the south-western portion of Nigeria, and somewhat lower rainfall amounts in the south-eastern region. During some years with below normal rainfall in the northern region, the ITD might not have penetrated far enough into the Sahelian zone, thereby causing above-normal rains in both southern regions. Tanaka *et al* (1975) analyzed August precipitation data in Africa and found that, in 1968, heavy flooding rainfall occurred along the normally dry southern coast while the Sahel experienced the driest August on record. Similarly, during some years of above-normal rainfall in the northern region, the ITD might have traveled further north than normal resulting in a prolonged little dry season in the southern region.

This space is for Table 4.1

This space is for figure 4.3

4. 2. Standardized precipitation anomalies in Nigeria

The standardized precipitation anomaly for Nigeria was also derived from the long-term mean for the period 1901-2000 and presented in figure 4. 4. It is interesting to note the dynamic nature of Nigerian climate over the last century. The pattern of rainfall variability since 1901 in the country shows that the 20^{th} century begins with relatively long period of wet conditions which are fairly persistent until about 1936 with occasional breaks of dry climatic condition. This is followed by a relatively dry climatic condition through the 1940s which lasted until about 1953. From 1953, the country experiences a relatively wet period which lasted until about 1963. From about 1964, droughts have been relatively persistent in Nigeria and particularly since 1969 reaching a peak in 1973. The country, however, witnessed an occasional break of normal or wet conditions between 1979-81 which was again followed by another long dry climatic condition from about 1982 and lasted until about 1994.

The observed country-wide occurrence of above average rainfall (wet condition) in the country, particularly during the first two decades of the 20^{th} century as well as in the 1950s, supports the hypothesis of considerable northward incursion of the Inter-Tropical Discontinuity (ITD) in wet years as reported by Ilesamni (1972), Nicholson (1983), Burnting *et a*l (1976) and Olaniran (1991a). However, the country-wide rainfall deficits experienced in the country from 1971-73 and 1983-87 supports the hypothesis of a restricted northward advance of the Inter-Tropical Discontinuity (ITD) in dry years as documented in the literature (see Winstanley, 1973a; Ayoade, 1974; Oguntoyinbo and Richards, 1977; Acheampong, 1987; Olaniran, 1988a, b; Olaniran, 1991); though Farmer and Wigley (1985), argued that it was difficult to determine the specific causes of drought on a regional scale in West Africa. Among the possible explanations for lack of moisture over the Sahel at the beginning of the rainy season, as given by Rasool (1984), are the displacement of ITD and changes in global east-to-west air circulation patterns caused, in part, by changing conditions over distant oceans (Indian, Atlantic and Pacific).

Oguntoyinbo and Richards (1977) presented results on ITD movements from 1969 to 1973 and showed that the northernmost limit reached by the Intertropical Discontinuity (ITD) progressively decreased at between 0.5 1^0 latitude per annum from 1969 to 1973 (i.e from 21.5^0 N to 19.5^0 N). They further noted that the rate of retreat was particularly rapid in 1971 and 1972 (7^0 of latitude between September and

October, 1972). They did not, however, report any corresponding southward displacement of the ITD at its southern limit during this period.

Similarly, an increase in the intensity of the subtropical high will also block the northwards movement of the Intertropical Discontinuity (ITD) (Stranz, 1978). Stranz (1978) presented results on mean deviations of air pressure to show that the anomalous rainfall of the period 1970-1972 over the Sahel is characterized by higher than normal pressure over North Africa where the sub-tropical highs are normally situated. The strong easterly winds that develop on the southern flank of the Saharan High picks up more dust into the atmosphere and thereby inhibiting precipitation processes, as well as block the northward movement of the ITD (see also Nieuwolt, 1977; Olaniran, 1989, 1991).

This space is for figure 4.4

Figures 4.5*a*-*h* show the pattern of the annual rainfall departure from the long-term mean during some selected wet and dry years in Nigeria. These results also support the pattern of annual rainfall anomalies in Nigeria shown in figure 4.4. A wet year is taken as a year when all the stations in the country recorded above average rainfall while a dry year is taken as a year when all the stations in the country recorded below average rainfall. The selected wet and dry years in this study were determined by dividing the country's total annual rainfall amount of 36619mm with the total number of stations considered in this study over the period 1951 to 2000 when all the stations had adequate rainfall record. The country's mean annual rainfall amount over the period 1951-2000 is 1356mm. A deviation from the long-term mean is then obtained from each year's value. Years with positive deviation represent wet years while those with negative deviations represent the dry years.

It could be observed from figures 4.5*a* and *b* that during the wet years 1954-1955, virtually the entire country experiences above average rainfall, which indicates a wetter climatic conditions during the period. In the wet year 1957, all the twenty-seven selected stations in the country have recorded total annual rainfall above their respective long-term means (see Fig. 4. 5*c*). It is also interesting to note that in the year 1962 (a wet year), only 3 stations have recorded an annual total rainfall below their respective long-term means. In other words, a total of 24 out of 27 selected stations have recorded a total annual rainfall above their respective long-term means which indicate a wetter climatic condition during the period (see Fig. 4.5*d*).

This space is for figure 4.5*a*

This space is for figure 4.5*b*

This space is for figure 4.5*c*

This space is for figure 4.5*d*

An examination of the pattern of annual rainfall departure from normal in Nigeria during the selected dry years of 1972-1973 and 1982-1983 (Figs. 4.5*e* - *h*) reveals that with the exception of few stations, the entire country experiences a pronounced rainfall deficits during those years which indicate a dry climatic condition during the period. In other words, most of the stations in the country have recorded a total annual rainfall below their respective long-term means. The problem becomes more glaring in the year 1983 (Fig. 4.5*h*) when all the twenty-seven selected stations in the country recorded a total annual rainfall far below their respective long-term means. Physical mechanisms for the occurrence and persistence of the anomalous wet and dry years in West Africa particularly during the 1950s, 1970s and 1980s have been suggested (see Grist and Nicholson, 2000; Long *et al*, 2000). Notable among them is the African Easterly Jet (AEJ) and associated shears coupled with restricted northward incursion of the ITD in dry years and its considerable northward advance in wet years. The African Easterly Jet (AEJ) is an important dynamic feature of the circulation over West Africa. It provides the instability needed to produce the African wave disturbances that characterized the rainy season (Burpee, 1972).

Most previous studies (e.g. Kidson, 1977; Kanamitsu and Krisnarmurti, 1978; Newell and Kidson, 1984; Fontaine and Janicot, 1992; Fontaine *et al*., 1995) showed a consistent association between the AEJ and rainfall, with the AEJ being anomalously strong (higher core speeds, stronger shear) in dry years and anomalously weak in wet years. It is generally assumed that the stronger jet is a factor in anomalously dry conditions.

However, a summary of the several changes in the general atmospheric circulation that have accompanied the shift to the drier conditions include: the African Easterly Jet (AEJ) is further south and more intense, the Inter-Tropical Discontinuity (ITD) is further south, the Tropical Easterly Jet (TEJ) is weaker, the equatorial westerlies are shallower and weaker, the southwesterly monsoon flow is weaker and relative humidity is lower (but not consistently so) (see among others Kidson, 1977; Newell and Kidson, 1984; Hastenrath, 1990; Janicot, 1992; Fontainne and Janicot, 1992; Miller and Lindzen, 1992; Fontaine *et al*., 1995; Goldenberg and Shapiro, 1996; Grist and Nicholson, 2000; Long *et al*., 2000; Grist, 2002; Paeth and Stuck, 2004; Okumura and Xie, 2004).

This space is for figure 4.5*e*

This space is for figure 4.5*f*

This space is for figure 4.5*g*

This space is for figure 4.5*h*

A comparison of a regional analysis of standardized precipitation anomalies in Nigeria (Fig. 4. 6) shows that the northern region of Nigeria experiences above average rainfall between 1905 to late 1910s which persisted until about 1936. The wet period of the 1950s is observed in virtually all the regions except the south-eastern region which experiences a pronounced drought throughout the 1950s. However, a relatively wet period is observed in the 1930s in all the regions except the central (Middle Belt) region of Nigeria. It could also be observed that the first and second decades (1901-10 and 1911-20) are dry decades in south-western region of Nigeria while the north, central and south-eastern regions of Nigeria experienced above average rainfall during the same period. Another feature observed is the country-wide occurrence of below average rainfall during the periods 1944-1948, 1968-76 and 1983-88 which also coincides with the droughts of 1940s, 1970s and 1980s, which almost ravaged the entire West African region as reported by (Winstanley, 1973b; Tanaka *et al*, 1975; Landsberg, 1975; Oguntoyinbo and Richards, 1977; Mortimore, 1989; Nicholson, 1989; Adejuwon, *et al*, 1990; Damaree and Nicolis,1990; Druyan, 1988; Hulme, 1992). The results presented in Fig. 4. 6 are consistent with the result presented in table 4.1.

It should be noted, however, that the large-scale rainfall deficits experienced in the country during these periods suggest the southward displacement of the Inter-Tropical Discontinuity (ITD) during the dry years coupled with weakened intensity of rain-bearing south-west monsoon due to sea surface temperature anomalies. Folland *et al* (1986) linked the observed rainfall anomalies in the Sahel to the extra-ordinary warming of the Atlantic, Pacific and Indian oceans respectively.

This space is for figure 4.6

This space is for figure 4.6

4. 3. Drought incidence based on the SPI index

The entire rainfall data for each of the twenty-seven selected stations are also subjected to Standardized Precipitation Index (SPI) technique in order to detect drought events, magnitude and severity. The SPI technique is developed by McKee et al (1993) for the purpose of defining and monitoring droughts. The technique has been adjudged a good indicator of moisture supply (Wu et al, 2005; Girux, 2001; Bordi et al, 2001). The SPI is based on statistical probability and is designed to be a spatially invariant indicator of drought. The nature of the SPI allows an analyst to determine the rarity of a drought or an anomalous wet event at a particular time scale for any location in the world that has precipitation records. This study adopts a 12-month SPI scale for drought analysis in Nigeria. All negative SPI values are taken to indicate the occurrence of drought, while all positive values show no drought. The temporal analysis of drought using the SPI index is carried out and results presented in table 4.2. Table 4. 2 shows the drought episodes station by station in Nigeria. It shows the number of mild to extreme droughts station by station across the country. It is interesting to note that all the stations have recorded high number of mild droughts on the SPI scale (0 to -0.99) compared to moderate droughts.

However, the striking feature of Nigerian climate in terms of rainfall variability is the country-wide occurrence of severe and extreme droughts across the country during the periods 1971-1973 and 1983-1987 as shown in figure 4. 4, and frequent occurrence of mild to moderate droughts as shown in table 4.2. The severe and extreme droughts are associated with extremely dry conditions and as such recovery can be very difficult. The resultant effects of drought of this magnitude are the loss of agricultural output, reduced biomass production and a great reduction in animal quality and quantity. These types of droughts are known to have caused outright migration and abandonment of farmlands (Appa, 1987; Oyebande, 1990). Indirectly, it may lead to change in land use practice and create heavy pressure on urban centers there by putting severe strain on the economy (Oguntoyinbo and Richard, 1977). It should however be noted that the occurrence of severe to extreme droughts is on the decrease when compared to the frequency of occurrence of mild to moderate droughts in the country. This supports the hypothesis of decreasing trend in the occurrence of drought of various intensities in Nigeria over the period of the study.

Table 4.2. Drought events station by station based on the SPI index

Station	Extreme drought	Severe Drought	Moderate drought	Mild Drought	Total Events
Bauchi	-	06	06	33	45
Jos	02	01	07	31	41
Kano	-	06	06	37	49
Katsina	01	03	11	23	38
Kaduna	-	03	09	23	35
Nguru	-	06	03	21	30
Maiduguri	01	05	12	22	40
Sokoto	02	02	09	29	42
Zaria	01	03	04	22	30
Zelwa	01	01	08	23	33
Yola	03	01	07	30	41
Makurdi	-	05	05	26	36
Lokoja	-	02	12	28	42
Ikom	01	01	01	07	10
Ilorin	01	04	12	24	41
Bida	-	02	08	25	35
Minna	03	01	09	36	49
Osogbo	02	01	03	26	32
Ondo	-	01	09	41	51
Ibadan	01	04	10	32	47
Lagos	01	02	12	30	45
Enugu	01	01	02	38	42
Portharcourt	-	06	07	36	49
Calabar	01	02	11	35	49
Warri	03	03	07	32	45
Benin	03	05	05	29	42
Gusau	-	-	06	24	30
TOTAL	28(03%)	77 (07%)	201 (19%)	763 (71%)	1069

4. 4. Drought frequency-duration analysis using SPI index

The frequency - duration analysis of drought in Nigeria is carried out using the McKee et al's (1993) SPI index and the result is presented in table 4.3. From table 4.3., it could be deduced that a total of 254 cases of sporadic (single) drought occurred in Nigeria with Bauchi and Lagos having the highest number of 15 cases each. Ikom records the least cases of such sporadic drought having only 2 cases. There are about 103 cases of a 2-year drought at a stretch in the country with Enugu recording the highest number of 10 cases and Yelwa and Gusau recorded the least number of only one case each. Table 4.3 also shows that there are 51 cases of a 3-year drought at a stretch in the country and Lagos recorded the highest number of 5 cases. Ondo records the highest number of 4 cases of a 3-year drought at a stretch. The result further shows that a total of 21 cases of a 4-year drought stretch are recorded in the country and Ondo recorded the highest number of 5 cases of a 4-year drought at a stretch. A 5-year drought at a stretch is not many in the country. There are only 6 of such cases in the country within the period of the study. However, a drought stretch of more than 5 years has occurred about 38 times with Calabar having the highest number of 4 cases of such droughts. A detailed description of the station by station analysis of drought-duration is given in the following section.

Table 4. 3. Drought duration as determined using the SPI index.

S/No.	Station	1-year Drought	2-year Drought	3-year Drought	4-year Drought	5-year Drought	Above 5-year Drought
1	Bauchi	15	07	01	01	-	01
2	Kano	11	06	02	01	-	02
3	Katsina	05	02	01	-	-	03
4	Sokoto	09	-	01	-	01	02
5	Maiduguri	12	03	01	-	01	02
6	Jos	09	04	01	01	-	01
7	Kaduna	12	04	01	-	-	01
8	Nguru	06	03	01	-	-	01
9	Zaria	07	04	-	01	-	01
10	Yelwa	08	01	02	03	-	01
11	Yola	08	07	03	01	-	01
12	Makurdi	07	04	02	-	-	02
13	Lokoja	12	04	01	01	-	02
14	Ilorin	11	04	04	01	-	01
15	Bida	09	04	04	-	-	01
16	Minna	12	04	-	-	-	02
17	Gusau	05	01	02	-	-	01
18	Osogbo	08	03	02	02	-	-
19	Ondo	11	02	03	04	01	01
20	Ibadan	12	06	04	01	02	-
21	Lagos	15	04	05	-	-	02
22	Warri	11	05	03	-	-	02
23	Benin	09	06	02	01	01	02
24	Enugu	11	10	01	02	-	-
25	P/Harcourt	08	02	01	02	-	02
26	Calabar	09	02	01	-	-	04
27	Ikom	02	01	02	-	-	-
Total		254	103	51	21	06	38

A drought duration analysis at individual stations is carried out using a 12-month SPI analysis. It reveals quite an interesting result. For example, from 1908-2000, Bauchi records 15 sporadic (single) drought events with 7 cases of a 2-year drought at a stretch. A 3-year drought at a stretch occurs only once in Bauchi. There is one episode of a 4-year drought at a stretch from the year 1925-1928. A 5-year drought stretch was lacking at Bauchi and, in place of it, was the longer duration droughts. The longest drought duration of ten years at a stretch occurs only once during the year 1982-1991 as shown in appendix 1.

At Kano, there are 11 cases of sporadic (single) drought events over the period 1905-2000 with a 2-year drought at a stretch which occurs 6 times. There are very few cases of a 3-year drought at a stretch in Kano. It occurs 2 times followed by a 4-year drought at a stretch which occurs only once from 1912-1915. There is no 5-year drought at a stretch in Kano. There is however a 6-year drought at a stretche from 1939-1944. It is noteworthy that the longest drought duration of seven years at a stretch occurs twice in Kano. These drought events occurred from 1971-1977 and 1981-1987 (see appendix 1). This incidentally coincides with the period of the 1970s and 1980s droughts which ravaged the West African Sahel (Winstanly, 1973a; Landsberg, 1975; Ojo, 1977; Lamb, 1982; Nicholson, 1986a, 1989a; Shukla, 1990).

Katsina experiences a sporadic (single) drought event 5 times from 1918-2000 with 2 cases of a 2-year drought at a stretch. A 3-year drought stretch occurs only once in Katsina from (1940-1942). There is no 5-year drought at a stretch in the area. However, an eight year drought at a stretch occurs twice in the area. This chain of drought events occurs from 1971-1978 and 1981-1988, and coincides with the period of the West African droughts of the 1970s and 1980s as documented in the literature (see Lamb, 1978, 1985; Gribbin and Lamb, 1979; Glantz, 1994). The longest drought duration of ten years at a stretch occurs only once in Katsina during the year 1990-1999 (see appendix 1).

Sokoto records 9 sporadic (single) drought events from 1916-2000 with a single case of a 3-year drought at a stretch. A 5-year drought at a stretch occurs from 1940-1944. The area also experiences a 9-year drought at a stretch from the year 1979-1987. However, the longest drought duration of ten years at a stretch occurs only once in Sokoto during the year 1966-1975 (see appendix 1). It is interesting to note that a 9-year and 10-year drought at a stretch experienced in Sokoto also coincides with the drought period of 1968-75 and 1982-87 as shown in table 4.1. At Maiduguri, a sporadic (single) drought event occurs 12 times over the period 1915-2000 followed by 3 cases of a 2-year drought at a stretch. A 3-year drought at a stretch occurs only once from 1971-1973. There is a 5-year drought at a stretch in Maiduguri between 1940-1944 followed by a 6-year drought at a stretch from 1989-1994. The longest drought duration of 7 years at a stretch occurs only once during the period 1981-1987 (See appendix 1).

Temporal analysis of drought at Jos using the SPI index also shows 9 cases of a sporadic (single) drought in the area with 4 cases of a 2-year drought at a stretch. Results further show a 3-year drought at a stretch from the year 1964-1966 followed by a 4-year drought at a stretch between 1977-2000. There is however no 5-year drought at a stretch in Jos during the period 1922-2000. It should be noted, however, that the longest drought duration of 17 years at a stretch occurs only once in Jos during the period 1979-1995 (see appendix 1). It is interesting to note that, out of the 17 years of drought during that period, it is only one year (1995) that shows an extreme drought of a magnitude -2.84 on the SPI scale. The remaining 16 years record droughts of lesser magnitudes that ranged from mild (0 to -0.99) to moderate (-1.00 to -1.49) droughts on the SPI scale. The water shortage associated with their occurrences is quite insignificant so that damages caused by this type of drought events do not create panic.

Kaduna records 12 cases of sporadic (single) drought and 4 cases of a 2-year drought at a stretch. It also experiences a 3-year drought at a stretch during the period 1947-1950. There is however no 5-year drought at a stretch in Kaduna. The longest drought duration of 8 years occurs once from 1983-1990 and coincides with the country-wide drought of the 1980s as revealed in table 4.1.

At Nguru, a 12-month SPI analysis shows that there are 6 cases of sporadic (single) drought events in the area with 3 cases of a 2-year drought at a stretch. There is single case of a 3-year drought stretch in the area from 1998-2000. However, the longest drought duration of fourteen years at a stretch occurs only once during the 1980-1993. It is noteworthy that 8 out of 14 years recorded only a mild drought (0 to -0.99) on the SPI scale. Such drought events have no serious consequences on rain-fed farming since the water shortage associated with them is insignificant, and the recovery rate from this type of drought event is quite fast if the precipitation situation appreciates upwards slightly.

Zaria records 7 cases of sporadic (single) drought events with 4 cases of a 2-year drought at a stretch. There is a single case of a 4-year drought at a stretch in Zaria during the period 1980-1984. The longest drought duration of 7 years at a stretch occurs once in Zaria during the period 1967-1973. At Yelwa, a 12-month SPI analysis shows 8 cases of sporadic (single) drought events. A 2-year drought at a stretch occurs only once in the area. A 3-year drought at a stretch occurs 2 times in Yelwa with 3 cases of a 4-year drought at a stretch. The longest drought duration of 6 years occurs only once in Yelwa during the period 1943-1948 (see appendix 1).

A 12-month SPI analysis for Yola shows that the area experiences 8 cases of sporadic (single) drought events and 7 cases of a 2-year drought at a stretch. A 3-year drought at a stretch occurs 3 times followed by a 4-year drought at a stretch during the period 1930-1933. A 6-year drought at a stretch is the longest drought duration in Yola which occurs from 1982-1987 (see appendix 1). Makurdi experiences 7 sporadic droughts and 4 cases of a 2-year drought at a stretch. A 3-year drought at a stretch occurs twice in Makurdi followed by a 6-year drought at a stretch during the period 1976-1983. The longest drought duration of 7 years at a stretch occurs only once at Makurdi during the period 1985-1991 (see appendix 1).

A 12-month SPI analysis at Lokoja shows that the area experiences 12 cases of sporadic (single) drought events and 4 cases of a 2-year drought at a stretch. It further reveals that the area experiences a single case of a 3-year, 4-year and 7-year drought at a stretch during the period 1950-52, 1935-38 and 1979-85 respectively. However, the longest drought duration of eight years at a stretch occurs once in Lokoja during the period 1940-1947 (see appendix 1). It should be noted, however, that the 8-year drought experienced at Lokoja coincides with the period of the 1940s drought.

Ilorin experiences 11 cases of sporadic (single) drought events . A 2-year and 3-year drought at a stretch occurs 4 times each which is then followed by a 4-year drought at a stretch between 1987-1990. However, the longest drought duration of 6 years at a stretch occurs during the period 1975-1980 (see Fig. 4.7 and appendix 1). At Bida, the result shows that the area experiences 9 cases of sporadic (single) drought events and 4 cases of a 2-year and 3-year drought at a stretch. The longest drought duration of 6 years at a stretch occurs during the period 1985-1990 (see appendix 1).

Minna records 12 cases of a sporadic (single) drought event and 4 cases of a 2-year drought at a stretch. Minna also experiences a 6-year drought at a stretch during the period 1995-2000. The longest drought duration of 11 years at a stretch occurs from 1980-1990, with droughts of various magnitudes. Gusau experiences 5 cases of sporadic (single) drought events and a 2-year drought at a stretch during the period 1964-65. This is followed by 2 cases of a 3-year drought at a stretch. The result further shows that Gusau experiences a single case of 8-year drought at a stretch during the period 1980-1987. However, the longest drought duration of 9 years at a stretch occurs between 1969-1977 (see appendix 1).

Osogbo, a southern station, records 8 cases of sporadic (single) drought events and 3 cases of a 2-year drought at a stretch. It, however, experiences 2 cases of a 3-year drought at a stretch and 2 cases of a 4-year drought at a stretch (see appendix 1). There is no drought longer than four years in the area. At Ondo, the result shows the area experiences 11 cases of sporadic (single) droughts and 2 cases of a 2-year drought at a stretch. The area also experiences 3 cases of 3-year drought at a stretch followed by 4 cases of a 4-year drought at a stretch. A 5-year drought at a stretch occurs from 1942-1946. The longest drought duration of 7 years at a stretch occurs during the period 1923-1929 (see appendix 1).

Ibadan records 12 cases of a sporadic (single) drought event and 6 cases of a 2-year drought at a stretch.The result further shows that the area experiences 4 cases of a 3-year drought at a stretch. There is however a single case of a 4-year drought at a stretch in Ibadan during the period 1974-1977. Ibadan also records 2 cases of a 5-year drought at a stretch (see appendix 1). There is however no drought of more than five years at Ibadan over the period investigated in this study. Lagos records 15 cases of a sporadic (single) drought events and 4 cases of a 2-year drought at a stretch. There are also 5 cases of a 3-year drought stretch in Lagos followed by a 6-year drought at a stretch between 1981-1986. The longest drought duration of eight years occurs during the period 1971-1978 (see appendix 1). This period coincides with the period of the country-wide rainfall deficits of the 1970s as shown in figure. 4. 4.

Warri records 11 cases of a sporadic (single) drought events and 5 cases of a 2-year drought at a stretch. This is followed by 3 cases of a 3-year drought at a stretch in the area. The result futher shows that Warri experiences a single case of a 7-year drought at a stretch during the period 1932-1938. The longest drought duration of 8 years at a stretch occurs only once in Warri from 1919-1926 (see appendix 1).

Benin records 9 cases of a sporadic (single) drought and 6 cases of a 2-year drought at a stretch. This is followed by a 3-year drought at a stretch during the period 1932-34. The result further shows that Benin experiences a single case of a 3-year, 4-year, 5-year and 6-year drought at a stretch during the period 1932-34, 1907-10, 1944-48 and 1981-86 respectively. The longest drought duration of 7 years at a stretch occurs only once during the period 1922-1928 (see appendix 1). Enugu records 11 cases of a sporadic (single) drought events and 10 cases of 2-year drought at a stretch in the area. A 3-year drought at a stretch occurs only once in the area during the year 1992-94. The longest drought duration of 4 years at a stretch occurs twice during the period 1981-84 and 1986-89 (see appendix 1).

PortHarcourt records 8 cases of a sporadic (single) drought events and 2 cases of a 2-year drought at a stretch in the area. A 3-year drought at a stretch occurs once in PortHarcourt during the year 1963-65. The area also experiences 2 cases of a 4-year drought at a stretch. The result further shows that the area experiences 13-year drought at a stretch during the period 1988-2000. It should be noted, however, that it is only 4 years (1989, 1991, 1997 and 2000) out of 13 years that recorded a moderate drought (-1.00 to -1.49) on the SPI scale, the remaining 9 years recorded only a mild drought (0 to -0.99) on the SPI scale. However, the longest drought duration of eighteen years at a stretch occurs only once during the period 1969-1986 (see appendix 1). It is interesting to note that it is only 3 years (1971, 1973 and 1983) that had recorded a moderate to severe drought on the SPI scale. The remaining 15 years of drought recorded only a mild drought (0 to -0.99) on the SPI scale.

Calabar experiences 9 cases of a sporadic (single) drought events and 2 cases of a 2-year drought at a stretch. This is followed by a 3-year drought at a stretch during the period 1998-2000. The area also experiences 2 cases of a 6-year drought at a stretch during the period 1970-75 and 1981-86. A 7-year drought at a stretch occurs in the area during the period 1988-1994. The longest drought duration of 10 years at a stretch occurs only once in the area during the year 1944-1953 (see appendix 1). It is noteworthy that PortHarcourt and Calabar had each recorded a total of 49 drought events from 1902-2000. However, while PortHarcourt had 36 out 49 drought events as a mere mild droughts, Calabar had 35 out 49 drought events as a mere mild droughts. Ikom records 2 cases of a sporadic (single) drought events and a 2-year drought at a stretch during the period 1988-1989. A 3-year drought at a stretch occurs twice at Ikom, with no further droughts over the period 1972-2000.

4. 5. Drought severity and magnitude based on SPI index

Drought severity/magnitude is determined using SPI which indicates the degree of departure from annual precipitation mean. The degree of departure is then read against an SPI prepared index (McKee *et al*, 1993). The results of a drought severity (magnitude) for selected Nigerian stations are presented in table 4. 4. It is interesting to note that all the stations showed a high percentage frequency of occurrence of a mild drought (0 to -0.99) on the SPI scale (see Table 4. 4). This is followed by a moderate drought (SPI scale -1.00 to -1.49). It should be noted, however, that the high percentage frequency of occurrence of a mild drought at all the stations studied do not pose any serious threat to rain-fed farming in the sense that the water shortage associated with this type of drought is quite insignificant. The recovery rate from this type of drought is quite fast if precipitation situation appreciates upwards slightly. However, table 4.4 also shows that all the stations in Nigeria recorded few cases of severe to extreme droughts over the period of the study when compared to the frequency of occurrence of mild to moderate droughts in the country.

Table 4. 4. Percentage frequency of occurrence of drought of various intensities based on the SPI index

Station	Length of record (years)	Drought events	Frequency Of occurrence (%)			
			Mild Drought	Moderate Drought	Severe Drought	Extreme Drought
Bauchi	89	45	74	13	13	00
Jos	79	41	76	17	02	05
Kano	95	49	76	12	12	00
Katsina	76	38	61	29	08	02
Kaduna	63	35	66	26	08	00
Nguru	59	30	70	10	20	00
Maiduguri	85	40	55	30	13	02
Sokoto	83	42	60	21	05	02
Zaria	58	30	73	13	10	04
Yelwa	62	33	70	24	03	03
Yola	86	41	73	17	03	07
Makurdi	72	36	72	14	14	00
Lokoja	83	42	67	29	04	00
Ikom	29	10	70	10	10	10
Ilorin	85	41	59	29	10	02
Bida	73	35	71	23	06	00
Minna	85	49	74	18	02	06
Osogbo	57	32	82	09	03	06
Ondo	95	51	80	18	02	00
Ibadan	96	47	68	21	09	02
Lagos	99	45	67	27	04	02
Enugu	81	42	91	05	02	02
Portharcourt	96	49	74	14	12	00
Calabar	97	49	72	22	04	02
Warri	93	45	71	15	07	07
Benin	95	40	69	12	12	07
Gusau	48	30	80	20	00	00

A station by station analysis of a 12-month SPI shows that the highest drought magnitude at Bauchi measures -1.98 (severe drought). This occured in 1985 (see Fig. 4.7 and appendix 1) and because of the extremely dry condition associated with it, recovery could be very difficult. It is believed that this 1985 severe drought set a chain of drought events of different magnitudes over the period 1985-1991. At Kano, the highest drought magnitude measures -1.98 (severe drought). This occured in 1973, the peak year of the 1968-73 droughts in West Africa.

The highest drought magnitude at Katsina measures -2.18 (extreme drought) on the SPI scale (see Fig. 4.7 and appendix 1) and occurs in 1993. At Sokoto the highest drought magnitude measures -2.11 (extreme drought) on the SPI scale. This occured in 1987 (see Fig. 4.7 and appendix 1). The highest drought magnitude in Maiduguri measures -2.34 (extreme drought). This occured in 1983 and because of the extremely dry condition associated with it, recovery could be very difficult. This extremely dry drought is believed to set a chain of drought events of different magnitudes covering 1983-1994, with the exception of 1998 (see Fig. 4.7 and appendix 1). The highest drought magnitude at Jos measures -2.84 (extreme drought) (see Fig. 4.7 and appendix 1) and occured in 1995. Kaduna has its highest drought magnitude measures -1.87 (severe drought) on the SPI scale. This occured in 1949 and coincides with the period of the 1940s drought in the country.

The highest drought magnitude at Nguru measures -1.72 (severe drought) This occured in 1983. At Zaria, the highest drought magnitude measures -2.18 (extreme drought) on the SPI scale and occurs in 1983. This coincides with the period of the country-wide rainfall deficits of the 1980s as shown in figure 4.4. Yelwa records the highest drought magnitude of -2.20 (extreme drought) in 1983. At Yola, the highest drought magnitude measures -3.17 (extreme drought). This coincides with the onset year of the 1968-73 droughts in West Africa (see Druyan, 1989; Adejuwon et al, 1990; Damaree and Nicholis, 1990; El Tahir and Gong, 1996). Markudi records its highest drought magnitude of -1.91 in 1988 while Lokoja records -1.86 as its highest drought magnitude (see Fig. 4. 7 and appendix 1).

Similarly, Ilorin records its highest drought magnitude of -2.01 (extreme drought) in 1989. At Bida, the highest drought magnitude measures -1.74 (severe drought) in 1972. Minna records its highest drought magnitude of -2.24 (extreme drought) in 1987 while Gusau records its highest drought magnitude of -1.41 (moderate drought) in 1984. It is noteworthy that Gusau, a northern station, has not experienced either a severe or extreme drought over the period 1953-2000. At Osogbo, the highest drought magnitude determined was -2.13 (extreme drought). This occured in 1977 (see Fig. 4. 7 and appendix 1). Ondo records its highest drought magnitude of -1.57 (severe drought) in 1920 while Ibadan records its highest drought magnitude of -2.07 (extreme drought) in 1912 (see Fig 4. 7 and appendix 1). The result further shows that Lagos records its highest drought of -2.59 magnitude (extreme drought) in 1982. At Warri, the highest drought magnitude determined was -2.25 (extreme drought) and occurred in 1977. In a similar vein, Benin recorded a highest drought magnitude of -2.33 (extreme drought) (see Fig. 4. 7 and appendix 1).

In addition, Ikom, a southestern station, records a highest drought magnitude of -3.48 (extreme drought). This occurs in 1972 (see Fig. 4. 7 and appendix 1). It is interesting to note that the 1972/1973 was the peak year of the 1968-73 droughts which ravaged the West Africal Sahel and caused untold hardship to the economy of the region as demonstrated by Wistanley (1973a), Tanaka *et al* (1975), Nicholson (1981), Mortimore (1989) and Hulme (1992). The highest drought magnitude of -2.16 (extreme drought) occurs at Enugu in 1983. PortHarcourt had its highest drought of magnitude -1.78 (severe drought) which ocurred in 1958. This indicates that PortHarcourt does not experience any extreme drought on record from 1902-2000.

The highest drought magnitude at Calabar measures -1.92 (severe drought) (see Fig. 4.7 and appendix 1). This shows that Calabar does not experience an extreme drought on record from1902-2000. The results of drought severity/magnitude analysis using the SPI index suggest that the large-scale droughts only rarely cover the country as a whole since there are distinct spatial differences that dominate the wet and dry years. It is noteworthy that the length and severity of drought in Nigeria vary from one area to another. This supports the hypothesis that drought occurrence in Nigeria is largely sporadic in its spatial patterns. It should however be noted that the spatial patterns of droughts in Nigeria also suggest that the seasonal incursion of the Inter-Tropical Discontinuity (ITD) may not likely be the leading cause of drought in West Africa, Nigeria inclusive, as observed by Lamb (1975), Kraus (1977a, b) and Greenhut (1977). More recently, researchers have found more plausible explanations for the large-scale droughts that ravaged the West African Sahel in the 1970s and 1980s by extending their analysis to additional factors that control the rainfall intensity such as the contrasting sea surface temperatures (SSTs) over the tropical Atlantic and Pacific oceans (Parker *et al*, 1987; Semezzi *et al* 1988; Nicholson, 2001).

This space is for figure 4.7

This space is for figure 4.7

This space is for figure 4.7

This space is for figure 4.7

4. 6. Drought analysis using the PDBM index

The drought episodes in Nigeria are also quantified using the PDBM index which also utilizes only the annual rainfall amounts of the individual stations. The results are then compared with that of the SPI index. The drought events detected at individual stations using the PDBM index are presented in table 4.5. The results are then compared with that of table 4. 2. It could be observed that the SPI index records more droughts in the country than the ones obtained using the PDBM index. For example, about three cases of extreme droughts were recorded each for Minna, Warri and Benin based on the SPI index with no such droughts recorded using the PDBM index. Similarly, Yola records three extreme droughts based on the SPI index but recorded only one diasatrous drought when analyzed using the PDBM index. Another noticeable feature of table 4.5 is the high percentage frequency of occurence of drought events in Nigeria. It could be observed that both the southern and northern stations had almost equal number of drought events based on the SPI index which is unrealistic (see Table 4.5). However, there are few drought cases in most of the southern stations when analyzed using the PDBM index. For example, Enugu, Lagos and Ikom had only one severe drought each whereas Nguru, Katsina and Maiduguri had six, three and two severe droughts respectively (see Table 4.5).

It should be noted, however, that the apparent high frequency of occurrence of droughts of various categories based on the SPI index arises from the standardized nature of the index itself; namely that extreme droughts (or any other drought threshold) measured by the SPI, when considered over a long time period, will occur with the same frequency at all locations. Thus, the SPI is not capable of identifying regions that may be more 'drought prone' than others. For example, areas like Warri and Calabar which have humid climate, a mere negative deviation of annual rainfall at a given year from the long-term mean would not have a pronounced effect on crop water requirements in that area when compared to such deviations at the stations such as Nguru, Katsina, Sokoto and Maiduguri located in the sahelian region of Nigeria. Unlike the PDBM index, which considers only more than 10% deviation below the mean as drought, the SPI index considers all negative deviations from the long-term mean as drought. This is unrealistic especially when we carefully examined table 4.5. For example, Calabar, with 97 years record, records a total of 49 droughts, while Minna, with 85 years of record, records a total of 49 droughts. It could also be noted that Ondo, with 95 years of record, records a total of 51 droughts based on the SPI index. The implication of this result is that on the average, drought occurs after every one year at these stations. This made the SPI index unrealistic when applied to the Nigerian rainfall series, even though, it has been adjudged the best drought index in recent times (McKee *et al*., 1995; Guttman, 1999; Agnew, 1999; Keyantash and Dracup, 2002). However, a carefull examination of table 4.5 shows that Calabar records 21 droughts when analyzed using the PDBM index as against 49 droughts detected using the SPI index while Ondo records 33 droughts as opposed to 51 droughts detected using the SPI index. Hayes *et al* (1999) cite the inability of the SPI to identify those regions that are more drought prone as a potential disadvantage. However, standardization in space and time is precisely what is required from a standard classification scheme where the frequency of a given event should be independent of location. Thus, the SPI should be viewed as superior to the PDBM index. The last two columns of table 4.5 represent numbers of years with more than 10% deviation below the long-term mean and years with more than 10% deviation above the long-term mean respectively (hereafter referred to as total drought years and total wet years repectively). This was represented in figures 4.8 and 4.9 respectively.

Table 4.5. Drought events station by station based on the PDBM index

STATION	Length of Record	Disastrous Drought	Severe Drought	Moderate Drought	Slight Drought	Total Drought Years	Total Wet Years
Bauchi	1906 - 2000	-	-	04	21	25	17
Jos	1922 - 2000	-	-	02	16	18	14
Kano	1905 – 2000	-	01	11	21	33	25
Katsina	1918 – 2000	-	03	14	11	28	22
Kaduna	1930 – 2000	-	-	02	13	15	13
Nguru	1942 – 2000	-	06	10	06	22	23
Maiduguri	1915 – 2000	-	02	15	11	28	32
Sokoto	1916 – 2000	-	-	11	16	27	27
Zaria	1943 – 2000	-	-	02	12	14	14
Yelwa	1926 – 2000	-	-	04	15	19	14
Yola	1914 – 2000	01	01	04	18	24	22
Makurdi	1927 – 2000	-	-	07	11	18	13
Lokoja	1916 – 2000	-	-	06	22	28	21
Ikom	1972 – 2000	-	01	-	04	05	06
Ilorin	1916 – 2000	-	-	07	18	25	23
Bida	1928 – 2000	-	-	02	17	19	14
Minna	1916 – 2000	-	-	04	15	19	18
Osogbo	1935 – 2000	-	-	03	13	16	13
Ondo	1906 – 2000	-	-	07	26	33	23
Ibadan	1905 – 2000	-	-	08	20	28	25
Lagos	1901 – 2000	-	01	08	21	30	34
Enugu	1916 – 2000	-	01	03	21	25	18
Porharcourt	1902 – 2000	-	-	10	22	32	27
Calabar	1902 – 2000	-	-	02	19	21	22
Warri	1908 – 2000	-	-	01	15	16	18
Benin	1906 – 2000	-	-	08	17	25	22
Gusau	1953 – 2000	-	-	01	15	16	12
TOTAL		01 (0.2%)	16(3.0%)	156(25.6%)	436 (71.9%)	609	537

The spatial pattern of the number of years with more than 10% deviation below the mean (i.e. total drought years station by station) is presented in figure 4. 8. It is interesting to note that the total drought years increases with increasing distance from the coast. This suggests that there are more dry years especially in the extreme northern part of the country than in the more humid zone of the south. This implies that droughts are more prevalent in the semi-arid northern part of Nigeria than in the more humid southern region of Nigeria. This result conforms to the findings of other previous studies which suggest that the semi-arid northern Nigeria especially at the stations on fringes of the Sahel are more drought- proned (Kowal and Knabe, 1973; Oguntoyinbo and Richards, 1977; Mortimore, 1989).

It is interesting to note that Ikom, a southern station, has only 5 dry years from 1972-2000. Other areas with the least number of dry years in relation to their length of records include Zaria (14 years), Osogbo and Warri (16years) each. Some of the areas with the highest number of drought years include Kano (33 years), Katsina (28years) and Sokoto (27years). Several mechanisms have been speculated on the likely causes of droughts in sub-Saharan Africa amongst which include southward displacement of the ITD (Landsberg, 1973; Nicholson, 1984), changing land surface conditions (Charney, 1975; Schukla *et al*., 1995), El-Nino/Southern Oscillation (ENSO) (Popelewski and Halpert, 1987, 1989; Nicholson and Jeeyoung, 1997) and sea surface temperature anomalies (Folland *et al*., 1986; Palmer, 1986; Janikot, 1992).

This space is for figure 4.8

The spatial pattern of the number of years with more than 10% deviation above the mean (i.e. total wet years) is presented in figure 4.9. It could be observed from figure 4.9 that the number of wet years decreases with increasing distance from the coast. This suggests that there are more wet years in the southern part of the country than in the semi-arid northern part of the country. Several reasons may be given for the pattern presented in figure 4.9. Notable among them is the fact that the southern stations are more closed to the source of moisture (Atlantic Ocean) (proximity effect) and tend to be influenced by the south-west monsoon than the far northern stations. The strength of the south-west monsoon also declines with increasing distance from the coast. The north-south movement of the ITD could also be an additional factor in explaining the spatial pattern of total wet years in Nigeria (Ayoade, 1974; Olaniran, 1991).

This space is for figure 4.9

4.7. Drought frequency and magnitude analysis using the PDBM index

The drought severity/magnitude in relation to length of record is computed for each station using the PDBM index and the result presented in table 4.6. It could be observed from table 4.6 that slight droughts have the highest percentage frequency of occurrence in Nigeria followed by the moderate and severe droughts. The disastrous drought occured once in Nigeria over the period of the study. It occured at Yola over the period 1914-2000. The percentage frequency of occurrence of slight, moderate and severe droughts are also presented in table 4.6.

Table 4.6. Percentage frequency of occurrence of drought of various intensities based on the PDBM index

Station	Length of record (years)	Drought events	Frequency Of Occurrence (%)			
			Slight Drought	Moderate Drought	Severe Drought	Disastrous Drought
Bauchi	89	25	84	16	00	00
Jos	79	18	89	11	00	00
Kano	95	33	64	33	03	00
Katsina	76	28	39	50	11	00
Kaduna	63	15	87	13	00	00
Nguru	59	22	27	45	28	00
Maiduguri	85	28	39	54	07	00
Sokoto	83	27	59	41	00	00
Zaria	58	14	86	14	00	00
Yelwa	62	19	79	21	00	00
Yola	86	24	75	17	04	04
Makurdi	72	18	61	39	00	00
Lokoja	83	28	79	21	00	00
Ikom	29	05	80	00	20	00
Ilorin	85	25	72	28	00	00
Bida	73	19	89	11	00	00
Minna	85	19	79	21	00	00
Osogbo	57	16	81	19	00	00
Ondo	95	33	79	21	00	00
Ibadan	96	28	71	29	00	00
Lagos	99	30	70	27	03	00
Enugu	81	25	84	12	04	00
Portharcourt	96	32	69	31	00	00
Calabar	97	21	90	10	00	00
Warri	93	16	94	06	00	00
Benin	95	25	68	32	00	00
Gusau	48	16	94	06	00	00

Table 4.6 shows the percentage frequency of occurrence of slight droughts in Nigeria in relation to length of records, and its spatial pattern is presented in figure 4.10. It could be observed from table 4.6 that the percentage frequency of slight droughts is generally higher in the country. This suggests that slight drought tends to occur more frequently than any other type of drought based on the analysis carried out using the PDBM index. It should, however, be noted that this type of drought has no any pronounced effect on crops since it is associated with insignificant moisture deficits (Oguntoyinbo and Richards, 1977). It could be observed that Gusau and Warri have the highest percentage frequency of 94% each. This is followed by Calabar and Bida having a percentage frequency of 90% and 89% respectively.

It could also be observed from figure 4.10 that the percentage frequency of occurrence of slight drought in the country decreases with increasing distant from the coast. This implies that the southern stations have the highest percentage frequency of such droughts and by implications they tend to have more slight droughts than their counterparts in the semi-arid northern part of Nigeria. Stations with the least percentage frequency of occurrence of slight drought include Nguru (27%), Maiduguri and Katsina (39%) each and Sokoto (51%).

This space is for figure 4.10

Table 4.6 and figure 4.11 show the percentage frequency of occurrence of moderate drought in relation to length of record at selected Nigerian stations. Unlike the slight droughts, the percentage frequency of occurrence of moderate droughts in Nigeria shows an increasing trend northwards. In other words, the stations located in northern part of the country tend to have high percentage frequency of occurrence of moderate drought than their southern counterparts. This shows that there are more moderate droughts in the northern part of the country than the south. Some of the areas with the highest percentage frequency of occurrence of moderate droughts include Maiduguri (54%), Katsina (50%), Nguru (45%) and Sokoto (41%). Areas with the least percentage frequency include Warri (6%), Calabar (10%) and Enugu (12%). This type of drought could posed a serious threat to rain-fed farming if persisted over time.

This space is for figure 4.11

Table 4. 6 and figure 4.12 show the spatial pattern of the percentage frequency of occurrence of severe droughts at selected Nigerian meteorological stations in relation to lengths of record. It is interesting to note that the semi-arid northern part of the country is more prone to drought than the humid southern part of the country. Figure 4.12 shows an increasing trend northward of the percentage frequency of occurrence of severe droughts in Nigeria. In other words, severe droughts are more pronounced in the semi-arid northern part of the country than in the southern part of the country. Table 4.6 shows that Nguru has the highest percentage frequency of occurrence of severe drought (28%) while Lagos has the least (3%). It should however be noted that though the occurrence of severe drought is not so frequent like moderate and slight droughts, such droughts are associated with extremely dry conditions and the recovery could be difficult. In the years 1972-73, droughts of catastrophic proportions persistently ravaged the West African Sahel, creating serious food shortages for human and animal population and dealing a crippling blow to the fragile economies of the affected countries. Such severe droughts have seriously impaired the rural economy of the affected areas because of the resulting famine, starvation, diseases and rural migration. Drought has often aggravated the 'hydra-headed' crisis already confronting policy makers in Africa (Glantz and Katz, 1987; Mortimore, 1973, 1989).

It could be observed from figure 4.12 that the large-scale droughts only rarely cover the country as a whole and there are distinct spatial differences that dominate wet and dry years. The length and severity of droughts also vary from sub-area to sub-area. However, the semi-arid northern part of the country tends to catch the brunt of drought on a year to year basis than the humid southern part of the country. It is interesting therefore to note that drought occurrence in Nigeria is largely sporadic in its spatial pattern. It has been stressed that the interannual variability of the drought areas in the savanna region of Nigeria was large and demonstrated large variations in the seasonal rainfall over the region (Oladipo, 1995). Statistical evidence suggests a significant long-term increasing trend in the areal extent of drought in savanna region of Nigeria, with a major shift towards an increase in the mean areal extent of drought between two climatic periods 1931-1960 and 1961-1987 (Oladipo, 1995).

This space is for figure 4.12

4.8. Spatial variations in drought occurrence for some years

Using the PDBM index, the nature and pattern of the interannual areal variations in drought occurrence for some years in Nigeria are examined for a period 1970-1987 and presented in figure 4.13. However, for the purpose of discussion, years with less than 10% deviation below the mean are considered as 'wet to normal condition' while years with 11-25%, 26-45%, 46-60% and above 60% are considered as slight, moderate, severe and disastrous drougthts respectively based on the PDBM index classification. Then, the information on the pattern of dryness and wetness conditions over the country for years 1970-1975 and 1982-1987 is presented in figure 4.13. It is interesting to note that annual dryness/wetness conditions are highly irregular and in no single year that the country is consistently affected by a particular drought throughout the period of the study. The areal coverage of individual drought changes considerably in both size and portion from year to year. Oladipo (1993) shows that there are remarkable seasonal changes in the pattern of wetness and dryness over northern Nigeria with no consistent recurrent spatial patterns in moisture anomalies.

A visual examination of the spatial variations of severe droughts in Nigeria over the period of the study (see Fig. 4.13) showed that they are highly localized in all years. This indicates that the PDBM drought intensity index of 46-60% deviation from the mean occurs very infrequently and for only few stations during the period of the study. This result is in agreement with Oladipo's (1993) result using the Bhalme and Mooley drought index on monthly growing season (April-October) rainfall totals for 34 stations in northern Nigeria. He shows that large-scale droughts only rarely cover the region as a whole, and there are distinct spatial differences dominating the wet and dry years. From figure 4.13, it could be deduced that the areal coverage of slight drought is considerably large in the years 1970-73 with localized moderate droughts in the years 1971-73. However, a minor portions in the semi-arid northern part of the country and part of South-Western Nigeria are being affected by severe drought between 1971-73 (see Fig. 4.13).

The drought hard-hit years are 1983 and 1987. In 1983, the areal coverage of both moderate and severe droughts increases significantly more than the previous years (1971-73). It could be observed that the moderate drought covers almost 65% of the entire country in 1983 while the extreme northeast and part of southwest are adversely affected by severe drought. It is interesting to note that parts of southwest are also seriously affected by drought ranging from moderate to severe droughts in 1982.

In 1987, the areal coverage of moderate drought also increases and covers part of northwestern Nigeria and farther northeast. A relatively large area in both northwest and northeastern parts of Nigeria is also being affected by severe drought in the year 1987. The irregularity in the pattern of wetness and dryness in the country as presented in figure 4.13 could be linked to irregularity in the movement pattern of the ITD and to some extent changes in land surface characteristics as demonstrated by Ayoade (1974), Charney (1975) and Oladipo (1993).

This space is for figure 4.13

This space is for figure 4.13

4. 9. Rainfall probability assessment over Nigeria

The probability assessment of a critical annual rainfall amount can be useful for evaluating the risks for drought and flood in a wet-dry climatic environment of Nigeria. Using the annual rainfall data of 27 Nigerian meteorological stations over the period 1901-2000, estimates were made of the probability of receiving less than the least annual rainfall ever recorded at individual stations in the country via the normal frequency distribution function. The probability of receiving more than the highest annual rainfall ever recorded at individual stations in Nigeria is also computed. The rainfall probability maps for Nigeria are plotted and the implications of the results for drought and flood risks are highlighted in this section.

In most cases, floods are associated with either abnormally high daily rainfall events or annual rainfall events. Droughts are often associated with a sustained period of abnormally low annual rainfall events. The flood and drought risks can be evaluated using information on probability of receiving less than the least annual rainfall ever recorded and of receiving more than the highest annual rainfall ever recorded over a period of time in a given country. Flood is defined as inundation of the flood plain when the river overtaxes its banks due to over flow of water in the river (Vente Chow, 1978). There is no universally acceptable definition of drought in the literature. It is, however, acknowledged that drought is a meteorological phenomenon which is associated with a sustained period of significantly below normal precipitation (see Oladipo, 1986; Glantz, 1994).

Ayoade (1976) examines the magnitude, frequency and distribution of intense rainfall in Nigeria using maximum daily rainfalls of 46 Nigeria stations with varying duration ranging from 35 to 59 years. He computes the recurrence interval (return period) of the highest daily rainfall ever recorded at each of 46 stations using the Gumbel's Extreme Value theory and shows that the return period of the extreme daily rainfall decreases from the south towards the northern part of the country. Using the Gumbel's Extreme Value theory, Ologunorisa and Tersoo (2006) also estimate the return period of extreme daily rainfall at Makurdi, Nigeria between 1979-2004 and show that the period between 1996 and 2001 records the highest frequencies of extreme rainfall events; that major floods are associated with higher return periods. Despite the socio-economic implications of floods and droughts, no attempt is made to assess the risk of occurrence of these weather extremes in terms of their probabilities. Estimates are made in this section of the probability of receiving less than the least annual rainfall ever recorded at individual stations in Nigeria as well as the probability of receiving more than the highest annual rainfall ever recorded at individual stations in Nigeria.

Table 4.7 provides the percentage probability of receiving less than the least annual rainfall ever recorded at selected meteorological stations in the country while figure 4.14 presents the spatial pattern of the probability of receiving less than the least annual rainfall ever recorded at some selected stations in Nigeria. From table 4.7 and figure 4.14, it could be observed that Kano and Gusau stations, all located in the north, have the highest probability (12.30%) and (7.93%) for Kano and Gusau respectively, of receiving less than the least annual rainfall ever recorded at those stations over their respective lengths of records. This is followed by Nguru (4.27%) and Zaria (4.09%). This implies that Kano, Gusau, Nguru and Zaria stations stand the risk of drought than any other station in the country. It is interesting to note that Ondo, a southern station, also records a relatively high probability (5.82%) of receiving less than the least annual rainfall ever recorded over the period 1906-2000. Fig. 4.14 shows that the semi-arid northern part of Nigeria stands the risk of drought than humid southern part of Nigeria.

Table 4. 7. Probability of receiving less than the least annual rainfall ever recorded at individual stations in Nigeria, 1901-2000.

Station	Length of record	Mean (mm)	Standard deviation (mm)	*LARE (mm)	Year of occurrence	Z-Score	Probability A (%)	Probability B (%)
Bauchi	1908-2000	1066	172	726	1985	-1.97	97.56	2.44
Jos	1922-2000	1332	175	928	1961	-2.31	99.95	0.05
Kano	1905-2000	845	369	416	1973	-1.16	87.70	12.30
Katsina	1918-2000	641	174	262	1993	-2.18	98.53	1.47
Kaduna	1930-2000	1246	184	893	1949	-1.92	92.26	2.74
Nguru	1942-2000	475	144	227	1983	-1.72	95.73	4.27
Maiduguri	1915-2000	622	150	264	1983	-2.39	99.15	0.85
Sokoto	1916-2000	705	149	373	1987	-2.23	98.71	1.29
Zaria	1943-2000	1066	161	786	1989	-1.74	95.91	4.09
Yelwa	1926-2000	1002	190	584	1983	-2.20	98.61	1.39
Yola	1914-2000	926	186	337	1968	-3.17	99.90	0.10
Makurdi	1927-2000	1308	239	581	1988	-3.04	99.86	0.14
Lokoja	1916-2000	1191	233	758	1942	-1.86	96.86	3.14
Ikom	1972-2000	2237	325	1104	1972	-3.49	99.96	0.04
Ilorin	1916-2000	1253	228	795	1989	-2.01	97.77	2.23
Bida	1928-2000	1206	216	831	1972	-1.76	96.08	3.92
Minna	1916-2000	1284	206	823	1987	-2.24	98.74	1.26
Osogbo	1935-2000	1278	223	708	1946	-2.56	99.47	0.53
Ondo	1906-2000	1582	329	1067	1920	-1.57	94.18	5.82
Ibadan	1905-2000	1249	262	708	1912	-2.06	98.03	1.97
Lagos	1901-2000	1813	416	737	1982	-2.59	99.52	0.48
Enugu	1916-2000	1792	407	913	1983	-2.16	98.46	1.54
P/Harcourt	1902-2000	2696	570	1683	1958	-1.78	96.25	3.75
Calabar	1902-2000	2995	460	2105	1919	-1.93	97.32	2.68
Warri	1908-2000	2787	317	2072	1977	-2.26	99.57	0.43
Benin	1906-2000	2086	359	1228	1986	-2.39	99.15	0.85
Gusau	1953-2000	933	182	676	1984	-1.41	92.07	7.93

*LARE = Least annual rainfall ever recorded

*Probability A = Probability of receiving more than the least annual rainfall ever recorded.

*Probability B = Probability of receiving less than the least annual rainfall ever recorded

This space is for figure 4.14

Table 4.8 provides the percentage probability of receiving more than the highest annual rainfall ever recorded at individual stations in the country while figure 4.15 shows the spatial pattern of the probability of receiving more than the highest annual rainfall ever recorded at individual stations in Nigeria. The results show that Ikom, a southern station, records the highest probability (9.01%) of receiving more than the highest annual rainfall of 2674mm over the period 1972-2000. This suggests that Ikom stands the riks of flood than any other station in the country. Fig. 4.15 presents the spatial pattern of the probability of receiving more than the highest annual rainfall ever recorded at individual stations in the country. The results further show that the risk of floods in the country tends to be localized rather than the large-scale pattern which is often associated with drought.

Table 4. 8. Probability of receiving more than the highest annual rainfall ever recorded at individual stations in Nigeria, 1901-2000.

Station	Length of record	Mean (mm)	Standard deviation (mm)	*HARE (mm)	Year of occurrence	Z-Score	Probability A (%)	Probability B (%)
Bauchi	1908-2000	1066	172	1454	1932	+2.26	98.09	1.91
Jos	1922-2000	1332	175	1711	1923	+2.17	98.50	1.50
Kano	1905-2000	845	369	1872	1998	+2.78	99.72	0.28
Katsina	1918-2000	641	174	956	1922	+1.81	96.49	3.51
Kaduna	1930-2000	1246	184	1697	1957	+2.45	99.28	0.72
Nguru	1942-2000	475	144	719	1954	+1.69	95.45	4.55
Maiduguri	1915-2000	622	150	889	1967	+1.78	96.25	3.75
Sokoto	1916-2000	705	149	1025	1936	+2.15	98.42	1.58
Zaria	1943-2000	1066	161	1457	1962	+2.43	99.24	0.76
Yelwa	1926-2000	1002	190	1566	1999	+2.97	99.85	0.15
Yola	1914-2000	926	186	1369	1941	+2.38	99.13	0.87
Makurdi	1927-2000	1308	239	2007	1953	+2.92	99.82	0.18
Lokoja	1916-2000	1191	233	1871	1927	+2.91	99.81	0.19
Ikom	1972-2000	2237	325	2674	1981	+1.34	90.99	9.01
Ilorin	1916-2000	1253	228	1737	1968	+2.12	98.30	1.70
Bida	1928-2000	1206	216	2083	1931	+4.06	99.99	0.01
Minna	1916-2000	1284	206	1891	1947	+2.60	99.53	0.47
Osogbo	1935-2000	1278	223	1907	1968	+2.83	99.76	0.24
Ondo	1906-2000	1582	329	2542	1985	+2.92	99.82	0.18
Ibadan	1905-2000	1249	262	1966	1980	+2.74	99.69	0.31
Lagos	1901-2000	1813	416	3265	1968	+3.50	99.97	0.03
Enugu	1916-2000	1792	407	4205	1918	+5.93	99.99	0.01
P/Harcourt	1902-2000	2696	570	4251	1905	+2.73	99.68	0.32
Calabar	1902-2000	2995	460	4252	1905	+2.73	99.68	0.32
Warri	1908-2000	2787	317	3586	1918	+2.52	99.41	0.59
Benin	1906-2000	2086	359	3147	1949	+2.96	99.84	0.16
Gusau	1953-2000	933	182	1507	1994	+3.15	99.97	0.03

*HARE = Highest annual rainfall ever recorded

*Probability A = Probability of receiving less than the highest annual rainfall ever recorded.

*Probability B = Probability of receiving more than the highest annual rainfall ever recorded

This space is for figure 4.15

CHAPTER FIVE

RELATIONSHIP OF EL-NINO/SOUTHERN OSCILLATION, SEA SURFACE TEMPERATURE ANOMALIES TO RAINFALL ANOMALIES IN NIGERIA

The links between the El-Nino/Southern Oscillation (ENSO) and large scale precipitation patterns have been examined since the earliest studies of this phenomenon. Indeed the pioneering studies by Walker (1923, 1924, 1928) and Walker and Bliss (1930, 1932, 1937), which first documented ENSO on a global scale, are motivated by attempts to understand and predict variations in Indian monsoon rainfall. The studies then expanded to studies of precipitation around the globe. Since the publication of these early works, several other studies have found evidences of precipitation-ENSO relationships. Berlage (1966) found evidence of significant correlation between indices of Southern Oscillation and precipitation at several individual stations around the globe. Ramusson and Carpenter (1982) and Shukla and Paolino (1983) also confirmed that the summer monsoon precipitation over Peninsular India was suppressed during ENSO years.

Although ENSO has been shown to be one of the primary determinants of the inter-annual variability of rainfall in low latitudes, its influence over Africa remains controversial. A number of studies have confirmed a relationship between rainfall and ENSO in parts of eastern and southern Africa (e.g. Lindesay et al., 1986; Farmer, 1988; Janowiak, 1988; Van Heerden et al., 1988; Nicholson, 1996; Pu and Cook, 2001; Reason, 2002, Reason *et al*, 2006). However, there is disagreement about its influence in the Sahel and elsewhere (e.g. Wolter, 1989; Semazzi et al., 1988; Nicholson and Palao, 1993, Nicholson, 2001). The most comprehensive studies of its influence on Africa are those of Nicholson and Entekhabi (1986), Popelewski and Halpert (1987, 1989). Using harmonic analysis method, Popelewski and Halpert (1987) investigate the global and regional scale precipitation patterns associated with the El-Nino/Southern Oscillation (ENSO). They used monthly precipitation time series from over 1700 stations for the period 1875-1982 and derived monthly ranked precipitation composites over idealized 2-year ENSO episodes for all the stations that include data for at least five ENSOs. The results show that, in addition to the Pacific Ocean basin where precipitation pattern could be directly related to the ENSO, several other regions, which showed consistent ENSO-related precipitation, were indentified. Specifically, four regions in Australia, two regions each in North America, South

America, the Indian subcontinent and Africa, and one region in Central America were all found to have coherent ENSO-related precipitation.

However, for a more regionally focused study, the present study examines the nature of the interaction between Nigerian rainfall and the coupled ocean-atmospheric phenomena of El-Nino/Southern Oscillation (ENSO) and Sea Surface Temperature anomalies. A number of techniques have been employed by several climate scientists to examine the relationship of the El-Nino/Southern Oscillation (ENSO) and Sea Surface Temperature anomalies to rainfall in different parts of the world. These include harmonic analysis (Popelewski and Halpert, 1987, 1989; Nicholson and Jeeyoung, 1997), principal component analysis (Mason, 1995), spectral analysis (Nicholson and Entekhabi, 1987), and the general circulation of the atmosphere simulation models (Folland *et al*., 1986). In the present study, a stepwise multiple regression model was employed to examine the nature of the relationships among El-Nino/Southern Oscillation (ENSO), Sea Surface Temperatures (SSTs) anomalies and rainfall over Nigeria. The choice of this model was informed by the numerical problems associated with the spectral and harmonic analyses in ENSO-rainfall teleconnections studies which include inability to deal with the multicolinearity between and among several independent variables as discussed in Ayeni (1979).

5.1. El-Nino/Southern Oscillation (ENSO), Sea Surface Temperature (SSTs) and Nigerian Rainfall.

The step wise multiple regression model is used to evaluate the relationships among the annual rainfall anomalies in Nigeria, the anomaly indices of sea surface temperature (SST) and ENSO. The relationship is assessed for each of the four zones considered in this study. The detail of the procedure is given in chapter three. The result is presented in table 5.1. It could be observed from table 5.1 that none of the critical variables (predictors) accounts for annual rainfall anomalies in Nigeria and for South-west and south-eastern regions of the country. However, the South Atlantic SST anomaly accounts for 12.0% (r^2 = 0.120) of the explained variance in annual anomalies for the northern part of the country while for the central (middle belt), it is only the North Atlantic SST anomaly that is significant. It, however, accounts for only12.4% (r^2 = 0.124) of the explained variance in annual rainfall anomalies for the central (middle belt) region of the country. This result could not be regarded as good enough because of the low values of r^2 and could not be used for any meaningful prediction of rainfall anomalies in the affected regions.

Table 5.1. Summary of the regression between annual rainfall anomalies and anomaly indices of SST and ENSO in Nigeria.

Country/Region	Critical Variable	R	r^2	Significance
Nigeria	-	-	-	-
North	South Atlantic SSTA	0.35	0.120	0.01
Central	North Atlantic SSTA	0.35	0.124	0.01
South-West	-	-	-	-
South-East	-	-	-	-

5.2. Regional impact of El-Nino/Southern Oscillation (ENSO) and Sea Surface Temperatures (SSTs) on rainfall in Nigeria.

Tables 5.2-5.6 provide correlation coefficients between annual rainfall anomalies and anomaly indices of SST and ENSO in Nigeria and for each of the four regions considered in this study. It could be observed that annual rainfall anomalies for Nigeria significantly correlate with annual rain days anomalies (0.17*) but none of the predictors significantly correlates with annual rainfall anomalies for Nigeria. The correlation is however significant between annual rainfall and rain days anomalies for the northern region (0.72**). Annual rainfall anomalies for the north correlate significantly with North Atlantic and South Atlantic SST anomalies (-0.35*) each. Similarly, none of the predictors significantly correlate with the annual rainfall anomalies for the central, south-west and south-east respectively. However, the correlation between annual rainfall and rain days anomalies is significant for the central, south-west and south-east (0.67**), (0.55**) and (0.23*) for the central, south-west and south-eastern regions respectively. The results obtained and presented in the form of correlation matrix show a poor relationship between annual rainfall anomalies and several predictors considered in this study.

Table 5.2. Correlation coefficients among annual rainfall anomalies, annual raindays anomalies and anomaly indices of SSTs and ENSO for Nigeria

X1	1.00									
X2	0.17*	1.00								
X3	-0.13	-0.32	1.00							
X4	-0.13	-0.32	1.00**	1.00						
X5	-0.05	-0.23**	0.54**	0.54**	1.00					
X6	-0.17	0.06	-0.03	-0.03	0.47**	1.00				
X7	-0.11	0.19	-0.01	-0.01	0.39**	0.92**	1.00			
X8	-0.04	-0.02	-0.08	-0.08	0.38**	0.59**	0.79**	1.00		
X9	-0.08	0.18	-0.17	-0.17	0.31*	0.78**	0.94**	0.93**	1.00	
X10	0.08	-0.06	0.11	0.11	-0.36**	-0.71**	-0.86**	-0.89**	-0.93**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9	X10

**Significant at 0.01 level of significance

*Significant at 0.05 level of significance

X1= Annual Rainfall Anomalies (Nigeria)

X2= Annual Rain days Anomalies (Nigeria)

X3= North Atlantic SST anomalies

X4= South Atlantic SST anomalies

X5= Global SST anomalies

X6= NINO 1+2 SST anomalies

X7= NINO 3 SST anomalies

X8= NINO 4 SST anomalies

X9= NNO 3.4 SST anomalies

X10= ENSO Index

Table 5.3. Correlation coefficients among annual rainfall anomalies, annual raindays anomalies and anomaly indices of SSTs and ENSO for the North.

X1	1.00									
X2	0.72**	1.00								
X3	-0.35*	-0.41**	1.00							
X4	-0.35*	-0.41**	1.00**	1.00						
X5	-0.07	-0.36	0.54**	0.54**	1.00					
X6	.017	0.12	-0.03	-0.03	0.47**	1.00				
X7	-0.09	-0.17	-0.01	-0.01	0.39**	0.92**	1.00			
X8	-0.01	0.02	-0.08	-0.08	0.38**	0.59**	0.79**	1.00		
X9	-0.11	-0.14	-0.17	-0.17	0.31*	0.78**	0.94**	0.93**	1.00	
X10	0.11	0.01	0.11	0.11	-0.36**	-0.71**	-0.86**	-0.89**	-0.93**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9	X10

**Significant at 0.01 level of significance

*Significant at 0.05 level of significant

X1= Annual Rainfall Anomalies (North)

X2= Annual Rain days Anomalies (North)

X3= North Atlantic SST anomalies

X4= South Atlantic SST anomalies

X5= Global SST anomalies

X6= NINO 1+2 SST anomalies

X7= NINO 3 SST anomalies

X8= NINO 4 SST anomalies

X9= NNO 3.4 SST anomalies

X10= ENSO Index

Table 5.4 .Correlation coefficients among annual rainfall anomalies, annual raindays anomalies and anomaly indices of SSTs and ENSO for the Middle Belt/Central

X1	1.00									
X2	0.67**	1.00								
X3	-0.15	-0.40**	1.00							
X4	-0.15	-0.40**	1.00**	1.00						
X5	-0.23	-0.49**	0.54**	0.54**	1.00					
X6	-0.06	-0.10	-0.03	-0.03	0.47**	1.00				
X7	-0.35	-0.27	-0.10	-0.10	0.39**	0.92**	1.00			
X8	0.03	-0.14	-0.08	-0.08	0.38**	0.59**	0.79**	1.00		
X9	0.11	0.03	-0.17	-0.17	0.31*	0.78**	0.94**	0.93**	1.00	
X10	0.13	0.20	0.11	0.11	-0.36**	-0.71**	-0.86**	-0.89**	-0.93**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9	X10

**Significant at 0.01 level of significance

*Significant at 0.05 level of significance

X1= Annual Rainfall Anomalies (Central)

X2= Annual Rain days Anomalies (Central)

X3= North Atlantic SST anomalies

X4= South Atlantic SST anomalies

X5= Global SST anomalies

X6= NINO 1+2 SST anomalies

X7= NINO 3 SST anomalies

X8= NINO 4 SST anomalies

X9= NNO 3.4 SST anomalies

X10= ENSO Index

Table 5.5. Correlation coefficients among annual rainfall anomalies, annual raindays anomalies and anomaly indices of SSTs and ENSO for the South-West

X1	1.00									
X2	0.55**	1.00								
X3	0.05	-0.45**	1.00							
X4	0.05	-0.45**	1.00**	1.00						
X5	-0.09	-0.35*	0.54**	0.54**	1.00					
X6	-0.13	-0.13	-0.03	-0.03	0.47**	1.00				
X7	-0.28	-0.32	-0.10	-0.10	0.39**	0.92**	1.00			
X8	-0.08	-0.16	-0.08	-0.08	0.38**	0.59**	0.79**	1.00		
X9	-0.04	0.07	-0.17	-0.17	0.31*	0.78**	0.94**	0.93**	1.00	
X10	0.10	0.11	0.11	0.11	-0.36**	-0.71**	-0.86**	-0.89**	-0.93**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9	X10

**Significant at 0.01 level of significance

*Significant at 0.05 level of significance

X1= Annual Rainfall Anomalies (South-West)

X2= Annual Rain days Anomalies (South-West)

X3= North Atlantic SST anomalies

X4= South Atlantic SST anomalies

X5= Global SST anomalies

X6= NINO 1+2 SST anomalies

X7= NINO 3 SST anomalies

X8= NINO 4 SST anomalies

X9= NNO 3.4 SST anomalies

X10= ENSO Index

Table 5.6. Correlation coefficients among annual rainfall anomalies, annual raindays anomalies and anomaly indices of SSTs and ENSO for the South-East

X1	1.00									
X2	0.24*	1.00								
X3	-0.15	-0.14	1.00							
X4	-0.15	-0.14	1.00**	1.00						
X5	-0.12	0.11	0.54**	0.54**	1.00					
X6	-0.14	-0.17	-0.03	-0.03	0.47**	1.00				
X7	-0.26	-0.01	-0.10	-0.10	0.39**	0.92**	1.00			
X8	0.02	-0.24	-0.08	-0.08	0.38**	0.59**	0.79**	1.00		
X9	0.03	0.11	-0.17	-0.17	0.31*	0.78**	0.94**	0.93**	1.00	
X10	0.12	0.21	0.11	0.11	-0.36**	-0.71**	-0.86**	-0.89**	-0.93**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9	X10

**Significant at 0.01 level of significance

*Significant at 0.05 level of significance

X1= Annual Rainfall Anomalies (South-East)

X2= Annual Rain days Anomalies (South-East)

X3= North Atlantic SST anomalies

X4= South Atlantic SST anomalies

X5= Global SST anomalies

X6= NINO 1+2 SST anomalies

X7= NINO 3 SST anomalies

X8= NINO 4 SST anomalies

X9= NNO 3.4 SST anomalies

X10= ENSO Index

The relationship is further examined using the wet season rainfall of the months within which each region receives bulk of its annual rainfall. The rainfall series is then normalized for each region and for the whole country which is then regressed on the anomaly indices of SST and ENSO using the step wise multiple regression model. For the entire country, a composite of (Apr+May+Jun+Jul+Aug) rainfall is generated from all the stations used for this study for the period 1951-2000. For the north and central, a composite of (Jun+Jul+Aug) rainfall is derived from the stations in each of the zones while for the south-west and south-east, a composite of (Apr+May+Jun) rainfall is derived from the stations considered in each of the zones for the period 1951-2000. These series are then normalized and regressed on the anomaly indices of SST and ENSO using step wise multiple regression model. The result is presented in table 5.7. It could be observed from table 5.7 that none of the critical variables accounts for the wet season rainfall anomalies for Nigeria and south-eastern part of the country. However, for the north, it is only South Atlantic SST anomaly that is significant and accounts for only 12.7% of the explained variance in rainfall anomalies. In the middle belt region, the North Atlantic SST anomaly seems to be the only critical variable that is significant and accounts for only 10.2% of the explained variance in rainfall anomalies for the region. For the south-western part of Nigeria, South Atlantic SST anomaly appears to be the only critical variable that dominates the rainfall anomlies in the region, even though it accounts for only 8.0% of the explained variance in rainfall anomalies for the region. The results obtained here could still not be regarded as good enough for any meaningful prediction of rainfall anomalies in Nigeria due to lower values of r^2.

Table 5.7.Summary of the regression between wet season rainfall anomalies and anomaly indices of SST and ENSO in Nigeria

Country/Region	Critical Variable	r	r^2	Significance
Nigeria	-	-	-	-
North	South Atlantic SSTA	0.356	0.127	0.01
Central	North Atlantic SSTA	0.319	0.102	0.01
South-West	South Atlantic SSTA	0.282	0.080	0.01
South-East	-	-	-	-

The results of the multiple-correlation analysis between the wet season rainfall anomalies for Nigeria and anomaly indices of SSTs and ENSO are presented in table 5.8. It could be observed that the wet season rainfall anomalies for Nigeria significantly correlates with the rain days anomalies for Nigeria (0.56**). This implies that any anomaly in rainfall over Nigeria could lead to significant anomalies in rain days in Nigeria. It should be noted, however, that none of the predictors considered that significantly correlates with rainfall anomalies in Nigeria as presented in table 5.8. However, the correlation between rain days anomalies and global SST anomalies is significant but negative (-0.40**). There is also a significant but negative correlation between rain days anomalies for Nigeria and NINO 1+2 SST anomalies (-0.29*). The rain days anomalies for Nigeria also correlate with the North and South Atlantic SST anomalies having a coefficient of (-0.42**) each for the North and South Atlantic SST anomalies respectively.

Table 5.8. Correlation coefficients among wet season rainfall anomalies, wet season raindays anomalies and anomaly indices of SSTs and ENSO for Nigeria

X1	1.00									
X2	0.56**	1.00								
X3	-0.09	0.18	1.00							
X4	0.11	-0.40**	-0.36**	1.00						
X5	-0.05	-0.29*	-0.71**	0.47**	1.00					
X6	0.05	-0.23	-0.86**	0.39**	0.92**	1.00				
X7	0.09	-0.15	-0.93**	0.31*	0.78**	0.94**	1.00			
X8	0.17	-0.11	-0.89**	0.38**	0.59**	0.79**	0.93**	1.00		
X9	0.04	-0.42**	0.11	0.54**	-0.03	-0.10	-0.17	-0.08	1.00	
X10	0.04	-0.42**	0.11	0.54**	-0.03	-0.10	-0.17	-0.08	1.00**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9	X10

**Significant at 0.01 level of significance

*Significant at 0.05 level of significance

X1= Rainfall Anomalies (Nigeria)

X2= Rain days Anomalies (Nigeria)

X3= ENSO Index

X4= Global SST Anomalies

X5= NINO 1+2 SST anomalies

X6= NINO 3 SST Anomalies

X7= NINO 3.4 SST anomalies

X8= NINO 4 SST anomalies

X9= North Atlantic SST Anomalies

X10= South Atlantic SST Anomalies

Similarly, the results of the correlation analysis between wet season rainfall anomalies for the North and anomaly indices of SST and ENSO are presented in table 5.9. It could be noted that the wet season rainfall anomalies for the North significantly correlate with the rain days anomalies (0.51**). However, apart from the North and South Atlantic SST anomalies, none of the predictors significantly correlates with the wet season rainfall anomalies for the North. The rain days anomalies for the North significantly correlate with the global SST anomlalies (-0.36**), North Atlantic SST anomalies (-0.41**) and South Atlantic SST anomalies (-0.41**). For the Central (Middle belt) region of the country, the wet season rainfall anomalies only correlate with North Atlantic SST anomalies (0.32*) and South Atlantic SST anomalies (0.32*). The remaining predictors did not show any significant correlation with the wet season rainfall anomalies for the central region of Nigeria. Surprisingly, the correlation between wet season rainfall and rain days anomalies for the central region of Nigeria is not significant (0.16). Besides, rain days anomalies for the central region of Nigeria significantly but negatively correlate with the global SST anomalies (-0.49**), NINO 1+2 SST anomalies (-0.71**), NINO 3 SST anomalies (-0.86**), NINO 3.4 SST anomalies (-0.93**) and NINO 4 SST anomalies (-0.89**). (see Table 5.10)

Table 5.9. Correlation coefficients among wet season rainfall anomalies, wet season raindays anomalies and anomaly indices of SSTs and ENSO for the North

X1	1.00									
X2	0.51**	1.00								
X3	0.16	0.10	1.00							
X4	-0.04	-0.36**	-0.36**	1.00						
X5	-0.15	-0.14	-0.71**	0.47**	1.00					
X6	-0.17	-0.10	-0.86**	0.39**	0.92	1.00				
X7	-0.16	-0.08	-0.93**	0.31**	0.78**	0.94**	1.00			
X8	-0.17	-0.13	-0.89**	0.38**	0.59**	0.79**	0.93**	1.00		
X9	-0.36**	-0.41**	0.11	0.54**	0.59**	-0.10	-0.17	-0.08	1.00	
X10	-0.36**	-0.41**	0.11	0.54**	-0.03	-0.10	-0.17	-0.08	1.00**	1.00
	X1	X2	X3	X4	X5-.032	X6	X7	X8	X9	X10

**Significant at 0.01 level of significance

*Significant at 0.05 level of significance

X1= Rainfall Anomalies (North)

X2= Rain days Anomalies (North)

X3= ENSO Index

X4= Global SST Anomalies

X5= NINO 1+2 SST anomalies

X6= NINO 3 SST Anomalies

X7= NINO 3.4 SST anomalies

X8= NINO 4 SST anomalies

X9= North Atlantic SST Anomalies

X10= South Atlantic SST Anomalies

Table 5.10. Correlation coefficients among wet season rainfall anomalies, wet season raindays anomalies and anomaly indices of SSTs and ENSO for Middle Belt/Central

X1	1.00									
X2	0.16	1.00								
X3	0.01	0.20	1.00							
X4	0.05	-0.49**	-0.36**	1.00						
X5	-0.06	-0.71**	-0.71**	0.47**	1.00					
X6	0.01	-0.86**	-0.86**	0.39**	0.92**	1.00				
X7	-0.02	-0.93**	-0.93**	0.31*	0.78**	0.94**	1.00			
X8	0.02	-0.89**	-0.89**	0.38**	0.59**	0.79**	0.93**	1.00		
X9	0.32*	0.11	0.11	0.54**	-0.03	-0.10	-0.17	-0.08	1.00	
X10	0.32*	0.11	0.11	0.54**	-0.03	-0.10	-0.17	-0.08	1.00**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9	X10

**Significant at 0.01 level of significance

*Significant at 0.05 level of significance

X1= Rainfall Anomalies (Central/Middle Belt)

X2= Rain days Anomalies (Central/Middle Belt)

X3= ENSO Index

X4= Global SST Anomalies

X5= NINO 1+2 SST anomalies

X6= NINO 3 SST Anomalies

X7= NINO 3.4 SST anomalies

X8= NINO 4 SST anomalies

X9= North Atlantic SST Anomalies

X10= South Atlantic SST Anomalies

For the South-Western part of Nigeria, the wet season rainfall anomalies correlate with only North Atlantic SST anomalies (-0.28*) and South Atlantic SST anomalies (-0.28*) (see Table 5.11). However, all other predictors did not show any significant correlation with wet season rainfall anomalies in the region. It should be noted, however, that the correlation between wet season rainfall and rain days anomalies for the South-West is positive and significant (0.49**). Table 5.11 also shows that rain days anomalies for the South-Western region of Nigeria significantly correlate with global SST anomalies (-0.35*), North Atlantic SST anomalies (-0.45**) and South Atlantic SST anomalies (-0.45**). It could also be observed that the correlation analysis between wet season rainfall anomalies for the South-Eastern part of Nigeria and the selected predictors shows no significant correlation between the wet season rainfall anomalies in the region and the considered predictors. Simailarly, the rain days anomalies for the South-Eastern part of Nigeria did not significantly correlate with any of the predictors considered in this study (see Table 5.12).

The results so far discussed in this section confirmed several other studies in Africa and else where which revealed a weak relationship between El-Nino/Southern Oscillation (ENSO) and West African rainfall (see Popelewski and Halpert, 1987, 1989; Nicholson, 1980, 1981, 1986a, 1993). Some teleconnection researches, however, established a strong relationship between El-Nino/Southern Oscillation (ENSO) and Southern African rainfall (Nicholson and Kim, 1997; Masson, 1995; Mulenga *et al*, 2003; Reason and Jagadheesha, 2005).

Table 5.11. Correlation coefficients among wet season rainfall anomalies, wet season raindays anomalies and anomaly indices of SSTs and ENSO for South-West

X1	1.00									
X2	0.49**	1.00								
X3	0.14	0.11	1.00							
X4	-0.12	-0.35*	-0.36**	1.00						
X5	-0.19	-0.18	-0.71**	0.47**	1.00					
X6	-0.16	-0.13	-0.86**	0.39**	0.92**	1.00				
X7	-0.07	-0.07	-0.93**	0.31*	0.78**	0.94**	1.00			
X8	-0.01	-0.05	-0.89**	0.38**	0.59**	0.79**	0.93**	1.00		
X9	-0.28*	-0.45**	0.11	0.54**	-0.03	-0.10	-0.17	-0.08	1.00	
X10	-0.28*	-0.45**	0.11	0.54**	-0.03	-0.10	-0.17	-0.08	1.00**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9	X10

**Significant at 0.01 level of significance

*Significant at 0.05 level of significance

X1= Rainfall Anomalies (South-West)

X2= Rain days Anomalies (South-West)

X3= ENSO Index

X4= Global SST Anomalies

X5= NINO 1+2 SST anomalies

X6= NINO 3 SST Anomalies

X7= NINO 3.4 SST anomalies

X8= NINO 4 SST anomalies

X9= North Atlantic SST Anomalies

X10= South Atlantic SST Anomalies

Table 5.12. Correlation coefficients among wet season rainfall anomalies, wet season raindays anomalies and anomaly indices of SSTs and ENSO for South-East

X1	1.00									
X2	-0.08	1.00								
X3	-0.11	0.21	1.00							
X4	0.10	0.11	-0.36**	1.00						
X5	0.01	-0.13	-0.71**	0.47**	1.00					
X6	0.05	-0.18	-0.86**	0.39**	0.92**	1.00				
X7	0.11	-0.21	-0.93**	0.31*	0.78**	0.94**	1.00			
X8	0.16	-0.18	-0.89**	0.38**	.059**	0.79**	0.93**	1.00		
X9	-0.19	-0.14	0.11	0.54**	-0.03	-0.10	-0.17	-0.08	1.00	
X10	-0.19	-0.14	0.11	0.54**	-0.03	-0.10	-0.17	-0.08	1.00**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9	X10

**Significant at 0.01 level of significance

*Significant at 0.05 level of significance

X1= Rainfall Anomalies (South East)

X2= Rain days Anomalies (South East)

X3= ENSO Index

X4= Global SST Anomalies

X5= NINO 1+2 SST anomalies

X6= NINO 3 SST Anomalies

X7= NINO 3.4 SST anomalies

X8= NINO 4 SST anomalies

X9= North Atlantic SST Anomalies

X10= South Atlantic SST Anomalies

Following a relatively weak relationship between the rainfall anomalies in Nigeria and anomaly indces of El-Nino/Southern Oscillation (ENSO) and Sea Surface Temperature (SST) as shown in the previous tables (Tables 5.1-5.12), the relationship between Nigerian rainfall and the coupled ocean-atmospheric phenomena of El-Nino/southern Oscillation (ENSO) and Sea Surface Temperature (SST) is re-examined using the actual values of rainfall amounts and indices of ENSO and SSTs against the normalized series that was previously utilized in this work. It should however be noted that, though, the actual values of rainfall and indices of ENSO and SSTs are used against the normalized series in examining the relationship between Nigerian rainfall and ENSO and SSTs, the rainfall composites used earlier in this section have been retained for uniformity. However, the rainfall composites derived for Nigeria and each of the four regions comprised of actual values of rainfall (i.e non-normalized series). For Nigeria, a composite of (Apr+May+Jun+Jul+Aug) rainfall is generated from all the stations considered for the period 1951-2000. For the north and central, a composite of (Jun+Jul+Aug) rainfall is derived from the stations considered in each of the zones while for the south-west and south-east, a composite of (Apr+May+Jun) rainfall is derived from the stations considered in each of the zones.

The relationship between Nigrian rainfall and the coupled ocean-atmospheric phenomena of ENSO and SSTs is first re-exmined using the annual rainfall amounts for Nigeria and each of the four regions considerd for the same period 1951-2000. This is followed by the derived-composite rainfall series for Nigeria and for each of the four regions considered in this work. The result of the correlation analysis between annual rainfall for Nigeria and the indices of SSTs and ENSO is presented in table 5.13 while for the North, Central, South-West and South-Eastern regions of Nigeria, the result of the correlation analysis between annual rainfall and indices of SSTs and ENSO in each of the regions is presented in tables 5.14-5.17. It is interesting to note that for Nigeria (Table 5.13); the annual rainfall for Nigeria does not significantly correlate with any of the predictors considered in this study. The annual rainfall for the North does not also significantly correlate with any of the predictors considered in this work (see Table 5.14). There is also no significant correlation between annual rainfall for the Central (Middle belt) region and each of the predictors considered in this work (see table 5.15). Similar results are obtained for the South-West and South-Eastern regions of Nigeria (see Tables 5.16 and 5.17).

Table 5.13. Correlation coefficients between annual rainfall for Nigeria and indices of SSTs and ENSO

X1	1.00								
X2	0.23	1.00							
X3	-0.10	0.03	1.00						
X4	-0.11	0.49**	0.37**	1.00					
X5	-0.18	0.28	-0.05	0.72**	1.00				
X6	-0.11	0.26	-0.09	0.77**	0.92**	1.00			
X7	-0.05	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	-0.08	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	-0.03	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01 level of significance

*Significant at 0.05 level of significance

X1= Annual Rainfall (Nigeria)

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.14. Correlation coefficients between annual rainfall for the North and indices of SSTs and ENSO

X1	1.00								
X2	-0.09	1.00							
X3	-0.11	0.03	1.00						
X4	-0.21	0.49**	0.37**	1.00					
X5	-0.17	0.28	-0.05	0.72**	1.00				
X6	-0.18	0.26	-0.09	0.77*	0.92**	1.00			
X7	-0.11	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	-0.13	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	0.18	0.14	0.19	-0.40**	-0.45*	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01 level of significance

*Significant at 0.05 level of significance

X1= Annual Rainfall (North)

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.15. Correlation coefficients between annual rainfall for the Central/Middle belt and the indices of SSTs and ENSO

X1	1.00								
X2	-0.04	1.00							
X3	0.26	0.03	1.00						
X4	0.23	0.49**	0.37**	1.00					
X5	0.01	0.28	-0.05	0.72**	1.00				
X6	0.02	0.26	-0.09	0.77**	0.92**	1.00			
X7	0.05	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	0.01	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	0.04	0.14	0.19	-0.40**	-0.45**	0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01 level of significance

*Significant at 0.05 level of significance

X1= Annual Rainfall (Central/Middle belt)

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.16. Correlation coefficients between annual rainfall for South-West and indices of SSTs and ENSO

X1	1.00								
X2	0.16	1.00							
X3	0.05	0.03	1.00						
X4	-0.09	0.49**	0.37**	1.00					
X5	-0.23	0.28	-0.05	0.72**	1.00				
X6	-0.14	0.26	-0.09	0.77**	0.92**	1.00			
X7	0.01	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	-0.07	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	0.02	0.14	0.19	-.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01 level of significance

*Significant at 0.05 level of significance

X1= Annual Rainfall (South-West)

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.17. Correlation coefficients between annual rainfall for South-East and indices of SSTs and ENSO

X1	1.00								
X2	-0.01	1.00							
X3	-0.10	0.03	1.00						
X4	-0.21	0.49**	0.37**	1.00					
X5	-0.22	0.28	-0.05	0.72**	1.00				
X6	-0.21	0.26	-0.09	0.77**	0.92**	1.00			
X7	-0.12	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	-0.18	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	0.12	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Annual Rainfall (South-East)

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

The result of the multiple-correlation analysis between the wet season rainfall for Nigeria and indices of SSTs and ENSO is presented in table 5.18. It could be observed from table 5.18 that the wet season rainfall for Nigeria does not significantly correlate with any of the independent variables considered in this work. In other words, there is no significant relationship between wet season rainfall for Nigeria and indices of SSTs and ENSO. For the North (Table 5.19), the wet season rainfall appears to significantly correlate with only the South Atlantic SST indices, though, the correlation is negative (-0.35*). For the Central and South-Western regions of the country (tables 5.20 and 5.21), the wet season rainfall does significantly correlate with only the South Atlantic SST indices, with a coefficient of 0.32* and -0.28* for the Central and South-Western regions respectively. For the South-Eastern region of Nigeria, the wet season rainfall correlates with only North Atlantic SST indices (see Table 5.22).

Overall, these results further confirmed the conclusion of other teleconnection researches (Popelewski and Halpert, 1987, 1989; Folland *et al*., 1986; Semezzi *et al*., 1988; Nicholson and Kim, 1997; Janowiak, 1988) that Sea Surface Temperatures (SSTs) in the tropical oceans (Atlantic, Pacific and Indian oceans), rather than El-Nino/Southern Oscillation (ENSO), have the dominant control on rainfall variability over Africa. The apparent relationships between El-Nino/Southern Oscillation (ENSO) and rainfall are manifestations of the influence of ENSO on these oceans. A similar conclusion was reached by Webster (1981, 1983), Folland, *et al* (1986) and Goddard and Graham (1999), using Reg GCM model simulations.

From the result of the relationships between Nigerian rainfall and air-sea interaction phenomena of El-Nino/Southern Oscillation (ENSO) and Sea Surface Temperature (SSTs), it has increasingly becomes clear that the El-Nino/Southern Oscillation (ENSO) appears to be less significant in terms of the observed inter-anaual variability of rainfall in Nigeria while the Sea Surface Temperature (SSTs) over the tropical oceans (Atlantic and Pacific oceans) seems to have dominated the inter-annual and intra-seasonal rainfall variability in Nigeria as revealed in this section. It should however be noted that the results so far presented in tables 5.1 – 5.22 are based on the aggregated rainfall amounts for 27 stations in the case of Nigeria, 9 stations in the of the North, 8 stations for the Central region, 6 stations in the case of South-West and 4 stations in the case of South-Eastern region of Nigeria. What could be deduced from the results presented in tables 5.1-5.22 is that the ENSO influence on rainfall in

Nigeria is actually manifested through its effects in tropical oceans (Atlantic, Pacific and Indian oceans).

Table 5.18. Correlation coefficients between wet season rainfall for Nigeria and indices of SSTs and ENSO

X1	1.00								
X2	0.19	1.00							
X3	0.04	0.03	1.00						
X4	0.10	0.49**	0.37**	1.00					
X5	-0.05	0.28	-0.05	0.72**	1.00				
X6	0.05	0.26	-0.09	0.77**	0.92**	1.00			
X7	0.17	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	0.11	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	0.13	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Wet Season Rainfall (Nigeria)

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.19. Correlation coefficients between wet season rainfall for the North and indices of SSTs and ENSO

X1	1.00								
X2	0.22	1.00							
X3	-0.35*	0.03	1.00						
X4	-0.19	0.49**	0.37**	1.00					
X5	-0.15	0.28	-0.05	0.72**	1.00				
X6	-0.17	0.26	-0.09	0.77**	0.92**	1.00			
X7	-0.16	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	-0.14	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	0.19	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Wet Season Rainfall (North)

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.20. Correlation coefficients between wet season rainfall for the (Central/Middle Belt) and indices of SSTs and ENSO

X1	1.00								
X2	0.14	1.00							
X3	0.32*	0.03	1.00						
X4	0.11	0.49**	0.37**	1.00					
X5	-0.06	0.28	-0.05	0.72**	1.00				
X6	0.01	0.26	-0.09	0.77**	0.92**	1.00			
X7	0.02	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	0.01	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	-0.02	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Wet Season Rainfall (Central/Middle Belt)

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.21. Correlation coefficients between wet season rainfall for the South-West and indices of SSTs and ENSO

X1	1.00								
X2	0.17	1.00							
X3	-0.28*	0.03	1.00						
X4	-0.15	0.49**	0.37**	1.00					
X5	-0.19	0.28	-0.05	0.72**	1.00				
X6	-0.15	0.26	-0.09	0.77**	0.92**	1.00			
X7	-0.01	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	-0.08	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	0.03	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Wet Season Rainfall (South-East)

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.22. Correlation coefficients between wet season rainfall for the South-East and indices of SSTs and ENSO

X1	1.00								
X2	0.29*	1.00							
X3	-0.19	0.03	1.00						
X4	0.10	0.49**	0.37**	1.00					
X5	0.01	0.28	-0.05	0.72**	1.00				
X6	0.06	0.26	-0.09	0.77**	0.92**	1.00			
X7	0.16	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	0.10	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	-0.20	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Wet Season Rainfall (South-East)

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

The relationship between Nigerian rainfall and the couple ocean-atmospheric phenomena of El-Nino/Southern Oscillation (ENSO) and Sea Surface Temperature (SSTs) is also investigated using five selected stations located in the soudano-sahelian region of Nigeria which include Katsina, Nguru, Maiduguri, Kano and Sokoto. The annual rainfall series for each of the stations is utilized for a 50-year period (1951-2000). A composite of (Jun +Jul+Aug) rainfall is also derived for each station within the the same period (1951-2000). The soudano-sahelian stations considered receive the bulk of their rainfall during these months in any given year. The annual and composites rainfall series for these stations comprised the actual values of rainfall (i.e. non-normalized). Both series are then subjected to step wise multiple regression and multiple correlation analysis in an attempt to evaluate the nature of the relationship between rainfall in the soudano-sahelian region of Nigeria and indices of Sea Surface Temperature (SSTs) and El-Nino/Southern Oscillation (ENSO).

Some of the empirical researches that attempt to examine the influence of Sea Surface Temperature (SSTs) on rainfall in West African Sahel and elsewhere include Parker *et al* (1987, 1988), Folland *et al* (1986), Semezzi *et al* (1988), Adedoyin (1989a, b, c), Walker (1987, 1990), Ward (1992), Nichloson, 2000), Le Barbe et al (2000), Mo et al (2001) and Paeth and Henseen 92004). A limited analysis of atmospheric conditions that accompanied dry and wet Sahel years by Newell and Kidson (1984) also showed that the reason for drier years might be due to weaker moisture-flux convergence into the West African sector of the Sahel. They found that the lower tropospheric south-westerly winds were shallower and weaker in dry years. The persistence of drought in the Sahel over the last 20 years could therefore be understood easily in terms of the warmer-than-normal SST in the South Atlantic (Hastentrath, 1984; Lough, 1986) during this period.

Table 5.23 provides a summary of the regression of annual rainfall in the soudano-sahelian region of Nigeria on indices of SSTs and ENSO. It could be observed from table 5.23 that the most critical variable that dominates the inter-annual variability of rainfall at Katsina is NINO 3.4 SSTs which accounts for 44.2% of the explained variance in inter-annual variability of rainfall at Katsina. This supports the hypothesis that any slight change in sea surface temperature over this region is capable of influencing weather across the globe as put forward by Walker (1923), Walker and Bliss (1930) and Popelewski and Halpert (1987, 1989). The North Atlantic SSTs accounts for 35.9% of the explained variance in inter-annual variability of rainfall at Katsina. This is followed by the Global SSTs which accounts for 19.8% of the explained variance of the inter-annual variability of rainfall at Katsina. Table 5.23 further shows that the Global SSTs exercises greater control on inter-annual variability of rainfall at Nguru and accounts for 33.9% of the explained variance in inter-annual variability of rainfall at Nguru. This is followed by South Atlantic SSTs which accounts for 27.4% of the explained variance in inter-annual rainfall variability at Nguru.

It could also be observed from table 5.23 that the most critical variable that control annual rainfall variability at Maiduguri is North Atlantic SSTs and accounts for 28.8% of the explained variance. This is followed by Global SSTs which accounts for only 12.9% of the explained variance in annual rainfall variability. It could also be observed that the only critical variable that modulates annual rainfall variability at Kano is North Atlantic SSTs which accounts for 20.7% of the explained variance in inter-annual variability in rainfall at Kano. However, at Sokoto, South Atlantic SSTs accounts for only 16.0% of the explained variance in inter-annual variability of rainfall at Sokoto.This implies that extra ordinary warming of sea surface temperature over the Atlantic and Pacific oceans is significantly related to the observed inter-annual variability of rainfall in soudano-sahelian region of Nigeria. Such extra ordinary warming was found to significantly reduce the meridional gradient of SST south of the Inter-Tropical Discontinuity or (ITD), and as result, leads to a weakened Hadley meridional circulation. The weakened circulation reduces the intensity of the south-west monsoon flow into West and central Africa and consequently reduce rainfall in the region as demonstrated by Wolter (1989), Adedoyin (1989a), Nicholson (1981) and Folland *et al*. (1986).

Table 5.23. Summary of the regression of annual rainfall on SSTs and ENSO indices at the selected Soudano-Sahelian stations of Nigeria

Station	Critical Variable (s)	R	r^2	Significance
Katsina	Global SST	0.45	0.198	0.00
	North Atlantic SST	0.60	0.359	0.00
	NINO 3.4 SST	0.67	0.442	0.00
Nguru	South Atlantic SST	0.52	0.274	0.00
	Global SST	0.61	0.366	0.00
Maiduguri	Global SST	0.36	0.129	0.01
	North Atlantic SST	0.54	0.288	0.00
Kano	North Atlantic SST	0.46	0.207	0.00
Sokoto	South Atlantic SST	0.40	0.160	0.00

The result of the multiple correltion analysis shows that the annual rainfall at Katsina significantly correlates with only South Atlantic SSTs (-0.32*) and Global SSTs (-0.45**), though, the correlation is negative (see table 5.24). Similarly, the annual rainfall at Nguru significantly correlates with only South Atlantic SSTs (-0.52**) and Global SSTs (-0.47**), though, the correlations is also negative (see Table 5.25). It is interesting to note that the annual rainfall at Maiduguri significantly correlates with four out of eight independent variables considered in this work. It significantly correlates with South Atlantic SSTs (-0.33*), Global SSTs (-0.36*), NINO 3 SSTs (-0.32*) and NINO 3.4 SSTs (-0.30*). (see Table 5.26). At Kano, the annual rainfall significantly correlates with only North Atlantic SSTs (Table 5.27) while at Sokoto, annual rainfall significantly correlates with only South Atlantic SSTs (see Table 5.28). This result implies that the aberration in sea surface temperature over the tropical oceans (Atlantic and Pacific oceans), though a remote factor, could significantly influence the rainfall pattern in West Africa, Nigeria inclusive, by altering the tropical circulation (Hadley circulation). This leads to weakened meridional circulation which as a result reduce the intensity of the south-west monsoon flow into West and Central Africa and consequently reduce rainfall in the region as reported by Folland *et al* (1986), Adedoyin (1989a) and Mason (1995).

Table 5.24. Correlation coefficients between Katsina annual rainfall and indices of SSTs and ENSO

X1	1.00								
X2	0.14	1.00							
X3	-0.32*	0.03	1.00						
X4	-0.45**	0.49**	0.37**	1.00					
X5	-0.27	0.28	-0.05	0.72**	1.00				
X6	-0.27	0.26	-0.09	0.77**	0.92**	1.00			
X7	-0.24	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	-0.21	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	0.18	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Katsina Annual Rainfall

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.25. Correlation coefficients between Nguru annual rainfall and indices of SSTs and ENSO

X1	1.00								
X2	-0.01	1.00							
X3	-0.52**	0.03	1.00						
X4	-0.47**	0.49**	0.37**	1.00					
X5	-0.19	0.28	-0.05	0.72**	1.00				
X6	-0.18	0.26	-0.09	0.77**	0.92**	1.00			
X7	-0.25	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	-0.15	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	0.05	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Nguru Annual Rainfall

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.26. Correlation coefficients between Maiduguri annual rainfall and indices of SSTs and ENSO

X1	1.00								
X2	0.18	1.00							
X3	-0.33*	0.03	1.00						
X4	-0.68*	0.49**	0.37**	1.00					
X5	-0.27	0.28	-0.05	0.72**	1.00				
X6	-0.32*	0.26	-0.09	0.77**	0.92**	1.00			
X7	-0.25	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	-0.29*	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	0.26	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Maiduguri Annual Rainfall

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.27. Correlation coefficients between Kano annual rainfall and indices of SSTs and ENSO

X1	1.00								
X2	0.46*	1.00							
X3	-0.01	0.03	1.00						
X4	0.27	0.49**	0.37**	1.00					
X5	0.17	0.28	-0.05	0.72**	1.00				
X6	0.15	0.26	-0.09	0.77**	0.92**	1.00			
X7	0.13	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	0.13	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	-0.00	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Kano Annual Rainfall

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.28. Correlation coefficients between Sokoto annual rainfall and indices of SSTs and ENSO

X1	1.00								
X2	0.19	1.00							
X3	-0.40**	0.03	1.00						
X4	-0.08	0.49**	0.37**	1.00					
X5	0.08	0.28	-0.05	0.72**	1.00				
X6	0.08	0.26	-0.09	0.77**	0.92**	1.00			
X7	0.02	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	0.09	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	-0.10	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Sokoto Annual Rainfall

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.29 provides a summary of the result of regression of wet season rainfall of selected stations in the soudano-sahelian region of Nigeria on indices of SSTs and ENSO. It could be observed from table 5.29 that sea surface temperature variations in the East-Central Pacific region (NINO 3.4 SSTs) accounts for 40.8% of the explained variance in intra-seasonal variability of rainfall at Katsina. This is followed by North Atlantic SSTs which accounts for 29.0% of the explained variance in intra-seasonal variability of rainfall in Katsina. Higher than normal sea surface temperature of the global oceans between 40^0 S and 60^0 N (Global SSTs), also accounts for 18.9% of the explained variance in intra-seasonal variability of rainfall at Katsina. Table 5.29 further shows that the extra ordinary warming of the global oceans between 40^0S and 60^0N (Global SSTs) accounts for 37.7% of the explained variance in intra-seasonal variability of rainfall at Nguru. This is followed by South Atlantic SSTs which accounts for 29.4% of the explained variance in intra-seasonal variability of rainfall at Nguru.

It could also be observed from table 5.29 that the coupled ocean-atmospheric phenomenon of El-Nino/Southern Oscillation (ENSO) accounts for 21.8% of the explained variance in intra-seasonal variability of rainfall at Maiduguri. This is followed by South Atlantic SSTs which accounts for 12.1% of the explained variance at Maiduguri. The table further shows that North Atlantic SSTs is only the independent variable recognised by the step wise multiple regression model carried out for Kano and accounts for only 20.7% of the explained variance in intra-seasonal variability of rainfall at Kano while at Sokoto, the South Atlantic SSTs is the only critical variable recognised and accounts for only 15.1% of the explained variance in intra-seasonal variability of rainfall at Sokoto.The results presented in table 5.29 further confirm the findings of other studies which maintained that the perturbation in sea surface temperature over the tropical oceans (Atlantic, Pacific and Indian ocean), rather the than El-Nino/Southern Oscillation (ENSO), appears to dominate the observed inter-annual and intra-seasonal variability of rainfall in the Sahel and elsewhere (see Folland *et al*., 1986; Semazzi *et al*., 1988; Adedoyin, 1989a; Mason, 1995).

Table 5.29. Summary of the regression of wet season rainfall on SSTs and ENSO indices at the selected Soudano-Sahelian stations of Nigeria

Station	Critical Variable (s)	r	r^2	Significance
Katsina	Global SST	0.44	0.189	0.00
	North Atlantic SST	0.54	0.290	0.00
	NINO 3.4 SST	0.64	0.408	0.00
Nguru	South Atlantic SST	0.54	0.294	0.00
	Global SST	0.61	0.377	0.00
Maiduguri	South Atlantic SST	0.35	0.121	0.01
	ENSO Index	0.47	0.218	0.00
Kano	North Atlantic SST	0.46	0.207	0.00
Sokoto	South Atlantic SST	0.39	0.151	0.01

Table 5.30 presents the result of the multiple correlation analysis between the wet season rainfall at Katsina and indices of SSTs and ENSO. It could be observed that Katsina wet season rainfall significantly correlates with only South Atlantic SSTs (-0.38**) and Global SSTs (-0.46**) while table 5.31 reveals that Nguru wet season rainfall significantly correlates with only South Atlantic SSTs (-0.54**) and Global SSTs (-.47**). It is quite surprising that the wet season rainfall at Maiduguri does not significantly correlate with any of the independent variables considered in this work (see table 5.32). At Kano, the wet season rainfall significantly correlates with only North Atlantic SSTs (0.48**) while at Sokoto, the wet season rainfall correlates with only South Atlnatic SSTs (-0.39**) (see Tables 5.33 and 5.34). The results presented in tables 5.30-5.34 further confirm the dominant role of sea surface temperature (SSTs) on rainfall in the Sahel and elsewhere as demonstrated by Folland *et al* (1986), Semazzi *et al* (1988), Adedoyin (1989a), Ward (1992) and Mason (1995).

Table 5.30. Correlation coefficients between Katsina wet season rainfall and indices of SSTs and ENSO

X1	1.00								
X2	0.06	1.00							
X3	-0.38**	0.03	1.00						
X4	-0.46**	0.49**	0.37**	1.00					
X5	-0.23	0.28	-0.05	0.72**	1.00				
X6	-0.26	0.26	-0.09	0.77**	0.92**	1.00			
X7	-0.28	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	-0.23	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	0.16	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Katsina Wet Season Rainfall

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.31. Correlation coefficients between Nguru wet season rainfall and indices of SSTs and ENSO

X1	1.00								
X2	-0.02	1.00							
X3	-0.54**	0.03	1.00						
X4	-0.47**	0.49**	0.37**	1.00					
X5	-0.22	0.28	-0.05	0.72**	1.00				
X6	-0.21	0.26	-0.09	0.77**	0.92**	1.00			
X7	-0.22	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	-0.15	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	0.12	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Nguru Wet Season Rainfall

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.32. Correlation coefficients between Maiduguri wet season rainfall and indices of SSTs and ENSO

X1	1.00								
X2	-0.09	1.00							
X3	0.10	0.03	1.00						
X4	0.27	0.49**	0.37**	1.00					
X5	0.11	0.28	-0.05	0.72**	1.00				
X6	0.10	0.26	-0.09	0.77**	0.92**	1.00			
X7	0.36	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	0.22	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	-0.16	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Maiduguri Wet Season Rainfall

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.33.Correlation coefficients between Kano wet season rainfall and indices of SSTs and ENSO

X1	1.00								
X2	0.48**	1.00							
X3	-0.02	0.03	1.00						
X4	0.26	0.49**	0.37**	1.00					
X5	0.16	0.28	-0.05	0.72**	1.00				
X6	0.14	0.26	-0.09	0.77**	0.92**	1.00			
X7	0.12	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	0.13	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	0.04	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Kano Wet Season Rainfall

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Table 5.34. Correlation coefficients between Sokoto wet season rainfall and indices of SSTs and ENSO

X1	1.00								
X2	0.14	1.00							
X3	-0.39**	0.03	1.00						
X4	-0.12	0.49**	0.37**	1.00					
X5	-0.06	0.28	-0.05	0.72**	1.00				
X6	-0.02	0.26	-0.09	0.77**	0.92**	1.00			
X7	0.02	0.29*	-0.08	0.76**	0.58**	0.80**	1.00		
X8	0.04	0.25	-0.14	0.75**	0.77**	0.94**	0.93**	1.00	
X9	-0.07	0.14	0.19	-0.40**	-0.45**	-0.64**	-0.63**	-0.68**	1.00
	X1	X2	X3	X4	X5	X6	X7	X8	X9

**Significant at 0.01% level of significance

*Significant at 0.05% level of significance

X1= Sokoto Wet Season Rainfall

X2= North Atlantic SST

X3= South Atlantic SST

X4= Global SST

X5= NINO 1+2 SST

X6= NINO 3 SST

X7= NINO 4 SST

X8= NINO 3.4 SST

X9= ENSO Index

Folland *et al* (1986) examine the influence of the worldwide sea temperatures on Sahel rainfall over the period 1901-1985 using the comprehensively quality-controlled Meteorological Office Historical Sea Surface Temperature data set (MOHSST) via GCA circulation model. The result shows that the persistent wet and dry periods in the Sahel region of Africa are strongly related to contrasting patterns of sea-surface temperature (SST) anomalies on a near-global scale. The anomalies include relative changes in SST between hemispheres, on timescales of years to ten of years, which are most pronounced in the Atlantic. Experiments with an 11-level global atmospheric general circulation model (AGCM) supports the idea that the worldwide SST anomalies modulate summer Sahel rainfall through changes in tropical atmospheric circulation (Lamb, 1978, 1986; Newell and Kidson, 1984; Hastenrath, 1984).

Using an Empirical Orthogonal Function analysis (EFOs), Semazzi *et al* (1988) investigate the relationship between sea surface temperature (SST) and rainfall index anomalies over the sub-Saharan Africa for a 15-year period, 1970-1984. The authors used the objectively analysed monthly mean SST data for the global oceans 40^0S and 60^0N, and annual mean rainfall indices for the Sahel and Soudan belts over North Africa. The result shows that the most dominant eigenmode, EOF1, is characterized by warming over the central eastern Pacific, cooling over the eastern mid-latitude Pacific and warming over the entire Atlantic and Indian Ocean basins. The second EOF for the Atlantic Ocean SST analysis shows a dipole (north-south see-saw) pattern. The third EOF for the Atlantic SST analysis has the same sign over the entire Atlantic basin. Global SST EOF2 and EOF3 correspond to Atlantic SST EOF3 and EOF2 respectively. The result further shows that the correlation between the sub-Saharan rainfall index, which mainly represents the summer season rainfall from June to September, and SST EOFs shows that EOF1 has statistically significant monthly correlations for the Sahel and the Soudan regions and that the warm El-Nino-like phases of SST EOF1 correspond to drought conditions. They further argue that the large-scale SST anomalies might be responsible for a significant component of the observed vacillation of sub-Saharan rainfall.

Mason (1995) examines the sea-surface temperature-South African rainfall associations from 1910-1989 using principal component analysis. The result shows that the El-Nino type behaviour around the coast of southern Africa in the Benguela and Agulhas systems correlates with its Pacific Ocean counterpart during January-

March. He argued that the El-Nino type behaviour appears to be of minimal climatic significance to South Africa, and in contrast, sea-surface temperatures off the east coast associated with the second principal component and enhanced both tropical and temperate synoptic systems bringing rainfall to the sub-continent and could thus significantly affect rainfall receipts throughout most of the year. He concludes that the negative association with the Southern Oscillation dampens the climatic significance of both phenomena and may be responsible for the occasional faltering of the association between the Southern Oscillation and rainfall over South Africa.

Several other studies confirm the relationship between rainfall and ENSO in parts of eastern and southern Africa (Lindesay *et al*., 1986; Rasmusson and Carpenter, 1983; Van Hardeen *et al*., 1988; Nicholson and Kim, 1997; Nichloson and Saleto, 2000) However, there is a disagreement about its influence in the Sahel and elsewhere (e.g. Wolter, 1989; Semazzi *et al*., 1988; Nicholson and Palao,1992).

Using harmonic analysis method, Popelewski and Halpert (1987) also investigate the global and regional scale precipitation patterns associated with the El-Nino/Southern Oscillation (ENSO). They used monthly precipitation time series from over 1700 stations for the period 1875-1982 and derived the monthly ranked precipitation composites over idealized 2-year ENSO episodes for all the stations that include data for at least five ENSOs. The result shows that, in addition to the Pacific Ocean basin where precipitation pattern could be directly related to the ENSO, several other regions, which show consistent ENSO-related precipitation, are indentified. Specifically, four regions in Australia, two regions each in North America, South America, the Indian subcontinent and Africa, and one region in Central America are all found to have coherent ENSO-related precipitation.

In a more regionally-focused study, Nicholson and Kim (1997) examine the relationship of the El-Nino/Southern Oscillation (ENSO) to African rainfall. They applied the harmonic analysis method utilized by Popelewski and Halpert (1987) to 90 regionally averaged rainfall time series for the period 1901-1990. The analysis is a composite of 20 ENSO episodes within this period. The result shows that, 15 multiregion sectors where ENSO appears to modulate rainfall are indentified. However, the strongest signals are in eastern equatorial and south-eastern Africa. A continental scale is also apparent. The authors argue that the magnitude, seasonal timing and duration, and consistency of the rainfall response to ENSO vary among the sectors and from episode to episode. Nicholson and Kim (1997) conclude that the onset of the ENSO signal in rainfall commences far to the south and propagates latitudinally northward; and for this reason, the equatorial regions are out-of- phase with the continental pattern. It could be conjectured here that the influence of ENSO in the Sahel and elsewhere appears to be less significant as put forward by Wolter (1989), Semazzi *et al* (1988) and Nicholson and Palao (1992).

It is likely that this ENSO phenomenon directly influences rainfall in Africa by altering tropical circulations. For instance, the Pacific El-Nino event alters the Walker circulation by causing a shift, to the east, of the Pacific ascending branch, and creating subsidence over some parts of Africa (WMO, 1985). Webster (1983) proposes a schematic of the Walker circulation along the equator during the El-Nino and La-Nina periods, which are identified as the two extremes of the Southern Oscillation. From the schematic (Fig. 5.1), the descending (ascending) branch of the Walker circulation over the sub-Saharan region during the El-Nino (La-Nina) period of the Southern Oscillation could provide a link between ENSO and the sub-Saharan rainfall through large-scale subsidence (convection). This implies that during El-Nino (referring to warm phases) and La-Nina (referring to cold phases) episodes, the dominant tropical centers of convective activity and rising motion shifted eastward (westward) and resulted to an altered configuration of the Walker circulation and enhanced (reduced) subsidence over Africa.

This space is for figure 5.1

In view of the strong El-Nino signal over much of the African continent, a pronounced influence of La-Nina might also be anticipated. However, the only continental-scale study of its influence on African rainfall is that of Popelewski and Halpert (1989), who found a consistent La-Nina signal in two sectors. An area of equatorial East Africa tends to experience low rainfall during the period from November of the La-Nina year to March of the following year. An area of southeastern Africa experiences positive rainfall anomalies during this same period, up to the April of the year following La-Nina. Kiladis and Diaz (1989) examine the composite differences between cold and warm events and show an influence in roughly the same areas, but the analysis method does not permit them to distinguish between the influence of La-Nina and El-Nino. As with the El-Nino signal (Batt, 1989; Rasmusson *et al*., 1990; Meehl, 1993; Nicholson and kim, 1997), La-Nina's influence on African rainfall appears to be confined to those events that produce strong sea surface temperature (SSTs) anomalies (in this case, a cold phase) in the tropical oceans (Atlantic, Pacific and Indian oceans).

The relationship between El-Nino and La-Nina events to rainfall at selected stations of the soudano-sahelian region of Nigeria is also demonstrated in figures 5.2-5.7 using the monthly rainfall and Southern Oscillation Index or (SOI). Figure 5.2 shows the monthly rainfall distribution pattern at Maiduguri in the year 1955 which is associated with sustained positive values of the Southern Oscillation Index or (SOI). It has been stressed that the sustained positive values of the Southern Oscillation Index (SOI) is an indication of the colder than normal sea surface temperature over the eastern tropical Pacific which is equivalent to Cold-phase of ENSO (La-Nina) (Ayoade, 1988, 2004). The La-Nina phenomenon tends to enhance precipitation as reported by Nicholson and Selato (2000). It could be observed that the monthly rainfall at Maiduguri is considerably higher during the year 1955 which could be linked to the influence of La-Nina on rainfall during the year in question. A similar situation is obtained at Sokoto in the year 1955. There are sustatined positive values of the Southern Oscillation Index or (SOI) in the year 1955 and monthly rainfall distribution at Sokoto during the year 1955 is considerably higher as shown in figure 5.3. The pattern observed in figure 5.3 could as well be linked to the influence of La-Nina on rainfall in Sokoto during the year in question. It is interesting to note that the year 1955 is one of the La-Nina years as documented in the literature (see Popelewski and Halpert, 1989; Kiladis and Diaz, 1989).

This space is for figure 5.2

This space is for figure 5.3

The influence of El-Nino (a warm –phase of ENSO) on rainfall in the soudano-sahelian region of Nigeria is also demonstrated in figures 5.4 and 5.5. Figure 5.4 shows that the year 1982 is associated with sustained negative values of the Southern Oscillation Index or (SOI) and monthly rainfall is considerably low at Maiduguri during the year 1982, which also happens to be an ENSO year. The sustained negative values of Southern Oscillation Index or (SOI) is an indication of the warmer than normal sea surface temperature over eastern and central tropical Pacific which is associated with the warm-phase of ENSO. It has been pointed out that the sustained negative values of Southern Oscillation Index or (SOI) which represents El-Nino tends to suppress rainfall activity over the tropical areas and cause anomalously low rainfall (drought condition) in the affected areas (see Rasmusson and Carpenter, 1983; Popelewski and Halpert, 1987, 1989; Nicholson and Kim, 1997).

Like in Maiduguri, the monthly rainfall distribution at Nguru in 1972 is also considerably low. It is interesting to note that the year 1972 also happens to be an ENSO year and is characterized by the sustained negative values of the Southern Oscillation Index (SOI) as shown in figure 5.5. The rainfall distribution at Nguru shows evidence of anomalously low rainfall in 1972 (a drought year) and could be linked to the rainfall-suppression effect of El-Nino during the period.

This space is for figure 5.4

This space is for figure 5.5

A survey of the existing literature shows that, the year 1957 is an ENSO year and is characterized by the sustained negative values of the Southern Oscillation Index or (SOI) (see Trenberth, 1991; Zhang, 1993; Meehl, 1993), and the affected areas experienced pronounced negative rainfall anomalies during the period. It should however be noted that not all ENSO events are associated with negative rainfall anomalies. Few available studies carriedout in South Africa show that the ENSO events tend to explain not more than thirty percent of the observed anomalies in South African rainfall (see for example Willem and Mason, 1999; Reason and Roualt, 2002).

Figure 5.6 shows the monthly rainfall and Southern Oscillation Index or (SOI) distribution pattern at Kano during an ENSO year of 1957. It could be observed from figure 5.6 that the year 1957 is characterized by the sustained negative values of SOI from February through December which is a clear indication of El-Nino event (i.e.a warm-phase of ENSO). Such an El-Nino event is usually associated with dry condition due to rainfall-supression effect of El-Nino. However, the monthly rainfall distribution at Kano is considerably high in 1957 in spite of being an ENSO year. A similar situation is observed at Sokoto in 1957 (see Fig. 5.7). There are sustained negative values of Southern Oscillation Index or (SOI) in the year 1957 which indicate the occurrence of El-Nino (a warm-phase of ENSO). However, a visual examination of figure 5.7 shows that the monthly rainfall distribution pattern at Sokoto during the year in question is considerably high. This suggests that the El-Nino effects of 1957 were not significantly felt at Kano and Sokoto stations as evidenced from the monthly rainfall distribution of the said stations during the year 1957. This supports the idea that not all ENSO events are associated negative rainfall anomalies.

This space is for figure 5.6

This space is for figure 5.7

CHAPTER SIX

INTER-TROPICAL DISCONTINUITY AND RAINFALL VARIATIONS IN NIGERIA

The role of ITD in inter-annual variability of rainfall is somewhat controversial, but there is agreement that both its intensity and location are important (Nichloson, 1981; Ayoade, 1983; Newell and Kidson, 1984; Citeau *et al*., 1989; Hastenrath, 1990). The association between ITD and rainfall variability is generally assumed to involve the humid air lying to the south of it. However, attempts to link the moisture flux to inter-annual variability have been inconclusive (Lamb, 1983; Newell and Kidson, 1984; Fontaine and Janicot, 1992; Long *et al*., 1998, 2000). In a recent study, Nicholson and Grist (2001) suggest that the reason for the disagreement among the various studies is that the ITD and the westerly 'monsoon' flow south of it control mainly the low-level moisture. However, the moisture at higher levels seems to be more important than low-level moisture in the development of the rainy season (see for example Miller and Lindzen, 1992) and this appears to be independent of the moist, 'monsoon' flow. In Nigeria, however, the ITD assumes prominence, though independent of the rain producing systems, it has been demonstrated that the prominent precipitation band moves with the inter-tropical discontinuity or (ITD) in its south-north annual course (Adedokun, 1978; Olaniran, 1991a,; Ayoade, 2004).

In the present study, attempts are made to evaluate the influence of ITD on the distribution patterns of rainfall in Nigeria using simple linear regression analysis. Among the several meteorological parameters that could be used to determine the surface position of the Inter-tropical Discontinuity or (ITD) include the surface winds characteristics, the surface difference in dew point or vapor pressure between the northern and southern air streams. In Nigeria, as in most of West Africa, the Inter-tropical Discontinuity (ITD) is located by the surface difference in dew point or vapor pressure between the northern and southern air streams. The dew point in the northern air is very much below 15^0 C while that in the southern air is usually over 15^0 C. The ITD is therefore frequently located at the surface by a threshold dew point of 15^0 C. The monthly dew point temperature data from twenty-seven selected Nigerian meteorological stations and five meteorological stations in Niger republic are employed in the present study to determine the mean monthly surface position of the Inter- Tropical Discontinuity or (ITD). The reason for using the dew point temperature data from meteorological stations in Niger republic is that, the ITD attains its northernmost position around Lat. 18-20^{0}N in August which is considerably beyond Nigerian border. The justification for using the dew point temperature in this study has been provided in chapter three of this work (see P.73).

Table 6.1 shows the mean monthly positions of the Intertropical Discontinuity (ITD) and mean monthly rainfall in Nigeria. It could be observed from table 6. 1 that, the ITD movements are very irregular. The northward advance of the Inter-tropical Discontinuity (ITD) is about 1.9^0 of latitude per month while the southward retreat is about 3.4^0 of latitude per month. In other words, the southward retreat of the ITD is about twice than the northward advance of the ITD. While the northward advance is about 164km a month, the southward retreat is almost twice as rapid. This fact accounts for the gradual advance of the rains over the country at the beginning of the wet season and their rather abrupt retreat towards its end as shown in figure 6.1. It could also be observed from table 6.1 that, the mean monthly rainfall increases from the month of February through August (ITD northward advance); and thereafter decreases from the month of September through January (Southward retreat). This further suggests an inherent association of rainfall pattern with space-time variation in the depth of the moist layer and in vertical motion, which in turn, depends on the motion of the Inter-tropical Discontinuity or (ITD).

The last two columns of table 6.1 show the observed mean monthly rainfall and the observed difference in mean monthly rainfall respectively following the north-south movements of the inter-tropical discontinuity or (ITD). In January, when the ITD attains its southern limit position of 5.2^{0}N, the observed mean monthly rainfall at the stations within the latitudinal band 4^{0}N - 18^{0}N is 6.2mm. The ITD begins its northward incursion in February with a mean monthly surface position of 7.4^{0}N and the observed mean monthly rainfall in February at the stations located within the latitudinal band 4^{0}N - 18^{0}N is 13.0mm, giving a difference of +6.8mm from the amount recorded in January. The mean position of ITD in March is 9.3^{0}N and the observed mean monthly rainfall at the stations located within the latitudinal band 4^{0}N - 18^{0}N is 34.5mm, giving a difference of +21.5mm from the amount recorded in February. The mean monthly rainfall recorded at the stations located within the latitudinal band 4^{0}N - 18^{0}N is 71.4mm when the ITD attains its mean position of 12^{0}N in April, giving a difference of +36.9mm from the amount recorded in March.

However, the ITD attains its mean monthly position of 14.5^0 N in May and the mean monthly rainfall recorded at the stations located within the latitudinal band 4^0N - 18^0N is 110mm, giving a difference of +38.6mm from the amount recorded in April. In June, the ITD attains a mean monthly position of 16.8^0N and a total of 159mm is recorded as a mean monthly rainfall at the stations located within the latitudinal band 4^0N - 18^0N, giving a difference of +49mm from the amount recorded in May. The mean monthly position of ITD in July is 17.5^0 N and the mean monthly rainfall recorded at the stations located within the latitudinal band 4^0N - 18^0N is 198mm. This gives a difference of +39mm from the amount recorded in June. The ITD attains its northernmost limit in August with a mean monthly position of 18.7^0N; and the mean monthly rainfall recorded at the stations located within the latitudinal band 4^0N - 18^0N is 207mm, giving a difference of +9mm from the amount recorded in July.

However, the ITD retreats southward in September and attained a mean monthly position of 17.2^0N. The mean monthly rainfall recorded in September at the stations located within the latitudinal band 4^0N - 18^0N is 117mm which gives a difference of -30mm from the amount recorded in August. The ITD continues to retreat southward and reaches a mean monthly position of 12.3^0N in October; and the recorded mean monthly rainfall at the stations located within the latitudinal band 4^0N - 18^0N is 103mm. This further gives a difference of -74mm from the amount recorded in September. The ITD further retreats southward and attains a mean monthly position of 9.3^0 N in November. The observed mean monthly rainfall in November at the stations located within the latitudinal band 4^0N - 18^0N is 23mm which further gives a difference of -80mm from the amount recorded in October. In December, the ITD attains a mean monthly position of 7.2^0N in the course of its southward retreat and a total of 7mm is recorded as the mean monthly rainfall at the stations located within the latitudinal band 4^0N - 18^0N. This gives a difference of -16mm from the amount recorded in November. The ITD attains its southernmost limit in January with a mean surface position of 5.2^0N. The mean monthly rainfall recorded in January at the stations located within the latitudinal band 4^0N - 18^0N is 6.2mm which gives a difference of –0.8mm from the amount recorded in December.

The results presennted above show that, the observed space-time variations in rainfall could be ascribed largely to the variations in the north-south movements of the inter-tropical discontinuity or (ITD). It could also be observed from table 6.1 that the mean monthly rainfall at selected stations within the latitudinal band 4^0N - 18^0N increases northward from February through August (northward advance of ITD) and decreases southward from September through January (southward retreat of the ITD). This, therefore, suggests a strong relationship between monthly rainfall distribution and ITD positions in Nigeria. Several scholars linked the observed inter-annual and intra-seasonal variability in rainfall to north-south movements of the ITD which is highly irregular (see among others Adejuwon *et al*, 1990; Ayoade, 1974; Olaniran, 1991).

It could be observed from table 6.1 that the highest mean monthly rainfall of 207mm is recorded in August, when the ITD reaches its northernmost position of 18.7^0 N. The northward advance of the ITD from February heralds the beginning of the rainy season in Nigeria. Table 6.1 also shows that the mean monthly rainfall increases from February through August following the northward incursion of the ITD. However, the southward retreat of the ITD from September through January leads to progressive decline in the mean monthly rainfall from September through January. The last column of table 6.1 shows the observed difference in the mean monthly rainfall in Nigeria. The anomalies in south-north movements of the ITD have been linked to the large-scale anomalies in rainfall over West Africa particularly in the 1950s, 1970s and 1980s. It has been stressed that the northward displacement of the ITD accounts for the above average rainfall observed in most of the stations over West Africa in the 1950s while the large-scale rainfall deficits experienced in the West African region in the 1970s and 1980s which led to the catastrophic droughts of 1968-73 and 1982-87 in the region has been ascribed to the restricted northward advance of the ITD (see Winstanley, 1973; Nichloson, 1981, 1988; Olaniran, 1991).

Several reasons have been provided for the large scale rainfall deficits observed in West Africa in years of Sahelian drought among which include the weakened intensity of south-west monsoon flow into West and Central Afriaca due to weakened meridional Hadley circulation largely ascribed to higher than normal Sea Surface Temperature over the tropical oceans (Pacific, Atlantic and Indian oceans) (see Adedoyin, 1989; Nicholson and Kim, 1997; Semezzi *et al*, 1988). An increase in the intensity of the sub-tropical high during the period might have also blocked the northward movement of the Intertropical Discontinuity or (ITD) in those Sahelian dry years (Stranz, 1978). Stranz presents the results of mean deviations of air pressure and shows that the anomalous rainfall of the period 1970-1972 over the Sahel is characterized by a higher than usual pressure over North Africa, where the subtropical highs are normally situated. The strong easterly wind which develops on the southern flank of the Saharan high in such circumstances picks up more dust into the atmosphere and thereby inhibits precipitation processes, as well as block the northward movement of the ITD (Stranz, 1978).

Table 6. 1. Relationship of ITD position and rainfall over Nigeria and part of Niger republic, 1961-2000.

Month	Mean monthly ITD position (Lat. ^{0}N)	ITD Latitudinal Difference	Observed mean monthly rainfall (mm)	Observed difference in mean monthly rainfall (mm)
January	**5.2^0 N	2.0	6.2	-0.8
February	*7.4^0 N	2.2	13.0	+6.8
March	*9.3^0 N	1.9	34.5	+21.5
April	*12^0 N	2.7	71.4	+36.9
May	*14.5^0 N	2.5	110.0	+38.6
June	*16.8^0 N	2.3	159.0	+49.0
July	*17.5^0 N	0.7	198.0	+39.0
August	*18.7^0 N	1.2	207.0	+9.0
September	**17.2^0 N	1.5	177.0	-30.0
October	**12.3^0 N	4.9	103.0	-74.0
November	**9.3^0 N	3.0	23.0	-80.0
December	**7.2^0 N	2.1	7.0	-16.0

*Northward advance of the ITD

**Southward retreat of the ITD

Table 6. 2 provides the mean monthly rainfall distribution patterns by latitude within Nigeria and part of Niger republic. It could be observed from the table that, with the exception of the few stations located south of 6^0N and 4^0N, the entire country is almost rainless in January. This further demonstrates the influence of ITD on rainfall distribution patterns in Nigeria and supports the idea that the ITD might, on occasion, be south of 6^0N west of the Niger river and south of 4^0N east of the river. The mean monthly rainfall recorded at the stations south of 6^0N and 4^0N could also be attributed to the coastal effects. The moist unstable air that crosses the coast is subjected to convergence where the aerodynamically smooth sea gives way to the rough land (Ayoade, 1974). This narrow zone of convergence triggers showers which then drift inland as they dissipate. Besides, the variations in the orientation of the coast relative to the rain bearing southern air are found to have rain-inducing effects. The zone of maximum coastal effect is found where the coastline is convex to the Gulf of Guinea. Where the coast is concave, there is diffluence of winds which reduce the frictionally induced convergence at the coast. Where the coast is convex, on the other hand, the sea breeze adds a confluence and intensifies the coastal convergence (Ayoade, 1974; Adefolalu, 1986).

From February through August (northward advance of the ITD), the mean monthly rainfall progressively decreases with increasing latitude. This suggests that the south-west monsoon flow decreases in thickness northwards from the Gulf of Guinea. It should, however, be noted that the moist south-westerly winds that bring rainfall into Nigeria from tropical Atlantic Ocean across the Gulf of Guinea could penetrate beyond Nigeria's border to as far as the southern fringes of the Sahara Desert near latitude 20^0N as shown in table 6.1.

A careful examination of mean monthly rainfall by latitude withim Nigeria and part of Niger Republic presented in table 6.2 further shows that, in the month of June, when the mean monthly position of ITD is 16.8^0N, the recorded mean monthly rainfall at the stations located on latitudinal band 8^0N (195mm) is higher than the one recorded at the stations located on latitudinal band 9^0N (165.5mm). Similarly, in August, when the mean monthly position of ITD is 18.7^0N, the mean monthly rainfall recorded at the stations located on latitudinal band 10^0N (283mm) is higher than the one recorded at the stations located on latitude 11^0N (234.7mm) (see Table 6.2 and Fig. 6.1). This suggests that the zone of maximum rainfall that lies 500-600 meters south of the discontinuity is located at stations around latitudinal band 8^0N in June and displaced northward to around latitudinal band 10^0N in August.

Equally, in September, when the mean position of ITD is about 17.2^0N, the mean monthly rainfall recorded at the stations located on latitudinal band 8^0N (228.7mm) is higher than the one recorded at the stations located on latitudinal band 9^0N (205.5mm) (see Table 6.2 and Fig. 6.1). This suggests the southward displacement of the ITD zone of maximum rainfall from latitudinal band 10^0N to latitudinal band 8^0N. This further confirms the existence of the single rainfall maximum (May-September) for the semi-arid northern part of Nigeria) and the double rainfall maximum for the humid southern Nigeria (March-July and August-October). Omotosho (1985) examines the separate contributions of the squalls, thunderstorms and ordinary monsoons to the total rainfall and their variations with latitude. He shows that the total and thumderstorm rainfall decreases with increasing latitude but monsoon precipitation decreases exponentially while line squall rainfall is at the maximum around 9^0 N. He further shows that rainfall from thunderstorms displays a single annual peak in July/August at stations north of about 8^0 N whereas line squall precipitation exhibits a double maximum for all stations south of 12^0 N.

Table 6. 2. Mean monthly rainfall (mm) by latitude within Nigeria and part of Niger republic, 1961-2000.

					Latitude							
Month	4	5	6	7	8	9	10	11	12	13	14	16	18
January	27.6	18.5	17.3	7.1	8.8	1.0	0.1	0.0	0.0	0.0	0.0	0.0	0.0
February	52.9	48.3	31.8	21.0	11.1	2.9	1.5	0.1	0.1	0.0	0.0	0.0	0.0
March	134.2	124.8	84.4	65.5	14.3	13.4	6.7	2.9	0.8	1.2	0.0	0.1	0.0
April	195.6	196.1	153	123.4	100.4	68.3	39.2	27.2	10.1	13.5	1.0	0.0	0.1
May	253.3	257.4	213.8	156.8	160.8	135.8	99	64.3	33.1	33.3	15.1	11.1	0.6
June	335	322.2	295.8	192.7	195	165.5	160	107.6	85.1	82.6	119.5	9.0	0.9
July	412.5	398	307.1	193.3	154.9	223	228	196.9	165.3	161.6	92.1	31.8	3.2
August	362.8	366.4	221.6	177.8	142.1	241.8	283	234.7	223.6	190.9	136.3	87.6	19.7
September	381.1	393.3	278.7	215.6	228.7	205.5	205	134	94.8	95.4	28.3	34.8	0.9
October	302.1	327.6	215.2	156.2	141.1	75.6	52.4	28.7	11.6	15	10.7	0.8	0.0
November	133.4	74.3	46.5	24.5	16.9	2.0	0.9	0.6	0.1	0.3	0.0	0.0	0.0
December	33.4	20.4	17.6	6.48	10.6	0.1	0.0	0.0	0.0	0.0	0.0	0.0	0.0

Source 1: Nigerian Meteorological Agency, Lagos. **Source 2:** National Meteorological Service, Niamey, Republic of Niger.

The mean monthly rainfall distribution pattern within Nigeria and part of Niger republic is expressed graphically as a function of latitude and presented in figure 6.1. It could be observed from figure 6.1 that the mean monthly rainfall decreases from the coast with increasing latitude northwards. This further demonstrates the influence of the Inter-tropical Discontinuity or (ITD) on distribution patterns of rainfall in Nigeria. The variations observed in mean monthly rainfall distribution by latitude within Nigeria and part of Niger republic and presented in table 6.2 and figure 6.1 might be explained in terms of the mode of advance and retreat of the rain-producing mechanisms. In particular, the ITD invades from the south at the beginning of the rainy season in a highly irregular manner, in a series of surges, stagnations and retreats. On the other hand, its net final retreat at the end of the rainy season is even more irregular (Oguntoyinbo and Richards, 1977; Olaniran and Sumner, 1989; Ayoade, 2004). This differential pattern of advance and retreat of the ITD, and possibly the enhancement of rainfall producing processes by the highlands have combined to determine the observed pattern of mean monthly rainfall distribution within Nigeria and part of Niger republic as shown in table 6.2 and figure 6.1. The annual march of rainfall amount and the onset, advance and retreat of rainy season therefore seems to depend primarily on the position and seasonal displacement of the discontinuity relative to the stations.

This space is for figure 6.1

This space is for figure 6.1

This space is for figure 6.1

This space is for figure 6.1

This space is for figure 6.1

This space is for figure 6.1

The influence of the Intertropical Discontinuity or (ITD) on rainfall distribution patterns in Nigeria is further examined using simple linear regression analysis between the independent variable (mean monthly position of the ITD) and the dependent variable (mean monthly rainfall) over the period 1961-2000. The result is then presented in a scatter graph. The scatter graph (Fig. 6.2) summarises the quasi-linear but positive relationship between mean monthly rainfall and mean monthly ITD positions in Nigeria. The result further shows that a large proportion of the variation in rainfall amounts is explained by the Inter tropical Discontinuity or (ITD) ($r^2 = 0.96$, 0.01 level of significance). A visual examination of the scatter graph shows that the scattered points are closely tied to the trend line which suggests a strong association between the Intertropical Discontinuity (ITD) and mean monthly rainfall distribution patterns in Nigeria. This further demonstrates the influence of the ITD on rainfall distribution in West Africa in general, and Nigeria in particular.

This space is for figure 6.2

CHAPTER SEVEN

TRENDS AND VARIABILITY IN RAINFALL AND RAIN DAYS IN NIGERIA

7.1. Mean annual rainfall distribution pattern in Nigeria.

The mean annual rainfall pattern in Nigeria is presented in figure 7.1. It shows that rainfall decreases with increasing distance from the coast. Heavy rainfall totals of more than 2995mm occurred in Calabar within the Niger Delta region and the area bordering the Cameroon Mountains in the southeast. Thereafter, rainfall declines rapidly towards the west, with Lagos area having a mean annual rainfall of 1812.8mm. The lower mean annual rainfall observed in the south-west which is in contrast to the southeast, has been attributed to the development of the so-called 'little dry season' or 'August break' (Ireland, 1962; Adefolalu, 1986; Archaempong, 1987; Olaniram, 1991). This phenomenon is believed to have occurred due to purely the development of the inversion layer within the south-west monsoon which inhibits convective activity during the period and hence a reduced rainfall in the region. Other factors might have contributed to the dryness of this period, such as upwelling of cold waters at the Gulf of Guinea coast or stabilization of the south-westerlies caused by the Southern Africa subtropical high pressure (See Ojo, 1977; Olaniran, 1991).

Other scholars (Hamilton and Archbold, 1945; Eldridge, 1957; Obasi, 1965) have ascribed the lower mean annual rainfall observed in the south-west compared to the southeast, to the diminishing intensity and frequency of the disturbance lines, one of the major rain-producing mechanisms in West Africa, as they travel from the east towards the west. Another plausible explanation for the occurrence of the lower mean annual rainfall in the south-west, as given by Ayoade (1974), is the difference in the alignments of the coastline in the south-western and south-eastern parts of the country. In the former, the coastline is concave while in the latter is convex. The convex slope triggers convective activity and favours the routing of storms towards the southeast.

It should be noted, however, that rainfall continued to decline towards the interior with the mean annual rainfall of about 2995mm in the Niger Delta to as low as 470mm in the far north, particularly in the Chad and Sokoto Rima basins. These areas are predominantly under the influence of the Harmattan weather (Zone A) for most of the year and given their locations far from the source of the moisture (Atlantic ocean), the intensity of the moist southwesterly airstream decreases with increasing distance from the coast which aggravates the rainfall situation in those areas. The presence of the Hausa highlands in those areas predisposed them to occasionally suffer the orographic effects. These areas sufferred from being in the rain-shadow of these highlands, the highest part of which, the Jos Plateau, protrudes into the overlying Harmattan air and triggers convectional overturning around its flanks. This made Jos Plateau to receive a mean annual rainfall of about 1340mm and becomes wetter than most parts of the southwest despite its being located farther away from the coast. It could also be observed from figure.7.1 that rainfall has relatively declined along the Niger-Benue troughs. This could also be ascribed to the local topographic effects on rainfall distribution patterns in the country.

This space is for figure 7.1

7.2. Mean annual rain days distribution pattern in Nigeria

From figure 7.2, it could be observed that the distribution pattern of the mean annual rain days in the country exhibits similar pattern to that of the mean annual rainfall distribution pattern. The number of rain days also decreases from the south to the north. The highest mean annual rain days in the country is 182 days recorded at Calabar in the Niger Delta region. The mean annual rain days decreases from the southeast towards the southwest. From the coast inwards, mean annual rain days is found to be decreasing considerably, with least amounts of 38 and 49 days recorded for Nguru and Sokoto respectively.

Comparing the map of the number of rainy days with the map of the number of rain events ***n***31 (1.e., cumulated over the 31 days of August) over West Africa, Le-Barbe *et al* (2002) show that over the Sahel the number of rain events is larger-approximately by a factor of 125% than the number of rainy days. Moving southward, the number of rain events becomes equal to the number of rainy days around latitude 9^0 N. Farther south, the number of rainy days becomes far larger than the number of rain events. This is well in agreement with the known climatology of the convective rain events of the region. They further note that in the Sahel in August, most of the rain events are associated with convective systems with a separation time that could be less than one day. On the coast, at that time of the year there are relatively few rain events of different nature and the probability of having a rain event spreads over two consecutive days is relatively large.

Generally, annual rain days have been decreasing in Nigeria, particularly in Sokoto Rima and Chad basins of the semi-arid northern part of the country. The implications of such a reduction in the annual rain days include reduced annual rainfall amounts and shorten growing season as observed by Olaniran (1988b), Olaniran and Sumner (1989b), Sivakumar (1992) and Bello and Oni (2000).

This space is for figure 7.2

7.3. Annual rainfall trends in Nigeria.

In the West African region, rainfall exhibits a high degree of spatial and temporal variability. It is influenced by small-scale and large-scale factors such as land-sea breeze, relief and sea surface contrast; and is proned to large observational errors. Hence, for any meaningful study on rainfall patterns, reliable estimates of the magnitude of trends are needed particularly to assess the severity of drought and changes in rainfall characteristics in the region. In Nigeria, several recent studies (Anyadike, 1993; Hess et al, 1995) estimate trends in annual rainfall amounts based on only the so-called Climatic Normal Periods (CNP) defined by the World Meteorological Organization as 1931-1960 and 1961-1990 for Sub-Saharan Africa (WMO, 1971). In the present study, trends in annual rainfall and rain days time-series data for the entire period of available data (1901-2000) are analyzed using both simple linear regression and Mann-Kendall *tau* statistic. Both techniques were used in this study to examine trends in climatic time-series data (annual rainfall and rain days) and their performances compared.

In linear regression analysis, the slope indicates the average rate of change in the climatic parameter over the time period. One advantage of this method is that, it is easy to apply to a large number of sites. A disadvantage is that it could fail to detect trends that are non-linear but still monotonic (generally in one direction). However, in the case of Mann-kendall test, it could be used to detect trends that are monotonic but not necessarily linear, but these only indicate the direction, and not the magnitude of trends.

Using the total annual rainfall of twenty-seven representative Nigerian meteorological stations with a varying lengths of records, the equations of the lines of best fit for the rainfall fluctuations are computed. The way in which the annual rainfall totals varies with time is shown in the table of the correlation coefficients (see Table 7.1). It is noteworthy that out of 27 stations studied, 12 (44.4%) of the stations reveal a statistically significant downward trend in total annual rainfall over their respective length of records (see Table 7.1 and Fig. 7.3). Besides, 7 (25.9%) of the stations also have a negative correlation coefficients, though not statistically significant. The apparent decline in annual rainfall at these seven stations might have been due to purely chance variations and could not be statistically proved.

It is interesting to note that the correlation coefficients at Ikom, Osogbo, Ondo, Ibadan, Warri and Benin, all located in the southern part of the country are positive which indicate a tendency for the total annual rainfall to increase over their respective length of records. However, it is only at Osogbo that the correlation coefficient is significant (0.87, at 0.05%). The correlation coefficients at the other 5 stations failed to be significant at either 0.05% or 0.01% probability level (see Table 7.1 and Fig. 7.3). In other words, the apparent upward trend in total annual rainfall in these stations might have been due to purely chance variations and therefore not statistically proved. Despite their location in the semi-arid northern part of the country, no discernible trend in the total annual rainfall at Kano and Yelwa could be detected over their respective length of records having recorded the correlation coefficients of 0.00 each.

It appears then that the significant downward trend in the total annual rainfall observed in the country is more pronounced in the northern part of the country where about 9 out of 12 stations that indicate a significant downward trend in annual rainfall are located. If such trends continued, it could lead to adverse hydroclimatic consequences as put forward by Ojo (1986), Oyebande (1990), Ayoade (1995) and Bello (1998).

Table 7.1. Correlation coefficients (r) and regression equations for the total annual rainfall in Nigeria

S/No.	Station	Length of record	*(r)*	Regression Equation
1	Bauchi	1908-2000	-0.42*	Y= 1121.91-1.3474t
2	Jos	1922-2000	-0.17**	Y= 1471.08-3.2746t
3	Kano	1905-2000	0.00	Y= 849.389-0.0037t
4	Katsina	1918-2000	-0.35**	Y= 819.182-4.6365t
5	Kaduna	1930-2000	-0.05	Y= 1313.8-2.247t
6	Nguru	1942-2000	-0.29**	Y= 602.445-4.4049t
7	Maiduguri	1915-2000	-0.54*	Y= 673.281-1.4050t
8	Sokoto	1916-2000	-0.76**	Y= 755.831-1.6847t
9	Zaria	1943-2000	-0.04	Y= 1108.89-2.0610t
10	Yelwa	1926-2000	0.00	Y= 997.823+0.1377t
11	Yola	1914-2000	-0.02	Y= 973.707-1.0947t
12	Makurdi	1927-2000	-0.11*	Y= 1444.02-3.7341t
13	Lokoja	1916-2000	-0.01	Y= 1201.99-0.2566t
14	Ikom	1972-2000	0.01	Y= 2191.75+2.9838t
15	Ilorin	1916-2000	-0.02	Y= 1312.94-1.3849t
16	Bida	1928-2000	-0.11**	Y= 1330.54-3.3585t
17	Minna	1916-2000	-0.12**	Y= 104.856-3.2229t
18	Osogbo	1935-2000	0.87*	Y= 1163.48 +3.9610t
19	Ondo	1906-2000	0.34	Y= 1476.70+2.2028t
20	Ibadan	1905-2000	0.01	Y= 1271.81+0.6496t
21	Lagos	1901-2000	-0.02	Y= 1846.07-0.6650t
22	Enugu	1916-2000	-0.49*	Y= 1948.10-3.8174t
23	Portharcourt	1902-2000	-0.40**	Y= 3321.06-12.888t
24	Calabar	1902-2000	-0.64**	Y= 3198.10-4.1438t
25	Warri	1908-2000	0.03	Y= 2757.05+0.6290t
26	Benin	1906-2000	0.02	Y= 2000.10+1.7965t
27	Gusau	1953-2000	-0.01	Y= 959.009-1.0741t

*Significant at 0.05%

**Significant at 0.01%

This space is for figure 7.3

This space is for figure 7.3

This space is for figure 7.3

The annual rainfall time-series data of twenty-seven selected synoptic stations in Nigeria were also subjected to the Mann-Kendall *t* test for trends. Several scholars have utilized Mann-Kendall *tau* statistics to study trends in total annual rainfall in Africa and elsewhere (see Jagannathan and Parthasarathy, 1973; Parthasarathy and Dhar, 1975; Aondover and Woo, 1998; Aondover and Tarhule-Lips, 2001). Table 7.2 shows the Mann-Kendall correlation coefficients for the total annual rainfall of twenty-seven representative Nigerian stations with varying lengths of records.

The results obtained from the Mann-Kendall trend analysis are similar to those obtained through simple linear regression analysis for the total annual rainfall in Nigeria. It could be observed from table 7.2 that a total of 19 (70.4%) stations indicate a tendency towards a decreasing total annual rainfall in Nigeria while a total of 8 (29.6) stations indicate a tendency towards an increasing total annual rainfall in Nigeria. However, it is only 13 stations that show a statistically significant downward trend in total annual rainfall over their respective lengths of records. The remaining 6 stations, though, have negative correlation coefficients; the trends are not statistically significant and hence might have occurred due to purely chance variation. It is also interesting to note that out of 8 stations that indicate a tendency towards an increasing total annual rainfall in Nigeria, it was only at Osogbo that the trend was statistically significant (see Table 7. 2). It is noteworthy that 11 out of 13 stations that show significant downward trend in total annual rainfall are all concentrated in the semi-arid northern part of Nigeria.

It could also be observed that the Kano and Yelwa stations that have not indicate any discernible trend in their total annual rainfall as revealed through the linear regression analysis, have shown a tendency towards an increasing total annual rainfall via the Mann-Kendall trend analysis. However, the correlation coefficients obtained for Kano and Yelwa stations through the Mann-Kendall trend analysis are not statistically significant and might have occurred due to purely chance variations. This suggestes that both the simple linear regression and Mann-Kendall *t* test are appropriate in studying trends in time series of precipitation.

Table 7.2. Mann-Kendall *t* test for trends in total annual rainfall in Nigeria

S/No.	Station	Length of rocord	*(t)*
1	Bauchi	1908-2000	-0.38*
2	Jos	1922-2000	-0.15**
3	Kano	1905-2000	0.02
4	Katsina	1918-2000	-0.32**
5	Kaduna	1930-2000	-0.03
6	Nguru	1942-2000	-0.27**
7	Maiduguri	1915-2000	-0.51*
8	Sokoto	1916-2000	-0.73**
9	Zaria	1943-2000	-0.02
10	Yelwa	1926-2000	0.01
11	Yola	1914-2000	-0.04
12	Makurdi	1927-2000	-0.09*
13	Lokoja	1916-2000	-0.01
14	Ikom	1972-2000	0.02
15	Ilorin	1916-2000	-0.01
16	Bida	1928-2000	-0.10**
17	Minna	1916-2000	-0.11**
18	Osogbo	1935-2000	0.79*
19	Ondo	1906-2000	0.29
20	Ibadan	1905-2000	0.02
21	Lagos	1901-2000	-0.01
22	Enugu	1916-2000	-0.46*
23	Portharcourt	1902-2000	-0.37**
24	Calabar	1902-2000	-0.59**
25	Warri	1908-2000	0.01
26	Benin	1906-2000	0.01
27	Gusau	1953-2000	-0.06**

*Significant at 0.05%

**Significant at 0.01%

7.4. Trends in total annual rain days in Nigeria.

The regression/correlation analysis is used to examine the trends in total annual rain days of twenty-seven representative Nigerian stations with varying lengths of records. The way in which the total annual rain days varies with time is shown in table 7.3. The result shows that it is only in 3 out of the 27 stations studied is there any tendency for the total annual rain days to increase over their respective length of records. However, with the exception of Calabar, the correlation coefficients for these stations failed to be significant at either 0.01% or 0.05% probability level (see Table 7.3). This implies that the apparent upward trend in the total annual rain days in these stations over their respective length of records might have been due purely to chance variations and could not be statistically proved. It should be noted however that, there is no any discernible trend in Ilorin, a northern station and Benin, a southern station, with each having a correlation coefficient of 0.00.

The correlation coefficients for the remaining 22 stations are negative which indicate a tendency towards a decrease in their total annual rain days over their respective length of records. However, the observed downward trend in total annual rain days is statistically significant at only 17 stations (see Table 7.3 and Fig. 7.4). Majority of these stations are located in the semi-arid northern part of the country. The correlation coefficients at Kaduna, Yola, Ikom, Enugu and Warri are negative but not significant. It appears then that, with the exception of these stations, there is a significant and pronounced downward trend in total annual rain days in most parts of Nigeria over the period 1901-2000. The agricultural implications of these trends include reduction in total annual rainfall and subsequently reduced length of the growing season. If such trends continued, it could lead to the initiation and persistence of meteorological drought as discussed by Watts (1983), White and Glantz (1987), Mortimore (1989), Oyebande (1990) and Shuckla (1995).

Table 7.3. Correlation coefficients (r) and regression equations for the total annual rain days in Nigeria

S/No.	Station	Length of Record	*(r)*	Regression Equation
1	Bauchi	1961-2000	-0.12*	Y= 81.2859-0.3912t
2	Jos	1922-2000	-0.27**	Y= 119.319-0.2905t
3	Kano	1916-2000	-0.21**	Y= 65.1263-0.1840t
4	Katsina	1922-2000	-0.28**	Y= 60.8386-0.2583t
5	Kaduna	1961-2000	-0.04	Y= 100.531-0.2344t
6	Nguru	1942-2000	-0.59**	Y= 51.7276-0.4327t
7	Maiduguri	1916-2000	-0.16**	Y= 58.3385-0.1964t
8	Sokoto	1916-2000	-0.71**	Y= 53.6705-0.0935t
9	Zaria	1969-2000	0.01	Y= 77-1794+0.0819t
10	Yelwa	1943-2000	-0.20**	Y= 84.2851-0.3057t
11	Yola	1961-2000	-0.01	Y= 68.4068-0.1113t
12	Makurdi	1961-2000	-0.16**	Y= 83.3739-0.4379t
13	Lokoja	1916-2000	0.03	Y= 82.0502+0.0307t
14	Ikom	1973-2000	-0.07	Y= 165.603-0.4160t
15	Ilorin	1961-2000	0.00	Y= 88.6423+0.0077t
16	Bida	1961-2000	-0.27**	Y= 98.0688-0.5611t
17	Minna	1916-2000	-0.12**	Y= 104.856-3.2229t
18	Osogbo	1961-2000	-0.15**	Y= 126.673-0.4828t
19	Ondo	1961-2000	-0.23**	Y= 145.896-0.6974t
20	Ibadan	1961-2000	-0.11*	Y= 113.492-0.4984t
21	Lagos	1901-2000	-0.30**	Y= 141.537-0.8229t
22	Enugu	1953-2000	-0.05	Y= 119.099-0.2872t
23	Portharcourt	1948-2000	-0.21**	Y= 168.496-0.5866t
24	Calabar	1916-2000	0.06*	Y= 173.022+0.2189t
25	Warri	1915-2000	-0.01	Y= 172.999-0.0438t
26	Benin	1961-2000	0.00	Y= 142.554+0.0059t
27	Gusau	1942-2000	-0.10**	Y= 74.4617-0.2702t

*Significant at 0.05%

**Significant at 0.01%

This space is for figure 7.4

This space is for figure 7.4

This space is for figure 7.4

The total annual rain days time-series data of twenty seven selected synoptic stations in Nigeria were also subjected to the Mann-Kendall *t* test for trends. Several scholars have utilized Mann-Kendall *tau* statistics to study trends in total annual rainfall in Africa and elsewhere (see Jagannathan and Parthasarathy, 1973; Parthasarathy and Dhar, 1975; Aondover and Woo, 1998; Aondover and Tarhule-Lips, 2001). Table 7.2 shows the Mann-Kendall correlation coefficients for the total annual rain days of twenty-seven representative Nigerian meteorological stations with varying lengths of records. It is interesting to note that the results obtained through the simple linear regression analysis are almost similar to those obtained from the Mann-Kendall *tau* statistics for trends in total annual rain days in Nigeria.

It could be observed from table 7. 4 that, a total of 22 (81.4%) stations show a tendency towards decreasing total annual rain days in Nigeria over their respective lengths of records. However, it is only at 17 stations that the trends are statistically significant. The remaining 5 stations, though indicated a downward trend in their total annual rain days, the trends are not statistically significant and hence might have occurred due purely to chance variation. Similarly, a total of 5 stations indicate a tendency for an increasing total annual rain days over their respective length of records but none is statistically significant (see Table 7. 4). In other words, the observed trends at these stations might have occurred due purely to chance variation.

A comparison of the trends in total annual rain days in Nigeria as obtained from the simple linear regression and Mann-Kendall test analyses shows that Benin, a southern station, and Ilorin, a northern station, indicate no discernible trends in the total annual rain days through the linear regression analysis but an upward trend via the Mann-Kendall trend analysis, though not statistically significant. It should however be noted that in both the total annual rainfall and rain days trend analyses, the Mann-Kendall *tau* statistics tends to give lower estimates of the coefficients when compared with the correlation coefficients (*r*) obtained via the simple linear regression model. However both techniques are suitable for the analysis of trends in the time series of precipitation as shown in this work.

Table 7.4. Mann-Kendall *t* test for trends in total annual rain days in Nigeria

S/No.	Station	Length of Record	*(t)*
1	Bauchi	1961-2000	-0.10*
2	Jos	1922-2000	-0.24**
3	Kano	1916-2000	-0.19**
4	Katsina	1922-2000	-0.23**
5	Kaduna	1961-2000	-0.01
6	Nguru	1942-2000	-0.51**
7	Maiduguri	1916-2000	-0.15**
8	Sokoto	1916-2000	-0.65*
9	Zaria	1969-2000	0.01
10	Yelwa	1943-2000	-0.16**
11	Yola	1961-2000	-0.01
12	Makurdi	1961-2000	-0.14**
13	Lokoja	1916-2000	0.01
14	Ikom	1973-2000	-0.04
15	Ilorin	1961-2000	0.01
16	Bida	1961-2000	-0.17**
17	Minna	1916-2000	-0.10**
18	Osogbo	1961-2000	-0.12**
19	Ondo	1961-2000	-0.19**
20	Ibadan	1961-2000	-0.10*
21	Lagos	1901-2000	-0.23**
22	Enugu	1953-2000	-0.03
23	Portharcourt	1948-2000	-0.19**
24	Calabar	1916-2000	0.03
25	Warri	1915-2000	-0.01
26	Benin	1961-2000	0.01
27	Gusau	1942-2000	-0.08**

*Significant at 0.05%

**Significant at 0.01%

7.5. Inter-decadal variability of rainfall in Nigeria

The inter-annual variability in rainfall is an inherent feature of the wet –and-dry climatic environment of the tropics. Understanding the spatial-temporal pattern of rainfall fluctuations in any given area of the tropics requires a careful analysis of its rainfall variability. In this section, an attempt was made to elucidate the nature and pattern of rainfall variability in Nigeria using historical rainfall records of twenty-seven Nigerian meteorological stations over the period 1901-2000. The inter-decadal variability in rainfall over Nigeria is analyzed using the coefficient of variation. The results are presented in table 7.5. From table 7.5, it be could observed that the decade 1901-1910 records the highest value of average rainfall of about 2190.22mm followed by the decade 1911-1920 with an average rainfall of 1696.9mm. From 1920 onwards, there is a gradual decreasing decadal average rainfall, reaching the lowest value of about1250.7mm during the 1981-1990 decade.

The values of the coefficient of variation are relatively high in all the decades which reflect the characteristics of Nigerian rainfall of being highly variable. The decade 1981-1990 records the highest coefficient of variation of 52.1% and least value of 47.0% is recorded in the 1951-1960 decade. This implies that the rainfall is more variable in the decade 1981-1990 and less variable in the decade 1951-1960. With the exception of the 1951-1960 decade, the rainfall change between successive decades indicates a continuous decline in total annual rainfall in Nigeria from the decade 1911-1920 through the decade1941-1950. After the wet period of 1951-1960 decade, the successive decades are consistently drier except 1991-2000 decade which receives more rainfall than the 1981-1990 decade. In summary, the first four decades experienced above average rainfall of 77.36mm-762.13mm above the long-term mean while the remaining six decades recorded considerably below average rainfall of about 62.96mm-177.39mm below the long-term average. This shows that the decade 1981-1990 is the worst decade over the period 1901-2000.

The frequency of occurrence of anomalous rainfall during different decades is shown in the last two columns of table 7.5. The first four decades had not recorded a single rainfall failure. The decades 1941-1950, 1951-1960, 1961-1970, 1971-1980 and 1981-1990 recorded 2 years, 2 years, 1 year, 4 years and 5 years of rainfall failures respectively. Overall, the result of the analysis in table 7.5 shows that rainfall is decreasing in Nigeria at the rate of about 787mm per decade.

This space is for Table 7.5

Tables 7.6-7.9 present the results of the inter-decadal variability of rainfall for the North, Middle Belt (Central), South-West and South-Eastern regions of Nigeria. From table 7.6, it could be deduced that with the exception of the decade 1941-1950, the first four decades receive a decadal average rainfall values above the long-term mean of 801mm. Thereafter rainfall continuous to decline in the northern part of the country from 1961-1970 decade through the 1981-1990 decade with the 1981-1990 being the worst decade in history, with only a decadal average of 646.2mm. However, rainfall continue to appreciate slightly upward in the 1990s. The last two columns of table 7.6 represent the years of anomalous rainfall in the region. The decades 1941-1950, 1971-1980 and 1981-1990 are the driest decades in the northern part of Nigeria and have recorded 3, 3 and 8 years of rainfall failures respectively. Rainfall is found to be decreasing in the region at the rate of about 129mm per decade.

In table 7.7, it is revealed that with the exception of 1941-1950 decade, the decades 1911-1920 through the decade1961-1970, receive a decadal average rainfall above the long-term average of 1240.3mm. From the decade 1971-1980 through the decade 1991-2000, rainfall considerably declined in the middle belt (central) region of Nigeria. The decades 1941-1950, 1951-1960, 1961-1970, 1971-1980, 1981-1990 and 1991-2000 have recorded 1, 1, 2, 2 and 6 years of rainfall failure respectively. This shows that the decade 1981-1990 is also the worst decade in history of the region. The years of rainfall failures in the decades 1971-1980 and 1981-1990 coincide with the period of the great Sudano-Sahelian droughts of the 1970s and 1980s in West Africa as revealed in the literature (see Winstanley, 1973; Lamb, 1975; Landsberg, 1975; Nicholson, 1988). While rainfall decreased at the rate of 129mm per decade in semi-arid northern part of Nigeria, it was found to be decreasing at the rate of about 124mm per decade in the middle belt (central) region located in between the humid tropical climate of the south and semi-arid climate of the northern part of Nigeria.

The results of the inter-decadal variability analysis of rainfall presented in table 7.8 show a contrasting result to that of the north and the middle belt regions. In the western region of Nigeria, rainfall is found to be increasing at the rate of about 102mm per decade. This is evident from the correlation coefficients of the annual rainfall trends in Nigeria (Table 7.1), which show a positive correlation coefficients at most of the stations in the south-western region of Nigeria, though the positive correlation is significant only at Osogbo. The first decade 1901-1910 is a dry decade in the region with 3 years of rainfall failure. Rainfall appreciates slightly upward in the following two decades, alternating with two other decades of rainfall failure and again followed by another two decades of above average rainfall. Rainfall continued to decline in the region between 1971-1980 and 1981-1990. The decade 1991-2000 received a decadal average above the long-term mean of 1832mm. The decades 1931-1940, 1941-1950, 1951-1960, 1961-1970, 1971-1980 and 1981-1990 recorded 3, 3, 2, 1, 2 and 5 years of rainfall failures respectively. This shows that the 1981-1990 decade is also the worst decade in terms of rainfall variability in the region.

Contrary to what was obtained in the south-western region of Nigeria, an inter-decadal variability in rainfall over the south-eastern region of the country (Table 7.9) shows that with the exception of the 1919, there is no any single rainfall failure in the region from the decade 1901-1910 through the 1921-1930. When compared to what was obtained in the north and the middle belt, the difference in terms of regional variability of rainfall becomes clearer. For example, the decade 1931-1940 recorded 2 years and 3 years of rainfall failure for the south-east and south-west regions respectively with nil in both the north and the middle belt regions of Nigeria. Besides, the decade 1951-1960 is a wet decade in the northern part of the country but each of the three remaining regions had recorded at least 1 year of rainfall failure during the same decade. Though the decade 1901-1910 through the decade 1921-1930 could be described as the wet decades in the south-eastern region of the country, with each recorded a decadal average rainfall above the long-term mean of 2505.9mm; rainfall had considerably declined in the region from the decade 1931-1940 through the decade 1991-2000, with the decade 1981-1990 being the driest decade in the region over the period of the study. The results further show that rainfall is decreasing in the region at the rate of about 123mm per decade. This is evident from the correlation coefficients of the total annual rainfall in Nigeria (Table 7.1) which shows a significant downward trend in the total annual rainfall of most of the stations in the region particularly Enugu (-0.49, at 0.05%), PortHarcourt (-0.40, at 0.01%) and Calabar (-0.64, at 0.01%).

The results of the decadal variability in rainfall over Nigeria presented in tables 7.5-7.9 suggest the influence of the seasonal excursion of the Inter-Tropical Discontinuity (ITD) from the south to the northern part of the country, global sea surface temperature anomalies, large-scale atmospheric circulation and associated surface conditions on the regional variation in rainfall as extensively discussed in the literature (see Winstanley, 1973a, b; Tanaka *et al*, 1975; Palmer, 1986; Parker *et al*, 1987; Popelewiski and Halpert, 1987; Wolter, 1989; Sivakumar, 1992; Ward, 1992).

This space is for Table 7.6

This space is for Table 7.7

This space is for Table 7.8

This space is for Table 7.9

7.6. Inter-annual variability of rainfall in Nigeria.

Table 7.10 and figure 7.5 show the result of the analysis of inter-annual variability of rainfall over Nigeria. It could be deduced from table 7.5 that stations located north of latitude 9^0 N in Nigeria exhibit higher level of inter-annual variability in rainfall with most of them being located in the semi-arid northern part of the country. The Kano station has the highest value of the coefficient of variation of 43.6% followed by Nguru (30.3%), Katsina (27.1%) and Maiduguri (24.1%). This suggests that rainfall is highly variable (less reliable) in those areas and could significantly affect the rain-fed agriculture in the region if the variability in rainfall persists in the area. Umar (2010b) examines the recent trends and variability in the length of the growing season in northern Nigeria and shows that stations located north of latitude 9^0 N in Nigeria particularly Katsina, Nguru, Sokoto and Maiduguri, exhibit highest level of variability in the length of the growing season.

It is interesting to note that Jos and Kaduna are among the stations that record the least values of the coefficient of variation despite being located north of latitude 9^0 N. The stations with the least values of the coefficient of variation in the country include Warri (11.4%), Jos (13.1%), Kaduna (14.7%), Ikom (14.5%) and Calabar (15.3%). The pattern of the inter-annual variability of rainfall in Jos and Kaduna could also be linked to the relief-induced rainfall effect of Jos and Kaduna highlands. Other stations located south of latitude 9^0 N also recorded relatively lower values of the coefficient of variation. Lower values of coefficient of variation suggests that rainfall is less variable in the area. This supports rain-fed farming in the area since rainfall distribution characteristics are likely to be suitable for crop germination, establishment and full development.

It has been speculated that such variations in inter-annual rainfall might be explained in terms of the mode of advance and retreat of rain-producing mechanisms. Specifically, the Inter-tropical Discontinuity or (ITD) invades from the south at the beginning of rainy season in a highly irregular manner, in a series of surges, stagnations and retreats. On the other hand, its net final retreat at the end of the rainy season is far more irregular (see Adejokun, 1966; Ayoade, 1974; Oguntoyinbo and Richards, 1977; Olaniran, 1989, 1991). This differential pattern of advance and retreat of the ITD, and possibly the enhancement of rainfall producing processes by highlands, for example Jos and Kaduna highlands, have combined to determine the pattern of the inter-annual variability in rainfall observed in Nigeria (Olaniran, 1989, 1991)

Table 7.10. Inter-annual variability of rainfall over Nigeria

S/No.	Station	Length of record	Mean (mm)	S. D. (mm)	C.V. (%)
1	Bauchi	1908 – 2000	1066	172	16.1
2	Jos	1922 - 2000	1332	175	13.1
3	Kano	1905 – 2000	845	369	43.6
4	Katsina	1918 – 2000	641	174	27.1
5	Kaduna	1930 – 2000	1246	184	14.7
6	Nguru	1942 – 2000	475	144	30.3
7	Maiduguri	1915 – 2000	622	150	24.1
8	Sokoto	1916 – 2000	705	149	21.1
9	Zaria	1943 – 2000	1066	161	15.1
10	Yelwa	1926 – 2000	1002	190	18.9
11	Yola	1914 – 2000	926	186	20.1
12	Makurdi	1927 – 2000	1308	239	18.3
13	Lokoja	1916 – 2000	1191	233	19.5
14	Ikom	1972 – 2000	2237	325	14.5
15	Ilorin	1916 – 2000	1253	228	18.2
16	Bida	1928 – 2000	1206	216	17.9
17	Minna	1916 – 2000	1284	206	16.0
18	Osogbo	1935 – 2000	1278	223	17.4
19	Ondo	1906 – 2000	1582	329	20.8
20	Ibadan	1905 – 2000	1249	262	20.9
21	Lagos	1901 – 2000	1813	416	22.9
22	Enugu	1916 – 2000	1792	407	22.7
23	Porharcourt	1902 – 2000	2696	570	21.1
24	Calabar	1902 – 2000	2995	460	15.3
25	Warri	1908 – 2000	2787	317	11.4
26	Benin	1906 – 2000	2086	359	17.2
27	Gusau	1953 – 2000	933	182	19.5

This space is for figure 7.5

CHAPTER EIGHT

SUMMARY AND CONCLUSION

SUMMARY

The study examines the pattern of rainfall anomalies in Nigeria over a period 1901-2000 with emphasis on droughts, which have been a constant feature of the climate of semi-arid West Africa, and will in all indications, continue to plague the region in the future. The dominant influence of the coupled air-sea interaction phenomena of El-Nino/Southern Oscillation (ENSO) and Sea Surface Temperatures (SSTs) anomalies on rainfall and rain days anomalies on a regional and country-wide basis is also investigated. The south-north movements of the Inter-tropical Discontinuity (ITD) in relation to the observed space-time variations in rainfall distribution patterns in Nigeria is also examined.

The data used for this study include the monthly rainfall and rain days for twenty-seven meteorological stations in Nigeria with varying lengths of records ranging from 56 years to 100 years over the period 1901-2000. The monthly rainfall data of five selected meteorological stations in the Republic of Niger (Lat. 14^0 N-18^0 N) for the period 1961-2000 are also utilized for the study. The monthly surface dew point temperature data for twenty-seven selected meteorological stations in Nigeria (Lat. 4^0N-13^0 N) and five selected meteorological stations in the Republic of Niger (Lat. 14^0N-18^0 N) are also obtained and used for the study. Other sets of data include monthly Sea Surface Temperature (SST) anomalies and Southern Oscillation Index (SOI) obtained from the Climatic Prediction Centre (CPC).

A 5-year moving average is applied to the annual rainfall totals for each of the stations in order to discern the pattern of rainfall fluctuations and trends in the country. The result shows a decreasing trend in total annual rainfall in the country with increasing distance from the coast which suggests a diminishing influence of the rain bearing south-west monsoon northward. A regional comparison of the rainfall pattern in the country shows that the northern region receives the least amount of seasonal rainfall and is, therefore, the most drought-proned. The region shows a relatively long period of below normal rainfall in the 1940s, which was interrupted only by two years of relatively high rainfall in 1945 and 1946 follwed by another low rainfall period beginning in 1968 and lasting through 1987 which corresponds with the drought periods of 1940s, 1970s and 1980s in the West African region.

The corresponding rainfall series for the central region also shows similar tendencies towards below-normal rainfall during these two periods (1968 through 1987) but to a lesser degree. Examination of the rainfall records for the southern regions (south-west and south-east) shows no consistent year-to-year pattern in the rainfall series compared to the northern region. There are years within this drought period, however, which show below-normal rainfall throughout most of Nigeria (i.e., 1926-28, 1942, 1943, 1944, 1948, 1971, 1972, 1973, 1976, 1982, 1983, 1984, 1987 and 1988). The results suggest a strong association between abnormal rainfall and anomalous large-scale circulation pattern in this sub-tropical region during these years, especially since rainfall data for the 1971-1973 period indicate widespread drought conditions throughout the country.

The annual rainfall anomaly pattern for Nigeria shows a country-wide occurrence of above average rainfall in the 1950s which suggests a considerable northward incursion of the Intertropical Discontinuity or (ITD) in those years. There was, however, a country-wide occurrence of below average rainfall in 1972/73 and 1982-84 in the country which suggests a restricted northward advance of the Inter-tropical Discontinuity or (ITD) in those years. However, a comparison of annual rainfall anomaly pattern for the north, middle belt, south-west and south-east shows that, the wet period of the 1950s is observed in virtually all the regions except in the south-eastern region which experiences a pronounced drought throughout the 1950s. However, a relatively wet period is observed in the 1930s in all the regions except the central (Middle) region of Nigeria. The result also shows that, the first and second decades (1901-10 and 1911-20) were dry decades in the south-western region of Nigeria while the north, central and south-eastern regions experienced above average rainfall during the same period.

An examination of the spatial pattern of the annual rainfall departure from normal in Nigeria for the selected dry years (1972, 1973, 1982 and 1983) shows that with the exception of few stations, the entire country experiences pronounced rainfall deficits during those years which indicate dry climatic conditions during the period. The problem becomes more glaring in the year 1983 (a dry year) when all the twenty-seven selected meteorological stations in the country recorded a total annual rainfall far below their respective long-term means. Similarly, during the wet years (1954, 1955, 1957 and 1962), nearly all the stations in the country recorded annual rainfall totals far above their respective long-term means, which indicates wet climatic conditions during those years. The result further shows that in the year 1957, a wet year, all the twenty-seven selected meteorological stations in the country recorded total annual rainfall above their respective long-term means.

The Standardized Precipitation Index (SPI) and the Percentage Deviation Below the Mean (PDBM) index have been employed in this study for the analysis of meteorological droughts in Nigeria. The station by station analysis of annual rainfall totals using the SPI shows a high percentage frequency of occurrence of mild drought (SPI scale 0 to -0.99) in Nigeria which constitutes about 763(71.0%) of the total drought events detected using the SPI technique. This is followed by a moderate drought (SPI scale -1.00 to -1.49). The result of the SPI analysis further shows a relatively high percentage frequency of the occurrence of severe (-1.50 to -1.99) and extreme droughts (≤ -2.00) in Nigeria which constitute about 77(07%) and 28(03%) for severe and extreme droughts respectively.

However, the PDBM index shows a comparatively low percentage frequency of occurrence of meteorological droughts when applied to annual rainfall series of twenty seven selected synoptic stations in Nigeria. Specifically, a total of 436 slight droughts (11-25% deviation below the mean) were detected in Nigeria using the PDBM index which constitutes about 71.9% of the total drought events detected in Nigeria. In the case of moderate droughts (26-45% deviation below the mean), a total of 156 (25.6%) events are detected whereas a total of 16 severe droughts (46-60% deviation below the mean) were detected and constitute about 3.0% of the total drought events detected in Nigeria using the PDBM index. In the case of disastrous drought (more than 60.0% deviation below the mean), the index detected only one disastrous drought which occurred at Yola in 1973.

Furthermore, the maps of the spatial patterns of droughts in Nigeria show that the drought occurrence in Nigeria is largely sporadic in its spatial distribution but the semi-arid northern part of Nigeria records more droughts than its humid southern counterpart. The impact of drought is likely to be felt more by the drought-proned northern states than the southern states. The high percentage frequency of occurrence of droughts in the northern part of the country than in the southern part as detected by the PDBM index made this index more realistic than the so-called SPI index widely used in temperate regions of the world. This shows that the McKee *et al*.'s (1993) SPI index is defective in the sense that it could not objectively detect the regions that are 'drought prone' more than others in a wet and dry climatic environment as evident from meteorological drought analysis carried in this work. The SPI index simply considers all negative deviations from the mean as drought. This results in the occurrence of droughts in nearly fifty percent of the years of available rainfall records in most of the stations studied. This made the SPI index highly unrealistic when applied to annual rainfall series of Nigeria unlike the PDBM index which considers only more than 10% deviation below the mean as indication of drought.

It should however be noted that the high percentage frequency of occurrence of mild droughts in the SPI index and slight droughts, in the case of the PDBM index in all the stations considered does not pose any serious threat to rain-fed farming in the sense that the water shortage associated with this type of drought is quite insignificant. The recovery rate from this type of drought is quite fast if precipitation situation appreciates upwards slightly. It is interesting to note that, although, the rate of occurrence of severe and extreme/disastrous droughts is low in the country, such droughts are associated with extremely dry conditions and as such recovery could be very difficult. The resultant effects of drought of this magnitude include the loss of agricultural output, reduced biomass production and a great reduction in animal quality and quantity. These types of droughts are known to have caused outright migration and abandonment of farmlands. Indirectly, it might lead to change in land use practice and create heavy pressure on urban centers there by putting severe strain on the economy.

To understand and predict future occurrences or develop strategies to minimize and manage the effects of either flood or drought, information is required on the probability of receiving less than the least annual rainfall ever recorded which is often associated with drought. The probability of receiving more than the highest annual rainfall ever recorded in Nigeria is also computed using the normal distribution function. Such probabilities are associated with the occurrence of floods in Nigeria. The result shows that Kano and Gusau stations, all located in the north, have the highest probability of (12.30%) and (7.93%) respectively, of receiving less than the least annual rainfall ever recorded at those stations over their respective lengths of records (Table 4.10). This is followed by Ondo (5.82%), Nguru (4.27%) and Zaria (4.09%). This implies that Kano, Gusau, Ondo, Nguru and Zaria stations stand the risk of drought than all other stations in the country. It is interesting to note that Ondo, a southern station, also records a relatively high probability (5.82%) of receiving less than the least annual rainfall ever recorded over the period 1906-2000.

The results further show that show that Ikom, a southern station, records the highest probability (9.01%) of receiving more than the highest annual rainfall of 2674mm over its available length of record spanning 1972-2000. This suggests that Ikom stands the riks of flood than all other stations in the country. The results further show that the risk of floods in the country tends to be localized rather than the large-scale pattern which is often associated with drought.

The simple linear regression model and Mann-Kendall *tau* statistics are employed to examine the trends in annual total rainfall and rain days in Nigeria from twenty-seven selected Nigerian meteorological stations with varying lengths of records ranging from 56 years to 100 years over the period 1901-2000. The linear regression results are generally in agreement with those of Mann-Kendall; though the latter tends to give lower estimates of the coefficients than the former. Overall, the result shows a downward trend in both annual rainfall and rain days totals with increasing distance from the coast which suggests a diminishing influence of the rain bearing south-west monsoon northwards. It is noteworthy that out of 27 stations studied, 12 (44.4%) reveal a statistically significant downward trend in total annual rainfall over their respective length of records. Besides, 7 (25.9%) of the stations also have a negative correlation coefficients, though not statistically significant. The apparent decline in annual rainfall at these seven stations might have been due to purely chance variations and could not be statistically proved.

It is interesting to note that the correlation coefficients at Ikom, Osogbo, Ondo, Ibadan, Warri and Benin, all located in the southern part of the country, are positive which indicate a tendency for the total annual rainfall to increase over their respective lengths of records. However, it is only at Osogbo that the correlation coefficient is significant (0.87, at 0.05%). The correlation coefficients at the other 5 stations are not significant at either 0.05% or 0.01% probability level. The correlation coefficient at Kano and Yelwa is 0.00 each, which indicates no discernible trend in their annual rainfall totals within their respective length of records.

The trend in annual total rain days in Nigeria shows a progressive south-north decline in total annual rain days. A total of 22(81%) of the stations considered have negative correlation coefficients which indicate a tendency towards a decreasing annual rain days, with the correlation coefficients of 17 of the stations being statistically significant. Most of these stations are located in the semi-arid northern part of the country. However, the correlation coefficient at Benin, a southern station and Ilorin, a northern station, is 0.00 each, which indicates no discernible trend in their total annual rain days over their respective lengths of record. The possible agricultural implications of these trends include reduction in total annual rainfall and consequently reduced length of the growing season. If such trends continued, it could lead to the initiation and persistence of meteorological drought.

The analysis of inter-decadal variability of rainfall in Nigeria was carried out using the coefficient of variation (C.V). The result shows that total annual rainfall in Nigeria is decreasing at the rate of about 787mm per decade. A regional analysis of rainfall variability, however, shows that rainfall is decreasing at the rate of 129mm and 124mm per decade for the north and the middle belts respectively. However, in the case of south-western part of Nigeria, rainfall is found to be increasing at the rate of about 102mm per decade. In the south-eastern part of the country, the result of the inter-decadal variability of rainfall shows that rainfall is decreasing at the rate of about 123mm per decade. Overall, the result shows that rainfall is highly variable in the country, particularly in the semi-arid northern part of the country. This is supported by the result of the inter-annual variability of rainfall in Nigeria which also shows that rainfall is highly variable in the country.

The relationsqhip of the El-Nino/Southern Oscillation (ENSO) and Sea Surface Temperature (SST) anomalies to rainfall in Nigeria was investigated using stepwise multiple regression model and multiple correlation analysis. The result shows that higher than normal SST over the North Atlantic Ocean dominates the observed inter-annual rainfall anomalies in the Middle belt (Central) region of Nigeria and accounts for only 12.4% of the explained variance. A higher than normal SSTs over South Atlantic Ocean appears to dominate the observed inter-annual rainfall anomalies in the northern part of Nigeria, though, accounts for only 12.0% of the explained variance. However, no relationship could be established between annual rainfall anomalies in Nigeria and anomaly indices of SSTs and ENSO. Similar results were obtained for the South-West and South-Eastern regions of Nigeria which also show no significant relationship between the anomaly indices of SSTs and ENSO and annual rainfall anomalies for both regions.

The relationship of El-Nino/Southern Oscillation (ENSO) and Sea surface Temperatures (SSTs) anomalies to Nigerian rainfall anomalies is also examined using the wet season rainfall anomalies for Nigeria and for each of the four regions considered in this study using stepwise multiple regression model. The result shows that South Atlantic SSTs anomalies dominate the intra-seasonal rainfall anomalies in the northern part of Nigeria and accounts for 10.8% of the explained variance while North Atlantic SSTs anomalies control the intra-seasonal rainfall anomalies in the Central (Middle Belt) region of Nigeria and accounts for 10.2% of the explained variance. In the case of South-West, the South Atlantic SSTs anomalies appears to be the most critical variable that controls the intra-seasonal rainfall anomalies in the region. However, no relationship could be established between wet season rainfall anomalies in Nigeria and anomaly indices of SSTs and ENSO. Similar results were obtained for the South-Eastern region of Nigeria as none of the predictors appears to significantly correlate with wet season rainfall anomalies for the region.

Due to the weak relationship observed between rainfall anomalies in Nigeria and anomaly indices of SSTs and ENSO, the relationship is re-examined using the unstandardized rainfall series of Nigeria and indices of SSTs and ENSO for the period 1951-2000, using multiple correlation analysis. The result further shows that the annual rainfall series for Nigeria does not significantly correlate with any of the predictors considered in this study. Similarly, the annual rainfall series for the north, central, south-west and south-eastern regions of Nigeria does not significantly correlate with any of the set of independent variables considered in this study. The unstandardized wet season rainfall series for Nigeria correlates with the actual indices of SSTs and ENSO. The result reveals that the wet season rainfall series in Nigeria does not significantly correlate with any of the independent variables considered in this work. However, the wet season rainfall series for the north, central and south-western regions of Nigeria does significantly correlate with only South Atlantic SSTs, having a correlation coefficient of (-0.35*), (0.32*) and (0.28*) for the north, central and south-western regions respectively. For the South-Eastern region of Nigeria, the wet season rainfall series significantly correlate with only North Atlantic SSTs (0.29*).

The relationship of El-Nino/Southern Oscillation (ENSO) and Sea Surface Temperatures (SSTs) to annual rainfall series at selected stations in the soudano-sahelian region of Nigeria was also investigated using stepwise multiple regression model and multiple correlation analysis for the period 1951-2000. Five stations located in the Sahelian region of Nigeria were selected for the analysis and they include Katsina, Nguru, Maiduguri, Kano and Sokoto. The result shows that the most critical variable that dominates the inter-annual variability of rainfall at Katsina is the

East Central Pacific SSTs (NINO3.4 SSTs) and accounts for 44.2% of the explained variance. This is followed by the North Atlantic SSTs which accounts for 35.9% of the explained variance. Sea Surface Temperatures over the global oceans (Global SSTs) accounts for 19.8% of the explained variance in inter-annual variability of rainfall at Katsina.

At Nguru, the result shows that Global SSTs accounts for 36.6% of the explained variance in inter-annual variability of rainfall at Nguru. This is followed by South Atlantic SSTs which accounts for 27.4% of the explained variance. The North Atlantic SSTs appears to dominate the inter-annual variability of rainfall at Maiduguri and accounts for 28.8% of the explained variance. This is followed by the Global SSTs which accounts for 12.9% of the explained variance. The result further reveals that the North Atlantic SSTs is the only critical variable that controls the inter-annual variability of rainfall at Kano and accounts for only 20.7% of the explained variance whereas South Atlantic SSTs appears to be the only critical variable that modulates the inter-annual variability of rainfall at Sokoto, though, accounts for only 16.0% of the explained variance.

An examination of the relationship between wet season rainfall series for selected soudano-sahelian stations in Nigeria and indices of SSTs and ENSO further reveals an interesting result. The result shows that the East-Central Pacific SSTs (NINO 3.4 SSTs) accounts for 40.8% of the explained variance in intra-seasonal variability of rainfall at Katsina. This is followed by the North Atlantic SSTs which accounts for 29.05% of the explained variance. The Global SSTs accounts for 18.9% of the explained variance in intra-seasonal variability of rainfall at Katsina. For Nguru, the Global SSTs accounts for 37.7% of the explained variance in intra-seasonal variability of rainfall. This is followed by the South Atlantic SSTs which accounts for 29.4% of the explained variance. At Maiduguri, the coupled ocean-atmospheric phenomenon of El-Nino/Southern Oscillation (ENSO) appears to dominate the intra-seasonal variability of rainfall, though, accounted for only 21.8% of the explained variance. This is followed by South Atlantic SSTs which accounts for 12.1% of explained variance. The result further shows that North Atlantic SSTs is the only critical variable that dominates the intra-seasonal variability of rainfall at Kano, and accounts for only 20.7% of the explained variance while at Sokoto; the intra-seasonal variability of rainfall is controlled by the South Atlantic SSTs, though, accounts for only 15.1% of the explained variance. Such extra ordinary warming over the South Atlantic Ocean is found to significantly reduce the meridional gradient of SST south of the Inter-Tropical Discontinuity (ITD), and as result, leads to a weakened Hadley meridional circulation. The weakened circulation reduce the intensity of the south-west monsoon flow into West and central Africa, and consequently reduce rainfall in the region. These findings have further confirmed the results of other studies on the influence of large-scale sea surface temperatures (SSTs) on rainfall in the Sahel and elsewhere.

The influence of the Intertropical Discontinuity or (ITD) on the distribution patterns of rainfall in Nigeria is examined using a linear regression analysis. The result shows that a large proportion of the variation in rainfall amounts is explained by the Inter tropical Discontinuity (ITD) (r^2 = 0.959, 0.01 level of significance). The mean monthly surface positions of the Intertropical Discontinuity (ITD) are determined using the mean monthly surface dew point temperature data from the twenty-seven selected Nigerian meteorologfical stations (Lat. 4^0 N - 13^0 N) and five selected meteorological stations from the Republic of Niger (Lat. 14^0 N - 18^0). From table 7.2, it could also be deduced that the space-time variation in the rainfall distribution patterns in the country is a function of the seasonal migration of the Intertropical Discontinuity (ITD) which is highly irregular. The Intertropical Discontinuity (ITD) travels at the rate of about 1.9^0 of latitude per month during its northward advance from February through August and reaches its northernmost position of about 18.7^0 N in the month of August.

From February through August (northward advance of the ITD), the mean monthly rainfall progressively decreases with increasing latitude. This suggests that the south-west monsoon flow decreases in thickness northwards from the Gulf of Guinea. It should, however, be noted that the moist south-westerly winds that bring rainfall into Nigeria from tropical Atlantic Ocean across the Gulf of Guinea could penetrate beyond Nigeria as far as the southern fringes of the Sahara Desert near latitude 20^0 N.

The result further shows that, in the month of June, when the mean monthly position of ITD is 16.8^0 N, the recorded mean monthly rainfall at the stations located on latitudinal band 8^0 N (195mm) is higher than the one recorded at the stations located on latitudinal band 9^0 N (165mm). Similarly, in August, when the mean monthly position of ITD is 18.7^0 N, the mean monthly rainfall recorded at the stations located on latitudinal band 10^0 N (283mm) is higher than the one recorded at the stations located on latitudinal band 11^0 N (234mm). This suggests that the zone of maximum rainfall that lies 500-600 meters south of the discontinuity is located at stations located on latitudinal band 8^0 N in June and then displaced northward to around latitudinal band 10^0 N in August. Similarly, in September, when the mean position of ITD is 17.2^0 N, the mean monthly rainfall recorded at the stations located on latitudinal band 8^0 N (228.7mm) is higher than the one recorded at the stations located on latitudinal band 9^0 N (205.5). This suggests the southward displacement of the ITD zone of maximum rainfall from latitudinal band 10^0 N to latitudinal band 8^0 N.

CONCLUSION

The pattern of rainfall anomalies in Nigeria was examined in this study with emphasis on droughts. The factors responsible for the observed pattern of rainfall anomalies in Nigeria were indentified and discussed.The total annual rainfall series in the country reveals a decreasing trend with increasing distance from the coast. The annual rain days series also shows a similar pattern. The annual rainfall series in the country was also shown to be highly variable particularly in the semi-arid northern part of Nigeria. The possible implications of the observed trends and variability for rain-fed farming in the country include reduction in both the total annual rainfall and length of the growing season which will further shoot up the farm management expenses due to the need for irrigation to supplement the seasonal rainfall, particularly in the semi-arid northern part of the country.

Though it is observed that the percentage frequency of occurrence of severe and extreme droughts is low in the country, there is need to intensify efforts for drought prediction and issuance of drought early warning systems in the country in order to mitigate the effects of drought when it occurs, since droughts have not been assumed to have ended in the country. The connection between the Nigerian rainfall series and remote factors of El-Nino/ENSO as revealed in this study is a pointer to the need for active research on the prediction of El-Nino/ENSO and Sea Surface Temperature (SST) anomalies over the tropical oceans (Pacific, Atlantic and Indian) in order to reduce the harmful effects of these meteorological events.

Since the space-time variations in the rainfall distribution patterns in Nigeria are associated with the south-north seasonal migration of the intertropical discontinuity (ITD) as shown in this study, the prediction of the seasonal rainfall, onset and retreat of the rainy season in the country using both ground and upper air data of rainfall, oceanic temperatures, air temperature, air pressure and precipitable water becomes necessary in order to aid agricultural planning. This could reduce the crop failures associated with restricted northward advance of Intertropical Discontinuity (ITD) and its rapid retreat. The study contributes to our deeper understanding of the climate teleconnections and their implications for Nigerian rainfall climatology.

REFERENCES

Aceituno, P. 1992. El-Nino, the Southern Oscillation and ENSO: Confusing names for a complex ocean-atmosphere interaction. *Bulletin of American Meteorological Society* 73: 483-485.

Acheampong, P. K. 1987. Rainfall characteristics over West African Sahel-the Nigerian example. *Journal of Tropical Geography* 8: 72-83.

Adedokun, J.A. 1978. Towards an achieving an in- season forecast of West African precipitation.*Geophysics* 28: 19-38.

Adedoyin, J.A. 1989a. Global-scale sea surface temperature anomalies and rainfall characteristics in northern Nigeria. *International Journal of Climatology* 9:133-144.

______________1989b. Initiation of West African squall lines. *Journal of Atmospheric Physics* 41: 99-103.

______________1989c. Wave propagation and amplication along the surface of discontinuity between airmasses of different atmospheric stabilities: A case study of West African tropospheric airmasses. *Journal of Atmospheric Physics* 41: 105-112.

Adefolalu, D. O. 1986. Rainfall trends in Nigeria. *Theoretical and Applied Climatology* 37.4.

Adejokun, J. A. 1966. The three-dimensional structure of the inter tropical discontinuity. *Nigerian Meteorological Services*, *Lagos*. Note 39.

Adejuwon, J.O., Balogun, E. E. and Adejuwon, S. A. 1990. On the annual and seasonal patterns of rainfall fluctuations in Sub-Saharan West Africa. *Internationa Journal of Climatology* 10: 839-848.

Adejuwon, J. O. and Odekunle, T. O. 2006. Variability and severity of the little dry season in south-western Nigeria. *Journal of Climate* 19: 483-493.

Adelekan, I.O. 1998. Spatio-temporal variations in thunderstorm rainfall over Nigeria, *International Journal of Climatology* 18: 1273-1284pp.

Agnew, C.T. 1999. Using the SPI to identify Drought. *Drought Network News* 12. 1: 6-12.

Akintola, F. O .1982. Relief and drainage. *Nigeria in Maps*. K. M. Barbour, J. S. Oguntoyinbo, J. O. C. Onyemelukwe and J. C. Nwafor. Eds. Hodder and Stoghton.

Allen, R. J., Lindesay, J. A. and Reason, C. J. C. 1996. Multi-decadal variability in the climate system over the Indian ocean region during Austral summer. *Journal of Climate* 110: 1853-1873.

Alley, W. M. 1984. The Palmer Drought Severity Index: limitations and assumptions. *Journal of Climatology* 32: 1100 – 1109pp.

Anyadike, R.N.C .1993. Seasonal and annual rainfall variations over Nigeria. *International Journal of Climatology* 13: 567-580pp.

Anyadike, R. N. C. 1987. Measures for containing the effects of drought on the environment. *Austerity and the Nigeria society*. G.E.K.Ofomata and N.I. Ikpeze. Eds. University of Nigeria Press. 367pp

Aondover, A. T. and Woo, M .1998. Changes in rainfall characteristics in northern Nigeria. *Internaational Journal of Climatology* 18: 1261-1271.

Aondover, A. T. and Tarhule-Lips, R. F. A. 2001. Discontinuities in precipitation series in the West African Sahel. *Physical Geography* 22. 5: 430 - 448.

Appa, R. G. 1987. Drought probability maps. *WMO CAgm* 24.

Areola, O . 1982. Land use. *Nigeria in Maps*. K. M. Barbour, J. S. Oguntoyinbo, J. O. C. Onyemelukwe and J. C. Nwafor. Eds. Hodder and Stoghton.

__________1994. The spatial growth of Ibadan city and its impact on the rural hinterland. *Ibadan region*. M. O. Filani, F. O. Akintola and C. O. Ikporukpo Eds. Rex Charles Publication.

Arkin, P. A. 1982. The relationship between inter-annual variability in 200mb tropical wind field and the Southern Oscillation. *Monthly Weather Review* 110: 808-823.

Ayeni, Bola. 1979. *Concepts and techniques in urban analysis*. Croom Helm: London.

Ayoade, J.O. 1973. Annual rainfall trends and periodicity in Nigeria. *Nigerian Geographical Journal* 16: 167-176.

___________1974. A statistical analysis of rainfall over Nigeria. *Journal of Tropical Geography* 39: 11-23.

___________1977. On the use of the multivariate technique in climate classification and regionalization. *Arch. Met. Geophysics* 53: 257-267.

___________1983. *Introduction to climatology for the tropics*. John Wiley and Sons: Chichester. U. K.

___________1988. *Tropical hydrology and water resources*. MacMillan: London.

___________1995. *Climate and human welfare*. Inaugural lecture delivered at the University of Ibadan. 15th June 1995. Ibadan University Press. 43Pp.

___________2008. *Techniques in climatology*. Stirling- Horden Publishers. Ibadan.

Ba, M. B., Nicholson, S. E. and Frouin, R. 2001. Temporal and spatial variability of surface radiation budget over the African continent as derived from METEOSAT. Part 11: temporal and spatial variability of surface global solar irradiance, albedo and net radiation. *Journal of Climate* 14: 60-76.

Bakun, A. 1978. Guinea current upwelling. *Nature* 271: 147-150.

Balogun, E.E. 1981. Seasonal and spatial variations in thunderstorm activity over Nigeria. *Weather* 36: 192-196.

Balogun, O. Y. 1983. *Senior secondary atlas*. Longman.

Balling, R. C. 1991. Impact of desertification on regional and global warming. *Bulletin of American Meteorolofical Society* 72: 232-234.

Barbour, K. M., Oguntoyinbo, J. S., Onyemeluke, J. O. C. and Nwafor, J. C. 1982. *Nigeria in maps*. Hodder and Stoughton: London.

Barnett, T. P. 1977. An attempt to verify some theories of El-Nino. *Geophysics* 7: 633-647.

Barnett, T. P. 1983. Interaction of the monsoon and Pacific trade wind system at interannual scales. Part 1: The equatorial zone. *Monthly Weather Review* 111: 756-773.

Barry, R. G. and Chorley, R. J. 1976. *Atmosphere, weather and climate.* 3rd ed. Methuen: London.

Battisti, D. S. and Hirst, A. C. 1989. Inter-annual variability in a tropical atmosphere-ocean model: influence of the basic state, ocean geometry and non-linearity. *Journal of Atmospheric Science* 46: 1687-1712.

Beer, T., Greenhut, G. K. and Tandoh, S. E. 1977. Relations between the Z criterion for the subtropical high, Hadley cell parameters and the rainfall in northern Ghana. *Monthly Weather Review* 105: 849-855.

Bello, N. J. 1998. Evidence of climate change based on rainfall records in Nigeria. *Weather* 53. 12: 412-418.

Bello, N. J. and Oni, F. O. 2000. A study of occurrences of wet and dry spells during the rainy season in the Niger Basin Development Authority Area, Nigeria. *Int. J. Environment and Development* 4. 2: 44-48.

Berlage, H. P. 1966. *The southern oscillation and world weather*. Meteorological Institute. 152pp.

Bhalme, H. N. and Mooley, D. A. 1980. Large-scale drought/floods and monsoon circulation. *Monthly Weather Review* 108: 1197-1208.

Bhatt, U .1989. Circulation regimes of rainfall anomalies in the Africa-South Asian monsoon belt. *Journal of Climate* 2: 1133-1144.

Bigg, G. R. 1990. El-Nino and the Southern Oscillation. *Weather* 45: 2-8.

Bjerknes, J. 1966. A possible response of the atmospheric Hadley circulation to equatorial anomalies of ocean temperature. *Tellus* 18: 820-829.

__________1969. Atmospheric teleconnections from the equatorial Pacific. *Monthly Weather Review* 97: 163-172.

Bonaccorsa, B.I., Bordi, A., Canceliere, G. and Sutera, A. 2003. Spatial variability of drought: An analysis of the SPI in Sicily. *Water Resources Management* 17: 273-296.

Bordi, I., Fragio, S., Parenti, P., and Speranza, A. 2001. The analysis of the standardized Precipitation index in the Mediteranean area: part I. *Journal of the Annuli Di Geofisica* 44. 5: 979-993.

Bradley, R. S., Diaz, H. F., Kiladis, G. N. and Eischeid, J. K. 1987. ENSO signal in continental temperature and precipitation records. *Nature* 327: 497-501.

Bryson, R. A. 1973a. Climatic modification by air pollution 11: The Sahelian effect. *Institute for Environmental Studies*. University of Wisconsin: 12pp.

___________1973b. Drought in Sahelia: Who or what is to blame? *Ecologist* 3: 366 371.

Bum, D. H. and Hag-Elnur, M. A. 2002. Detection of hydrologic trend and Variability. *Journal of Hydrology* 255:107-122.

Bunting, A. H., Dennett, M. D., Elston, J. and Milford, J. R. 1976. Rainfall trends in the West African Sahel. *Quarterly Journal of Royal Meteorological Society* 102: 59-64.

Burpee, R. W. 1972. The origin and structure of easterly waves in the lower troposphere of north Africa. *Journal of Atmospheric Science* 29: 77-90.

____________1974. Chacteristics of North Africa easterly waves during the summers of 1968 and 1969. *Journal of Atmospheric Science* 31: 1556-1570.

Burt, J. E., Hayes, J. T., O'Rourke, P. A., Terjung, W. H. and Todhunter, P. E. 1981. A parametric cropwater use model. *Water Resources Research* 17: 1095-1105.

Cane, M. A. 1991. Forecasting El-Nino with a geophysical model. *Teleconnections linking worldwide climate anomalies*. M. H. Glantz, R. W. Katz and N. Nicholls. Eds. Cambridge University Press: Cambridge. 345-369.

Ceituno, P. 1988. On the functioning of the Southern Oscillation in the South American sector. Part 1: surface climate. *Monthly Weather Review* 16: 505-524.

Charney, J. G. and Stern, M. E. 1962. On the stability of internal baroclinic jets in a rotating atmosphere. *Journal of Atmospheric Science* 19: 159-172.

Charney, J., Stone, P. H. and Quirk, W. J. 1975. Drought in the Sahara: a biophysical feedback mechanism. *Journal of Atmospheric Science* 187: 434-435.

Charney, J. G. 1975. Dynamics of desert and drought in the Sahel. *Quarterly Journal of Royal Meteorologicaal Society* 101: 193-202.

Chen, W. Y. 1980. Assessment of the southern oscillation sea level pressure indices. *Monthly Weather Review* 110: 800-807.

Chervin, R. M. 1979. Response on the NCAR general circulation model to changed land surface albedo. Report of the JOC study Conference on Climate models. *GARP Publication Service* 22: 563-581.

Citeau, J., Finnaud, L., Cammas, J. P. and Demaroq, H. 1989. Questions relative to ITCZ migration over tropical Atlantic ocean, sea surface temperature and Senegal river run-off. *Journal of Meteorological and Atmospheric Physics* 41: 181-190.

Cohn, T. A. and Stedinger, J. R. 1987. The use of historical information in a maximum likelihood framework. *Journal of Hydrology*. 96: 215-223.

Cook, K. H. 2001. A Southern Hemisphere wave response to ENSO with implications for Southern Africa precipitation. *Journal of the Atmospheric Sciences* 58: 2146–2162.

Courel, M. F., Kandel, R. S. and Rasool, S. I. 1984. Surface albedo and the Sahel drought, *Nature* 307: 528-531.

Covey, D. L. and Hastenrath, S. 1978. The Pacific El-Nino phenomenon and the Atlantic circulation. *Monthly Weather Review* 106: 1280-1286.

Cunnington, W. M. and Rowntree, P. R. 1986. Simulations of the Saharan atmosphere-dependence on moisture and albedo. *Quarterly Journal of Royal Meteorological Society* 112: 971-999.

Dai, A. and Wigley, T. M. L. 2000. Global patterns of ENSO-induced precipitation. *Geophysical Research Letters*, 27.9: 1283-1286.

Dalezions, R.N., Lonkas, A., Vasiliades, L. and Liakapoulos, E. 2000. Severity – Duration - Frequency Analysis of droughts and wet periods in Greece. *Journal of Hydrological Science* 45.5: 751-769.

Damaree, G.R. and Nicolis, C. 1990. Onset of Sahelian drought viewed as a fluctuation – induced transition. *Quaterly Journal of Royal Meteorological Society* 116: 221-238.

Dennett, M. D., Elston, J. and Rodgers, J. A. 1985. A reappraisal of rainfall trends in the Sahel. Journal of Climatology 5: 353-361.

Deser, C. and Wallace, J. M. 1990. Large-Scale atmospheric circulation features of warm and cold episodes in the tropical Pacific. *Journal of Climate* 3: 1254-1281.

Dhonneur, G. 1971. General circulation and types of weather over Western and Central Africa. *Annex-iv*. GARP- GATE 23 Design: 22Pp.

Diaz, H. F. and kiladis, G. N. 1992. Atmospheric teleconnections associated with the extreme phases of the Southern Oscillation. *El-Nino: Historical and Paleoclimatic aspects of the Southern Oscillation.* H. F. Diaz and V. Margraf. Eds. Cambridge University Press. Cambridge: 7-28.

Diraper, N. R. and Smith, H. 1966. *Applied regression analysis*. JohnWiley. New York. 407Pp.

Dracup, J.A., Lee, K. S., and Paulson, E.G. 1980. On the definition of droughts.*Water Resources Research* 16.2: 297-302.

Dracup, J. A. and Keyantash, J. 2002. The quantification of drought: An evalauation of drought indices *BAMS* 83: 1167-1180.

Druyan, L. M. 1988. Comments on rainfall variability in equatorial and Southern Africa: relationship with sea surface temperatures along the South-western coast of Africa. *Journal of Climate* 1: 569-581.

______________1989. Advances in the study of Sub-Saharan drought. *International Journal of Climatology* 9: 77-90.

______________1991. The sensitivity of Sub-Saharan precipitation to Atlantic SST. *Journal of Climate* 18: 17-36.

_____________ 2011. Studies of the 21st –century precipitation trends over West Africa. *International Journal of Climatology* 31: 1415-1424.

Dury, G. H. 1964. Some results of a magnitude-frequency analysis of precipitation. *Australian Geographical Studies* 2: 21-34.

Ekpoh, I. J. and Nsa, E. 2011. Extreme climatic variability in North-Western Nigeria: An analysis of rainfall trends and patterns. *Journal of Geography and Geology*, 3. 1: 51-62.

Eldridge, R. H. 1957. A synoptic study of West African disturbance lines. *Quarterly Journal Royal Meteorological Society* 83: 303-314.

El-tahir, E.B and Gong, C. 1996. Dynamics of wet and dry years in West Africa. *Journal of Climate* 9: 1030-1042.

El-tahir, E. B. and Gong, W. 2000. Biosphere-Atmosphere interactions over West Africa: Development and validation of a coupled dynamic model. *Quarterly Journal of Royal Meteorological Society* 126: 1261-1280.

Emmanual, K. A. 1987. An air-sea interaction model of intraseasonal oscillation in the tropics. *Journal of Atmospheric Science* 44: 2324-2340.

Falkenmark, M. 1987. Water related constraints to African development in the next few decades. Water for the future: Hydrology in perspective. *IAHS publication* 164: 439-453.

Faniran, A. 1982. Relief and drainage. *Ibadan Region* . M.O Filani, F. O. Akintola and C. O. Ikporukpo. Eds. Rex Charles Publication.

Farmer G. and Wigley. T.M.L. 1985. Climatic Trends for tropical Africa. *A Research Report for Overseas Development*. England.

Farmer. G. 1988. Seasonal forecasting of the Kenya coast short rains, 1901-1984. *Journal of Climatology* 8: 489-497.

Fasheun, A. 1984: Modelling of daily rainfall sequence for farm operating planning in Ibadan. *Nigerian Meteorological Journal* 1: 102-109.

Fleer, H. E. 1981. Teleconections of rainfall anomalies in the tropics. *Monsoon Dynamics*. A. Lighthill and R. Perarce. Eds. Cambridge university Press: 5-18.

Folland, C.K., Palmer, T. N. and Parker, D. E. 1986. Sahel rainfall and world-wide sea surface temperatures. *Nature* 320: 602 -607.

Folland, C., Owen, J., Ward, M. N. and Colman, A. 1991. Prediction of seasonal rainfall in the Sahel region using empirical and dynamic methods. *Journal of Forecasting* 10: 21-56.

Foley, J. A., Levis, S., Prentice, I. C., Pollard, D. and Thompson, S. L. 1998. Coupling dynamic models of climate and vegetation. *Global Climate Biology* 4: 561-579.

Fontaine, B and Janicot, S. 1992. Wind-field coherence and its variations over West Africa. *Journal of Climate* 5: 512-524.

Fontaine, B and Bigot, S. 1993. West African rainfall deficits and sea surface Temperatures. *International Journal of Climatology* 13: 271-285.

Fontain, B., Janicot, S. and Moron, V. 1995. Rainfall anomaly patterns and wind-field signals over West Africa in August, 1958-89. *Journal of Climate* 8:1503-1510.

Fordyke, A.G. 1949. Weather forecasting in tropical regions .*Geophysics* 82: 1-47.

Fu, C., Diaz, H. F. And Fletcher, J. O. 1986. Characteristics of the response of sea surface temperature in the central pacific associated with warm episodes in the southern oscillation. *Monthly Weather Review* 114: 1715-1738.

Garnier, B.J. 1967. Weather conditions in Nigeria. *Climatological Research Series* 2: 21-32

Gibbs, W.J. 1987. Defining climate. *WMO Bulletin* 36. 4.

Giladis, G. N. and Diaz, H. F. 1989. Global climatic anomalies associated with the extremes in the Southern Oscillation. *Journal of Climate* 2: 1069-1090.

Gill, A. E. and Rasmusson, E. M. 1983. The 1982-83 climate anomaly in the equatorial Pacific. *Nature* 306: 229-234.

Gillooly, J. F. and Walker, N. D. 1984. Spatial and temporal behavior of sea surface temperatures in the South Atlantic. *South African Journal of Science* 80: 97-100.

Girux, L. A. 2001. Towards characterizing and planning for drought in Vermont- part 1: A climatological perspective. *Journal of Atmospheric and Water Resources* 37: 505-525.

Glantz, M. H. and Katz, R. W. 1987. African drought and its impacts: revived interest in a current phenomenon. *Desertification Control Bulletin* No. 4: 22-30.

Glantz, M. H. 1994. *Drought follows the plow*. Cambridge University Press.

Goddard, L. and Graham, N. E. 1999. Importnace of the Indian ocean for simulating rainfall anomalies over eastern and southern Africa. *Geophysical Research* 104: 19099-19116.

Goldenberg, S. B. and Shapiro, L. J. 1996. Physical mechanisms for the association of El-Nino and West African rainfall with Atlantic major hurricane activity. *Journal of Climate* 9: 1169-1187.

Gray, M. W. and Love, G. B. 1981. The role of general circulation in tropical cyclone formation: Program and abstracts. *IAMAP Third scientific assembly on the Dynamics of the General Circulation of the Atmosphere*, Hamburg, IAMAP, p. 23.

Greenhut, G. K. 1977. A new criterion for locating the subtropical high in West Africa. *Journal of Applied Meteorology* 16: 729-734.

Gribbin, J. and Lamb, H.H. 1979. Climatic change in historical times. *Climatic change*. J. Gribbin. Ed. Oxford University Press.

Griffiths, J. F. 1972. *Climates of Africa*. Elsevier. 604pp.

Grist, J. P. 2002. Easterly waves over Africa. Part I: The seasonal cycle and contrasts between wet and dry years. *Monthly Weather Review*130: 197–211.

Grist, J. P. and Nicholson, S. E. 2001. A study of the dynamic factors influencing the interannual variability of rainfall over West African Sahel. *Journal of Climate* 14:1337-1359.

Grove, A. T. 1972. Climatic change in Africa in the last 20, 000 years. *Septentrional* 2: 141-153.

___________1973. A note on the remarkably low rainfall of the Sudan in 1913. *Savannah* 2: 133-138.

Guo, S. L. 1990. Unbiased plotting position formulae for historical floods. *Journal of Hydology* 121: 45-61.

Gutman, N.B. 1999. Accepting the standardized precipitation index: a calculation Algorithm. *Journal of Water Resources* 322: 311-322.

Hastenrath, S. 1985. Inter annual variability and annual cycle: mechanisms of circulation and climate in the tropical Atlantic sector. *Monthly Weather Review* 112: 1097-1107.

_____________1990. Decadal-scale changes of the circulation in the tropical Atlantic sector associated with Sahel drought. *International Journal of Climatology* 10: 459-472.

_____________1995. Recent advances in tropical climate prediction. *Journal of Climate* 8. 6:1519-1532.

Hastenrath, S. and Heller, L. 1977. Dynamics of climatic hazards in northeast Brazil. *Quarterly Journal Royal Meteorological Society* 103: 77-92.

Hamilton, R. A. and Archbold, J. W. 1945. Meteorology of Nigeria and adjacent territory. *Quarterly Journal of Royal Meteorological Society* 17: 231-264.

Hamilton, K. and Allingham, A. M. 1988. A note on equatorial Atlantic sea surface temperature variations, 1890-1979. *Oceanography* 26: 666-678.

Halpert, M. S. and Popelewski, C. F. 1992. Surface temperature patterns associated with the Southern Oscillation. *Journal of Climate* 5: 577-593.

Hayes, M. J., Svoboda, M. D. and Vanyarkho, O. V. 1999. Monitoring the 1996 drought using the standardized precipitation index. *Bulletin of the of the American Meteorological Society*, 80, Pp. 429-438.

Hess, T. M., Stephens, W. and Maryah, U. M. 1995. Rainfall trends in the Northeast arid zone of Nigeria 1961-1990. *Agriculture and Forest Meteorology*. 74: 87-97.

Hirsch, R. M., Slack, J. R. and Smith, R. A. 1982. Techniques of trend analysis for monthly water quality data. *Water Resources Research* 18: 107-121.

Hisard, P. and Henin, C. 1987. Response of the equatorial Atlantic ocean to 1983-84 winds. *Geophysics* 92: 3759-3768.

Horel, J. D. and Wallace, J. M. 1981. Planetary-scale atmospheric phenomena associated with the Southern Oscillation. *Monthly Weather Review* 109: 813-829.

Horel, J. D., Kousky, V. E. and Kagano, M. T. 1986. Atmospheric conditions in the Atlantic sector during 1983 and 1984. *Nature* 322: 248-251.

Hoskins, B. J. and Karoly, D. J. 1981.The steady linear response of a spherical atmosphere to thermal and orographic forcing. *Journal of Atmospheric Science* 38: 1179-1196.

Hoskins, B. J. and Pearce, R. P. 1983. *Large-Scale dynamical processes in the atmosphere*. Academic Press.

Hosking, J. R. M. and Wallis, J. R. 1986. The value of historical data in flood frequency analysis. *Water Resources Research* 22: 1606-1612.

Hounam, C. E., Burgos, J. J., Palmer, W. C. and Rodda, J. 1975. Drought and agriculture. *WMO Technical Note* 138.

Houze, R. A. and Betts, A. K. 1981. Convection of GATE. *Reviews of Geophysics and Space Physics* 19: 541-576.

Hulme, M. 1992. Rainfall changes in Africa: 1931-1960 to 1961-1990. *International Journal of Climatology* 12: 685-699.

_________1996. Recent climatic change in the world's dry lands. *Geophysics Research Letters* 23: 61-64.

_________2000. Climatic perspective on Sahelian dessication: 1973-1998. *Global Environmenatal Change* 11: 19-29.

Hulme, M. and Kelly, M.1993. Exploring the links between desertification and climate change. *Environment* 35: 4-11.

Iloeje, N. P .1981. *A New Geography of Nigeria*. 3rd ed. Longman: Great Britain.

Ireland, A. W. 1962. The little dry season of southern Nigeria. *Nigerian Geographical Journal* 5: 7-20.

Jagannathan, P. and Parthasarathy, B. 1973. Trends and periodicities of rainfall over India. *Monthly Weather Review* 101: 691-700.

Janicot, S. 1992. Spatio-temporal variability of West African rainfall. Par 1: Regionalizations and typings. *Journal of Climate* 5: 489-497.

_________1996. Sahel droughts and ENSO dynamics. Geophysical Research Letters 23. 5: 515-518.

Janowiak, J. 1988. An investigation of interannual variabilitry in Africa. *Journal of Climate* 1: 240-255.

Jin, F. F. and Neelin, J. D. 1993. Modes of inter-annual tropical ocean-atmosphere interaction- A unified view. Part 1: Numerical results. *J. Atmos. Science* 50: 811-829.

Jury, M. R., Valentine, H. R. and Lutjeharms, J. R. E. 1993. Influence of the Agulhas current on summer rainfall along the southeast coast of Africa, *Journal of Applied Meteorology* 32: 1282-1287.

Kanamisu, M and Krisnamurti, T. N. 1978. Northern summer tropical circulations during drought and normal rainfall months. *Monthly Weather Review* 106: 331-347.

Katz, E. J., Hisard, P., Verstraete, J. M. and Garzoli, S. L. 1986. Annual change of sea surface slope along the equator of the Atlantic ocean in 1983 and 1984. *Nature* 222: 245-247.

Keer, A. 1985. Fifteen years of African drought. *Science* 227: 1453-1454.

Kendall, M. G. 1975. *Rank correlation measures*. Charles Griffin: London.

Khandekar, M. L. 1979. Climatic teleconnections from the equatorial Pacific to the Indian monsoon-analysis and implications. *Archives for Meteorological Geophysics* 28A: 159-168.

Kessler, W. S., Mcphaden, M. J. and Weikman, K. M. 1996. Forcing of the intraseasonal Kelvin waves in the equatorial Pacific. *Geophysics* 100: 10613-10631.

Kendall, M. G. and Stuart, A. 1961.*The advanced theory of statistics*. Hafner Publishing Company. NewYork. U. S. A.

Kidson, J. W. 1977. African rainfall and its relation to the upper air circulation. *Quarterly Journal of Royal Meteorological Society* 103: 441-456.

Kiladis, G. N. and Diaz, H. F. 1989. Global climatic anomalies associated with the extremes of the Southern Oscillation. *Journal of Climate* 2: 1069-1090.

Kiladis, G. N. and Van Loon, H. 1988. The Southern Oscillation Part vii: Meteorological anomalies over the Indian and Pacific sectors associated with the extremes of the oscillation. *Monthly Weather Review* 116: 120-136.

Kiladis, G. N., Von Storch, H. and Van Loon, H. 1989: Origin of the South Pacific Convergence Zone. *Journal of Climate* 2: 1185-1195.

Kitoh, A., Yamazaki, K. and Tokioka, T. 1988. Influence of soil moisture and surface albedo changes over the African tropical rain forest on summer climate investigated with the MRI-GCM-1. *Journal of Meteorological Science* 66: 65-85.

Klaus, D. 1978. Spatial distribution and periodicity of mean annual precipitation south of the Sahara. *Arch. Met. Geophys* B26: 17-27.

Koslowski, G. and Glaser, R. 1995. Reconstruction of the ice winter severity since 1701 in the Western Baltic. *Journal of Climate* 31: 79-98.

Kowal, J. M. and Knabe, D. T. 1972. *An agroclimatological atlas of northern states of Nigeria*. Ahmadu Bello University Press, Zaria.

Kowal, J. M. and Kassam, A. H. 1973. An appraisal of drought in 1973: Affecting groundnut production in the Guinea and Sudan savannah areas in Nigeria. *Savannah* 2.2: 13-19.

Kraus, E.B. 1977a. Subtropical droughts and cross equatorial energy transports, *Monthly Weather Review* 105: 1009-1018.

__________1977b: The seasonal excursion of the Intertropical Convergence Zone. *Monthly Weather Review* 105: 1052-1055.

Krueger, A. F. and Winston, J. S. 1974. A comparison of the flow over the tropics during two contrasting circulation regimes. *Journal of Atmospheric Science* 31: 358-370.

Kutzbach, J. E. 1980. Estimates of past climate at Palaeo Lake Chad , North Africa based on a hydrological and energy balance model. *Quaternary Research* 14: 210-223.

Landsea, C. W. and Gray, W. M. 1992. The strong association between western Sahelian monsoonal rainfall and intense Atlantic hurricanes. *Journal of Climate* 5: 435-453.

Landsea, C. W. and Knaff, J. A. 2000. How much skill was there in forecasting the very strong 1997-98 El Niño? *Bull Amer. Meteor. Soc*. 81: 2107—2119.

Lamb, P. J. 1973. Large- scale tropical Atlantic surface circulation patterns associated with Sub-Saharan weather anomalies. *Tellus* 30: 240-251.

_________1980. Persistense of sub-Saharan drought. *Nature*. 299: 46-48.

_________1985. Rainfall in Sub-Saharan Africa during 1941-1983. *Gletshcherk Glazialgeo* 21: 131-139.

Lamb, P. J. and Peppler, R. A .1992. Further case studies of tropical Atlantic surface atmospheric and oceanic patterns association with Sub-Saharan drought. *Journal of Climate* 5: 476-488.

Landsberg, H. E. 1975. Sahel Drought: change of climate or part of climate? *Archieves for Meteorology, Geophysics and Bioclimatology*, Ser. B. 23: 193-200.

Lau, K. M. and Peng, L. 1987. Origin of low frequency intraseasonal oscillation in the tropical atmosphere. Part 1: Basic theory. *Journal of Atmospheric Science* 44: 465-472.

Lau, K. M. and Sheu, P. J. 1988. Annual cycle, quasi-biennial oscillation and Southern Oscillation in global precipitation. *Geophysics* 93: 10975-10988.

Le Barbe, L., Lebel, T. and Tabsoba, D. 2002. Rainfall variability in West Africa during the years 1950-90. *Journal of Climate* 15: 187-202.

Lettenmaier, D. P., Wood, E. F. and Wallis, J. R. 1994. Hydro-climatological trends in the continental United States, 1948-88. *Journal of Climate* 7: 586-607.

Lim, H. and Chang, C. P. 1981. A theory for midlatitude forcing of tropical motions during winter monsoons. *Journal of Atmospheric Science* 38: 2377-2392.

Lindesay, J.A., Harrison, M. S. J. and Haffner, M. P. 1986. The southern oscillation and south African rainfall. *South African Journal of Science* 82: 196-198.

Lockwood, J.G.1979. *Causes of Climate*. Edward Arnold: London.

Long, M., Entekhabi, D. and Nicholson, S. E. 1998. Interannual variability in rainfall, water vapor flux and vertical motion over West Africa. *Water Resources Variability in Africa during the 2oth century*. IAHS Publications No. 252, Wallingford, Pp. 19-26.

________________________and____________2000. Interannual variability in rainfall, water vapor flux and vertical motion over West Africa. *Journal of Climate* 13: 3827-3841.

Lough, J.M. 1986. Tropical Atlantic sea surface temperatures and rainfall variability in Sub-Saharan Africa. *Monthly Weather Review* 114: 561-570.

Madden, R. A. and Jullian, P. R. 1994. Observations of the 40-50 dry tropical oscillation-a review. *Monthly Weather Review* 122: 814-837.

Mann, H. B. 1945. Non-Parametric tests against trend. *Econometrica* 13: 245-259.

Mason, S. J. 1995. Sea surface temperature-South African rainfall associations, 1910-1989. *International Journal of Climatology* 15: 119-135.

Mattai, H. F. 1979. Hydrologic and human aspects of the 1976-77 drought. *UGS Report* 1130.

McKee, B. T., Doeskin, N., Kleist, J. 1993. The relationship of drought frequency and duration to time scales. *Proceedings of the 8th International Conference on Applied Climatology*.17th -22nd January 1993. Anheim. California: 21-28.

McKee, B.T., Nolan, J., Doeskan, N.J. and Kleist, J. 1995. Drought Monitoring with multiple time scales. *Proceedings of the 9th International Conference on Applied Climatology*. 15th -20th January 1995. Boston: Massachusetts, U.S.A. 233-236.

Mechoso, C. R and Lyons, S. W 1988. On the atmospheric response to SST anomalies associated with the Atlantic warm event during 1984. *Journal of Climate* 1: 422-428.

Meehl, G. A. 1987. The annual cycle and interannual variability in the tropical pacific and Indian ocean regions. *Monthly Weather Review* 115: 27-50.

___________1993. A couple air-sea mechanisms in the tropical Indian and Pacific regions: role of the oceans. *Journal of Climate* 6: 31-41.

Michelchen, N. 1985a.The Southern Oscillation and the interannula upwelling variations off northwest Africa and in the Gulf of Guinea. *Ocean- Atmosphere News Letters* 10: 211-220.

Michelchen, N. 1985b. About interannula coastal upwelling variations off NW Africa with reference to changes of the Southern Oscillation. *International Symposium on Upwelling in Western Africa* 1: 93-100.

Moore, D. W., Hisard, P., McCreary, J. P., Merie, J., O'Brien, J., Picaut, J., Verstraete, M. and Wunsch, C. 1978. Equatorial adjustment in the eastern Atlantic. *Geophysics* 5: 637-640.

Morse, B. 1987. Forward. *Drought and hunger in Africa: Denying famine a future*. M. H. Glantz. Ed. Cambridge University Press. 457pp.

Mortimore, M. 1973. Famine in Hausa land. *Savannah* 33.2: 103-107.

____________1989. *Adapting to drought: Farmers, famine and desertification in West Africa*. Cambridge University Press.

____________2010. Adapting to drought in the Sahel. *WIRES Climate Change* 1: 134-143

Mo K. C., Bell, G. D. and Thiaw, M. W. 2001. Impact of sea surface temperature anomalies on the Atlantic tropical storm activity and West African rainfall. *Journal of Atmospheric Sciences* 58: 3477-3496.

Miles. M.K. and Folland, C.F. 1974. Changes in the latitude of the climatic zones of the Northern Hemisphere. *Nature* 252: 616-629.

Miller, R. I. and Lindenz, R. 1992. Organization of rainfall by an unstable jet with an application to African waves. *Journal of Atmospheric Science* 49: 1523-1540.

Morgan, W.B and Solarz, J.A. 1994. Agricultural crisis in Sub-Saharan Africa: Development Constraints and Policy problems. *The Geographical Journal* 160.1: 57-73.

Mulenga, H. M., Rouault, M. and Reason, C. K. 2003. Dry summers over northeastern South Africa and associated circulation anomalies. *Climate Research* 25; 29-41.

Munemoto, M. and Tachibana, Y. 2012. The recent trend of increasing precipitation in Sahel and the associated inter-hemispheric dipole of global SST. *International Journal of Climatology* 32. 1346-1353.

Namias, J. 1963. The art and science of long-range forecasting. *EOS* 61: 449-450

_________1969. Seasonal interactions between the North Pacific ocean and the atmosphere during the 1960's. *Mon. Wea. Rev*. 97: 173-192.

_________1974. Suggestions for research leading long-range precipitation forecasting for the tropics. *Preprints Institut for Tropical Meteorology. Kenya:* 141-144.

_________1981. State of the art predicting short period climatic variations. *Food-Climate interactions*. W. Bach, J. Pankrath and S. H. Schneider. Eds. Dordrecht: Reidel. 399 - 422.

Navarra, A., Tribbia, J. and Conti, G. 2013. Atmosphere-ocean interaction at strong couplings in a simple model of El-Nino. *American Meteorological Society* 26: 9633-9654.

Neelin, J. D. I., Held, M. and Cook, K. H. 1987. Evaporation-wind feedback and low frequency variability of the tropical atmosphere. *Journal of Atmospheric Science* 44: 2341-2348.

Newell, R. E. and Weare, B. C. 1976. Ocean temperatures and large- scale atmospheric variations. *Nature* 262: 40-41.

Newell, R.E. and Kidson, J.W.1984. Africa mean wind changes between Sahelian wet and dry periods. *Journal of Climatology* 4: 27-33.

Nicholson, S.E. 1978. Revised rainfall series for the West African sub-tropics. *Monthly Weather Review* 107: 620-623.

_______________1980. The nature of rainfall fluctuations in Subtropical West Africa. *Monthly Weather Review* 108: 473-487.

_______________1981. Rainfall and atmospheric circulation during drought period and wetter years in West Africa. *Monthly Weather Review* 109: 2191-2208.

Nicholson, S. E. 1983. Sub- Saharan rainfall in years 1976-1980: Evidence of continued drought. *Mon. Wea. Review* 111: 1646-1654.

_______________1985. Sub-Saharan rainfall, 1981-1984. *Journal of Applied Climatology* 24: 1388-1391.

_______________1986a. The spatial coherence of African rainfall anomalies- Interhemispheric teleconnections. *Journal of Applied Climatology* 25: 1365-1381.

_______________1986b. The Nature of rainfall variability in Africa south of the Equator. *Journal of Climatology* 6: 515-530.

______________1989a. African Drought: Characteristics, causal theories, and global teleconnections . *Understanding climate change*. A. Derger, R. E. Dickenson and T. Kidson. Eds. American Geophysics Union. 79-100.

______________1989b. Rainfall variability in Southern Africa and its relationship to ENSO and the Atlantic and Indian oceans. *Preprints of the 3rd International conference on Sothern hemisphere meteorology and oceanography.* American Meteorological Society. 366-367.

______________1993. An overview of African rainfall fluctuations in the last decade. *Journal of Climate* 6: 1463-1466.

______________1996. A review of climate dynamics and climate variability in eastern Africa. *The limnology, climatology and paleoclimatology of the East African lakes*. T. C. Johnson and E. Odada. Eds. Gordon and Breach: Amsterdam. 25-56.

______________ 2001. Climatic and environmental change in Africa during the last two centuries. *Climatic Research* 17: 123-144.

Nicholson, S. E. and Flohn, H. 1980. African environmental and climatic changes in late Pleistocene and Holocene. *Climatic Change* 23: 313-348.

Nicholson, S. E. and Chervin, R. M. 1983. Recent rainfall fluctuations in West Africa-Interhemispheric teleconnections. *Variations in the global water budget*. A. Perrott, M. Beran and C. Ratliffe. Eds.Reidel. Dordrecht: 221-238.

Nicholson, S.E and Entekhabi, D. 1986. The quasi-periodic behavior of rainfall variability in Africa and its relationship to southern oscillation. *Journal of Climate* 34: 331-348.

______________and_____________ 1987. Rainfall variability in equatorial and Sothern Africa: relationships with sea surface temperature along the southwestern coast of Africa. *Journal of Applied Climatology* 26: 561-578.

Nicholson, S. E. and Nyenzi, B. S. 1990. Temporal and spatial variability of SST in the tropical Atlantic and Indian oceans. *Archives for Meteorology* Series A 42: 1-17.

Nicholson, S.E and Palao, I.M. 1993. A re-evaluation of rainfall variability in the Sahel. Part I: Characteristics of rainfall fluctuations. *International Journal of Climatology* 13: 371-389.

Nicholson, S.E and Kim, J. 1997. The relationship of the El-nino/Southern

Oscillation to African rainfall. *International Journal of Climatology* 17.2: 117-135.

Nicholson, S. E., Turker, C. J. and Ba, M. B. 1998. Desertification, drought and surface vegetation: an example from the West African Sahel. *Bulletin of American Meteorological Society* 79: 815-829.

Nicholson, S. E. and Selato, J. C. 2000. The influence of La-Nina on African rainfall. *International Journal of Climatology* 20: 1761-1776.

Nicholson, S. E., B. Some, and B. Kone, 2000: An analysis on recent rainfall conditions in West Africa, including the rainy season of the 1997 ENSO year. *J. Climate,* **13,** 2628–2640.

Nicholson, S. E. 2001. Climate and environmental change in Africa during the last two centuaries. *Climate Reserach* 17: 123-144.

Nicholson, S. E. and Grist, J. P. 2001. A conceptual model for understanding rainfall variability in the West African sahel on inter-annual and inter-decadal timescales. *Int. J. Climatology* 21: 1733-1757.

Nicholson, S. E. 2005. On the question of the "recovery" of the rains in the West African Sahel. *Journal of Arid Environments* 63: 615-641.

Nicholson, S. E. and Webster, P. J. 2007. A physical basis for the inter-annual variability of rainfall in the Sahel. *Quarterly Journal of the Royal Meteorological Society* 133: 2065-2084.

Nicholson, S. E. 2013. The West African Sahel: A review of recent studies on the rainfall regime and its inter-annual variability. *ISRN Meteorology* 20: 1-32.

Nieuwolt, S. 1977. *Tropical climatology*. John Wiley. 207pp.

Norton, C. C., Mosher, F. R. and Hinton, B. 1979. An investigation of surface albedo variations during the recent Sahel drought. *Journal of Applied Meteorology* 18: 1252-1262.

Norquirst, D. C., Recker, E. E. and Reed, R. J. 1977. The energetic of African waves disturbances as observed during phase 111 of GATE. *Monthly Weather Review* 105: 334-342.

Obasi, G. O. P. 1965. Atmospheric, Synoptic and Climatological features over the West African region. *Nigerian Meteorological Service Technical Note* 28.

O'Brien, J. J., Adamec, D. and Moore, D. W. 1978. A simple model of upwelling in the Gulf of Guinea. *Geophysics* 5: 641-644.

Odekunle, T.O. 2004. Rainfall and the length of the growing season in Nigeria.

International Journal of Climatology 24: 1124-1133.

Odekunle, T. O. and Eludoyin, A. O. 2008. Sea surface temperature patterns in the Gulf of Guinea: their implications for the spatio-temporal variability of precipitation in West Africa. *International Journal of Climatology* 28: 1507-1517.

Ogallo, L. A. 1979. Rainfall variability in Africa. *Mon. Wea. Review* 107: 1113-1139.

Ogallo, L.A. 1994.Validity of ENSO-related impact in eastern and southern Africa. *Proceedings of the International Workshop on Usuable Science, Food Security, Early Warning and El-Nino.* 17th-20th October 1993. M. Glantz. Ed. Boulder: Colorado.

Oguntoyinbo, J.S. 1981. Aspects of Urban Microclimate: The case of Ibadan. *Urbanization processes and problems in Nigeria*. P.O. Sada and J.S Oguntoyinbo. Eds. Ibadan University Press.

_________________1982. *Climate and Mankind*. Inaugural Lecture. 24th March 1982. University of Ibadan.

_________________1986. Drought prediction. *Climate Change* 9:79-90.

Oguntoyinbo, J. S. and Richards, P. 1977. The extent and intensity of the 1969-73 drought in Nigeria: a provisional analysis. *Drought in Africa*. 2nd Ed. D. Dalby, J. H. Harrison-Church and F. Bezzaz. Eds. International African Institute. London: 114-126.

Ojo, O. 1977. *The climate of West Africa*. Heineman: London.

Ojo, O. 1986a. The Sahel droughts: 1941-1984. *Proc. ISLSCP Conference*. Rome, Italy. ESA SP-248: 411-415.

_______1986b. Drought persistence in tropical Africa since 1969. *WMO Report Series* 6.1: 73-85.

_______1987. Hydro climatic consequences of climatic events in West Africa: The lessons of the 1969-84 Sahelian droughts, Water for the future: Hydrology in perspective, *IAHS Publication* 164: 229-238.

Okumura, Y. and Xie, S. 2004. Interaction of the Atlantic equatorial cold tongue and the African monsoon. *Journal of Climate* 17: 3589-3602.

Oladipo, E. O. 1985. A comparative performance analysis of three meteorological drought indices. *International Journal of Climatology* 5: 655-668.

______________1995. Some statistical characteristic of drought area variations in the savanna region of Nigeria. *Theoretical and.Applied Climatology* 50: 143-155.

Olaniran, O. J. 1986. On classification of tropical climates for a study of regional climatology: Nigeria as a case study. *Geografiska Annalar* 68a. 4: 543-555.

______________1987. A case study of the seasonal variation of rainfall of different categories in Nigeria in relation to Miller's station types for tropical continents. *Journal of Theoretical and Applied Climatology* 38: 198-209.

______________1988a. The July-August rainfall anomaly in Nigeria. *Climatological Bulletin* 22: 26-38.

______________1988b. The distribution in space of rain days of rainfall of different amounts in the tropics: Nigeria as case study. *Geoforum* 19: 507-520.

______________1990. Changing patterns of rain-days in Nigeria. *Geojournal* 22.1: 99-107.

_______________1991. Evidence of climatic change in Nigeria based on annual rainfall series of rainfall of different daily amounts, 1919-1985. *Journal of Climate change* 19: 319 341.

Olaniran, O. J. and Sumner, G.N. 1989a. A case study of climatic variability in Nigeria based on the onset, retreat and length of the rainy season. *International Journal of Climatology* 9: 253-269.

_______________and______________1989b. Long term variation of annual and growing season rainfall in Nigeria. *Theoretical and Applied Climatology* 40: 2111-2123.

Oliver, J. E. 1991. Al-Biruni's chronology: a source for historical climatology, *Journal of Climate change* 18: 437-452.

Omotosho, J. B. 1985. The separate contribution of squall lines, thunderstorms and monsoon to the total rainfall in Nigeria. *Journal of Climatology* 5: 543-552.

Omotosho, J. B. and Abiodun, B. J. 2007. A numerical study of moisture build-up and rainfall over West Africa. *Meteological Applications* 14: 209-225.

Otterman, J. 1974. Baring high-albedo soils by overgrazing: a hypothesized desertification mechanism. *Science* 186: 531-533.

Oyebande, L. 1990. Drought policy and drought planning in Africa. *Water Resources Development* 6.4: 260-269.

Owen, J. A., Folland, C. K. and Bottomly, M. 1988. Numerical simulation of seasonal Sahel rainfall in four past years using observed sea surface temperatures. *Technical Note No. LRFC15*. Meteorological Office, Bracknell, U. K.

Paeth, H. and Stuck, J. 2004. The West African dipole in rainfall and its forcing mechanism in global and regional models. *Mausam* 55: 561-582.

Paeth, H. and Hensen, A. 2006. On the linear response of tropical African climate to SST changes deduced from regional climate model simulations. *Theoretical and Applied Climatology* 83: 1-19.

Palmer, W. C. 1965. Meteorological Drought. *US Department of Commerce Weather Bureau Research Paper* 45: 1-45.

Palmer, T.W. 1986. Influence of the Atlantic, Pacific and Indian oceans on Sahel rainfall. *Nature* 322: 251-253.

Pasch, R. J. and Avila, L. A.1994. Atlantic tropical systems of 1992. *Monthly Weather Review* 122: 539-548.

Philander, S. G. H. 1985. El-Nino and La-Nina. *Journal of Atmospheric Science* 42: 2652-2662.

________________1986. Unusual conditions in the tropical Atlantic ocean in 1984. *Nature* 222: 236-238.

________________1990. *El-Nino, La-Nina and Southern Oscillation*. Academic Press Incorporation: Toronto.

Parker, D.E., Folland, C. K. and Ward, M. N. 1987. The influence of global sea surface temperatures on African rainfall. *Proceedings of the 1st Technical Conference on Meteorological Research*. 12th-15th January 1987. Nairobi: Kenya.

___________________________and______________1988. Sea surface temperature anomaly patterns and prediction of seasonal rainfall in the Sahel region of Africa. *Recent Climatic Change*. Gregory, S. Ed. Belhaven Press: London.166-178.

Parthasarathy, B. and Dhar, O. N. 1975. Trend analysis of annual Indian rainfall. *Hydrological Sciences Bulletin*. 26: 257-260.

Popelewiski, C.F. and Halpert, M.S. 1987. Global and regional scale precipitation patterns associated with the El-nino/Southern Oscillation. *Monthly Weather Review* 115: 1606-1626.

Popelewiski, C.F. and Jones, P.D. 1987. An extension of the Tahiti-Darwin southern oscillation index. *Monthly Weather Review* 115: 2161-2165.

Popelewiski, C.F. and Halpert. M.S. 1989. Precipitation patterns associated with the high index phase of the southern oscillation. *Journal of Climate* 2: 268-284

Pu, B. and Cook, K. 2012. Role of West African westerly jet in Sahel rainfall variations. *Journal of Climate*, 25: 2880-2896.

Ramage, C. S. 1986. El-Nino. *Scientific American*, Vol. 254, Pp. 76-83.

Rasmusson, E. M. and Carpenter, T. H. 1982. Variations in tropical sea surface temperature and surface wind fields associated with the Southern Oscillation/El-Nino. *Monthly Weather Review* 110: 354-384.

Ramusson, E. M. 1984. El-Nino and variations in climate. *American Scientist* 73: 168-177.

Rasmusson, E. M., Wong, X. and Popelewski, C. F. 1990. The biennial component of ENSO variability. *Journal of Marine Systems* 1: 71-96.

Rasool, S. I. 1984. On dynamics of deserts and climates. *The global climate*. J. Houghton. Ed. Cambridge University Press: 107-116.

Raymond, P. M. 1980. Precipitation patterns in West Africa. *Monthly Weather Review* 108: 1567-1577.

Reason, C. J. C. 2002. Sensitivity of the southern African circulations to dipole sea surface temperature patterns in the South Indian ocean. *International Journal of Climatology* 22: 377-393.

Reason, C. J. C. and Roualt, M. 2002. ENSO-like decadal variability and South African rainfall. *Geophysical Research Letters* 29.13: 1638-1646.

Reason, C. J. C. and Jagadheesha, D. 2005. A model investigation of recent ENSO impacts over southern Africa. *Meteorology and Atmospheric Physics* 89: 181-205.

Reason, C. J. C., Landman, W. A. and tenant, W. 2006. Seasonal to decadal prediction of the southern Africa climate and its links with the variability of the Atlantic ocean. *Bulletin of the Amerian Meteorological Society* 87: 941-955.

Reed, R. 1988. On understanding of the meteorological causes of Sahelian drought. *Pontificae academiae scientiarum scripta varia.* C. Chagas and G. Puppi. Eds. The Vatican City. 180-212.

Rennick, M. A. 1976. The generation of African waves. *Journal of Atmospheric Science* 33:1955-1969.

Rodrigo, F. S., Estaben-Parra, M. J and Castro-Diez, Y. 1995. Reconstruction of total annual rainfall in Andalusa (Southern Sapain) during the 16th and 17th centuaries from documentary sources. *Theoretical and Applied Climatology* 52: 207-218.

Rosenfield, D., Rudich, Y. and Lahav, R. 2001. Desert dust suppressing precipitation: a possible desertification feedback loop. *Proceedings of the National Academy of Science of the United states of America* 98: 5975-5980.

Rosini, E. 1963. A quantitative definition of the meaning of constancy and steadiness of climate. *Proceedings of the Symposium on Changes of Climate*. UNESCO and WMO: 45-48.

Rowell, D. P. and Blondin, C. 1990. The influence of soil wetness distribution on short-range rainfall forecasting in West African Sahel. *Quarterly Journal of Royal Meteorological Society* 116: 1471-1485.

Rowell, D. P. 2001. Teleconnections between the tropical Pacific and the Sahel. *Q. J. R. Meteteorol. Soc.* 127: 1683-1706.

Rowntree, P. R. 1972. The influence of tropical east Pacific temperatures on the atmosphere. *Q. J. Roy. Met. Soc*. 98: 290-321

Rooy, M. P. 1965. A rainfall anomaly index independent of time and space. *Notos* 14: 43-52.

Salas, J. D., Wohl, E. E. and Jarret, R. D. 1994. Determination of flood characteristics using systematic, historical and paleo-flood data. *Coping with floods*. G. Rossi, N. Harmancioglu and V. Yevjevich. Eds. Kluwer Academic Publishers.

Sanford, S. 1978. Towards a definition of drought. *Proceedings of Symposium on Drought.* 21st -25th February 1978. Gaborone.

Sawyer, J.S. 1952. Memorandum on the International Front. *Meteorological Report* 10: 1-14.

Schnecider, J. L. 1969. Evolution du dernier lacutre et Schneiider peuplements prchistoriques aux pays bas du. *Tchad. Bull. Inst. Fr Afri Noire* A 31: 259-263.

Schupelius, G.D. 1976. Monsoon rains over West Africa. *Tellus* 28: 533-536.

Sekoni, I.O. 1992. Spatio - temporal variations in the synoptic origin of rainfall over Nigeria. Unpublished Ph.D.Thesis. Dept. of Geography. University of Ibadan. xvi+237pp.

Sellik, N.P. 1960. Note on humidity and rainfall: Tropical meteorology in Africa. *W.M.O. and Munitalp Foundation Proceedings*.Nairobi: 150-155.

Semazzi, F. H. M., Mehta, V. and Sud, Y. C. 1988. An investigation of the relationship between sub-Saharan rainfall and global sea surface temperatures. *Ocean-Atmosphere* 26.1: 118-138.

________________________and___________1996. A GCM study of the telecommunications between the continental climate of African and global sea surface temperature anomalies. *Journal of Climate* 9.10: 2490-2497.

Seo, K. and Kim, K. 2003. Propagation and initiation mechanisms of Madden-Jullian Oscillation. *Journal of Atmospheric Science* 47: 521-530.

Servain, J. and Legler, D. M. 1986. Emperical orthogonal function analyses of tropical Atlantic sea surface temperature and windstress,1964-1979. *Geophysics* 91: 4181-14191.

Servain, J., Picaut, J. Merle, J. 1982. Evidenced of remote forcing in the equatorial Atlantic ocean. *Geophysics* 12: 457-463.

Shannon, L. V. 1985. The Benguela ecosystem. Part 1: Evolution of the Benguela physical features and processes. *Ocean Marine Biology Annalar Review* 23: 105-182.

Shukla, J. 1990. Dynamical predictability of monthly means. *Journal of Atmospheric Science* 38: 2547-2572.

________1995. On the initiation and persistence of Sahel drought. *National Research Council* 3: 44-48.

Shukla, J. and Mintz, Y. 1982. Influence of land surface-surface evapotranspiration on the earth's climate. *Science* 215: 1498-1501.

Shukla, J. and Paolino, D. A. 1983. The southern oscillation and long range forecasting of the summer monsoon rainfall over India. *Monthly Weather Review* 111: 1830-1837.

Simmon, A. J. 1977. A note on the instability of the African Easterly Jet. *Jouurnal of Atmospheric physics* 34: 1670-1674.

Sivakumar, M.V.K. 1992. Empirical analysis of dry spells for agricultural applications in West Africa. *Journal of Climate* 5: 532-539.

________________ 2007. Interactions between climate and desertification. *Agricultural and Forest Meteorology* 142: 143-155.

Sling, J. M. and Gregory, S. 1996. Intraseasonal oscillation in 15 atmospheric general circulation models: results from an AMIP diagnostic subproject. *Climatological Dynamics* 12: 325-357.

Snijders, T.A.B. 1986. Interstation correlations and nonstationarity of Burkina Faso rainfall. *Journal of Climate and Applied Meteorology* 25: 524-531.

Soliman, K.H. 1958. On the Intertropical front and Intertropical convergence zone over Africa and adjacent oceans. *Symposium on monsoons of the world*. Meteorological office: New Delhi. 135-142.

Sonmez, F. K., Komuscu, A.U., Erkan, A. and Turgu, E. 2005. An analysis of spatial and Temporal dimension of drought vulnerability in turkey using The standardized precipitation index. *Natural Hazards* 35: 243-264.

Stedinger, J. R. and Baker, V. R. 1987. Surface water hydrology: historical and paleo-flood information. *Geophysics* 25: 119-124.

Strantz, D. 1978. Rain in Africa and the drought of recent years. *Applied Science Development* 11: 49-77.

Street, F. A. and Grove, A. T. 1976. Environmental and climatic implications of Quarternary lake –level fluctuations in Africa. *Nature* 261: 385-390.

Suarez, M. J. and Schopf, P. S. 1988. A delayed action oscillator for ENSO. *Journal of Atmospheric Science* 45: 3283-3287.

Subrahmanyam, V. P. 1967. Incidence and spread of continental drought. *WHO/IHD Report* 2.

Sud, Y. C. and Fennesy, M. 1982. A study of the influence of surface albedo on July circulation in semi-arid regions using the GLAS GCM. *Journal of Climate* 2: 105-125.

Sud, Y. C. and Molod. 1988. A GCM simulation study of the influence of Saharan evapotranspiration and surface albedo anomalies on July circulation and rainfall. *Monthly Weather Review* 116: 2388-2400.

Sutcliffe, V. J. 1987. The use of historical records in flood frequency analysis. *Journalof Hydrology* 96: 159-171.

Tabony, R. C. 1977. Drought classification and a study of drought at Kew. *Meteorological Magazine* 106: 1-9.

Tallaksen, L.M. and Lanen, H.A.J. 2004. Hydrological drought: Processes and estimation methods for streamflow and groundwater. *Development in Water Science* 48: 579-592.

Taleb, E. and Druyan, L. M. 2003. Relationships between rainfall and West African wave disturbances in station observations. *Int. Journal of Climatology* 23: 305-313.

Tanaka, M., Janocot, S. and Hulme, M. 1975. Recent African rainfall patterns. *Nature* 255: 201-203.

Tetzlaff, G and Adams, L. J. 1983. Present-day and early-Holocence evaporation of Lake Chad. *Variations in global water budget*. A. Perrot and K. Stigter. Eds. Reigdel. Dordrecht: 547-560.

Thompson, B. W. 1957. Some reflections on equatorial forecasting. East African Meteorological Department. *Technical Memo* 7: 1-13.

Thompson, R. M., Payne, S. W., Recker, E. E. and Reed, R. J.1979. Structure and properties of synoptic-scale wave disturbances in the intertropical convergence zone of the eastern Atlantic. *Journal of Atmospheric Science* 36: 53-72.

Thompson, S. 1999. *Hydrology for water Management*. A. A. Balkena. Ed. Rotterdam: The Netherlands.

Tickle, C. 1977. *Climate Change and World Affairs*. Centre for International affairs. Harvard University.

Torry, W. J. 1986. Economic development, drought and famines: some limitations of dependency explanation. *GeoJournal* 12: 5-8.

Tourre, B. and White, H. 1995. ENSO signals in global upper-ocean temperature. *Oceannography* 3:125-135.

Trenberth, K. E. 1976. Spatial and temporal variations in the southern oscillation. *Quarterly Journal of Royal Meteorological Society* 102: 639-653.

_______________1981. Signal versus noise in the Southern Oscillation. Monthly Weather Review 112: 326-332.

_______________1991. General characteristics of El-Nino/Southern Oscillation. *Teleconnections linking worldwide climate anomalies*. M. H. Glantz, R. W. Katz and N. Nicholls. Eds. Cambridge University Press. Cambridge: 13-42.

Trewartha, G. T. 1961. *The Earth's problem climates*. University of Wisconsin Press. 334Pp.

Tuckler, C. J., Harold, E. D. and Wilbur, N. W. 1991. Expansion and contraction of the Saharan desert from 1980 to 1990. *Science* 253.5017: 299-301.

Tyson, P. D. 1986. *Climate change and variability in South Africa*. Oxford University Press. Oxford.

Udo, R.K. 1971. *Geographical Regions of Nigeria*. Heineman: London.

Umar, A. T. 2010a: Rainfall variability in North-Western Nigeria and its implications for food security and livelihoods. *Nigerian Geographical Journal* 6.2: 151-162.

__________2010b. Recent trends and variability in the length of the growing season in northern Nigeria. *Journal of Meteorology and Climate Science* 8. 1: 40-52.

Van Apeldoorn, G. J. 1981. *Perspectives on drought and famine in Nigeria*. George Allen and Unwin. London. 184pp.

Van Heerden, J., Terblanche, D. E. and Schulze, G. C. 1988. The southern oscillation and South African summer rainfall. *Journal of Climatology* 8: 577-597.

Van Loon, H. and Rodgers, J. C. 1979. The seesaw in the winter temperatures between Greenland and the Northern Europe. Part 11: some oceanic and atmospheric effects in middle and higher latitudes. *Monthly Weather Review* 107: 509-515.

Van Loon, H. and Madden, R. 1981. The southern oscillation. Part 1: Global associations with pressure and temperature in northern winter. *Monthly Weather Review* 109: 1150-1162.

Van Loon, H. and Shea, D. J. 1985. The Southern Oscillation .Part viii: Precursors south of 15^0 S to the extremes of the oscillation. *Monthly Weather Review* 113: 2063-2074.

Van, V.S. and Francis, W. Z. 1999. *Statistical analysis in climate research*.Cambridge Universiy Press.

Verma, R.K., Subramaniam, K. and Dugam, S. S. 1985: Interannual and long-term variability of the summer smonsoon and its possible link with northern hemispheric surface air temperature. *Earth Planet Science* 94.3: 187-198.

Waliser, D. E., Lau, K. M. and Kim, J. 1999. The influence of coupled sea surface temperatures on Madden Jullian Oscillation: A model perturbation experiment. *Journal of Atmospheric Science* 56: 333-358.

Wallace, J. M. and Gutzler, D. S. 1981.Teleconnection in the geopotential height field during the northern hemisphere winter. *Monthly Weather Review* 109: 784-812.

Walker, H.O. 1958. The Monsoon in West Africa. *Ghana Meteorological Departmental Note* 5.

Walker, N. D. 1987. Interannual sea surface temperature variability and associated atmospheric forcing within the Benguela system. *South African Journal of Marine Science* 5: 121-132.

____________1990. Links between South African summer rainfall and temperature variability of the Agulhas and Benguela current systems. *Geophysics* 95: 3297-3319.

Walker, J. and Rowntree, P. R. 1977. The effect of soil moisture on circulation and rainfall in a tropical model. *Quarterly Journal of Royal Meteorological Society* 103: 29-46.

Walker, N. D., Taunton-Clark, J. and Pugh, J. 1984. Sea temperatures off the South African west coast as indicators of Benguela warm events. *South African Journal.of Marine Science* 80: 72-77.

Walker, G. T. 1923. Correlation in seasonal variations in weather, viii: A preliminary study of world weather. *Indian Meteorology* 24: 75-131.

___________1924. Correlation in seasonal variations of weather, ix: A further study of world weather. *Indian Meteorology* 24: 275-332.

___________1928. World weather 111. *Royal Meteorological Society* 2: 97-104.

Walker, G. T. and Bliss, E. W. 1930. World weather iv. *Royal Meteorological Society* 3:81-95.

____________and__________1932. World weather v. *Royal Meteorological Society* 4: 53-84.

____________and__________1937. World weather vi. *Royal Meteorological Society* 4: 119-139.

Ward, M. N. 1992. World -wide ocean-atmosphere surface fields in Sahelian wet and dry years using provisionally corrected surface wind data. *Journal of Climate* 5: 454-475.

Watts, M. 1983. *Silent violence: food, famine and peasantry in northern Nigeria.* University of California Press.

Watts, M. 1989. The agrarian question in Africa: Debating the crisis. *Progress in Human Geography* 13.1: 1-41.

Weare, B. C., Narato, A. R. and Newell, R. K. 1976. Emperical orthogonal analysis of Pacific sea surface temperatures. *J. Physical Oceanography* 6: 671-678.

Webster, P. J. 1981. Mechanisms determining the atmospheric response to sea surface temperature anomalies. Journal of Atmospheric Science 38: 554-571.

_____________1983. Large-scale structure of the tropical atmosphere. *Large-scale dynamical processes in the atmosphere*. B. J. Hopskin and R. P. Pearce. Eds. Academic Press.

Webster, P. J. and Chang, H. R. 1988. Equatorial energy accumulation and emanation regions: Impacts of a zonally varying basic state. *Journal of Atmospheric Science* 45: 803-829.

Webster, P. J. and Yang, M. 1992. Monsoons: Processes, predictability and the prospects for prediction. *J. Geophysical Research* 103: 14451-14510.

White, D. A. and Glantz, M. H. 1985: Understanding the drought phenomenon: the role of definitions. *Water International* 10: 111-120.

Wigley, T. M. L. 1984. The role of statistics in climate research. *Proceeings. of the 2nd International Meeting on Statistical Climatology*. 26th -30th September 1984. Lisbon: Portugal.

Willem, A. L. and Mason, S. J. 1999. Operational long-lead prediction of south African rainfall using canonical correlation analysis. *International Journal of Climatology* 19.10: 1073-1090.

Winstanley, D. 1973a. Recent rainfall trends in Africa, the Middle East and India. *Nature* 243 464-465.

_____________1973b: Rainfall pattern and general atmospheric circulation. *Nature* 245: 190-194.

World Meteorological Organisation. 1966. *Climate change*. WMO Technical note.

World Meteorological Oeganisation. 1971. Climatological normals (CLINO) for climate and climatic stations for the period 1931-1960. *WMO* 117, TP 52.

______1985. The global climate system: Autumn 1982-spring 1984. *World Climate Data Programme Contribution of Global Environmental Monitoring System* (GEMS), WMO, Geneva.

Wolter, K. 1989. Modes of tropical circulation, Southern Oscillation, and Sahel rainfall anomalies. *Journal of climate* 2: 149-172.

Woolnough, S. J., Slingo, J. M. and Hopskin, B. J. 2000. The relationship between convection and sea surface temperature on intraseasonal time scales. *Journal of Climatology* 13: 2086-2104.

Woo, M. K. and Tarhule, A. 1994. Streamflow droughts of northern Nigerian Rivers. *Journal of Hydrological Sciences* 39: 19-34.

Wu, H., Hayes, M. .J., Wilhite, D. and Svoboda, M.D. 2005. The effect of the length of record on the standardized precipitation index calculation. *International Journal of Climatology* 25: 505-520.

Wyrtki, K. 1975. El-Nino – the dynamic response of the equatorial Pacific Ocean to atmospheric forcimg. *Physical Oceanography* 9: 315-352.

Xue, Y., Liou, K. and Kasahara, A. 1990. Investigation of biogeophysical feedback on the African climate using a two dimensional model. Journal of Climate 3: 337-352.

Yarnal, B. and Diaz, H. F. 1986. Relationships between extremes of the Southern Oscillation and the winter climate of the Anglo-American Pacific coast. *Journal of Climatology* 6: 197-219.

Yasunari, T. 1985. Zonally propagating modes of the global east west circulation associated with the Southern Oscillation. *Quarterly Journal of Royal Meteorological Society* 63: 1013-1029.

___________1987a. Global structure of El-Nino/Sothern Oscillation. Part 1: El-Nino composites. *Quarterly Journal of Royal Meteorological Society* 65: 67-80.

___________1987b. Global structure of the El-Nino/Southern Oscillation. Part 11: Time evolution. *Quarterly Journal of Royal Meteorological Society* 65: 81-102.

Yeh, T. C., Wetherald, R. T. and Manabe, S. 1984. The effect of soil moisture on the short-term climate and hydrology change-a numerical experiment. *Monthly Weather Review* 112: 474-490.

Yin, Y., Levin, Z., Reisin, T. G. and Tzivion, S. 2000. The effects of giant cloud condensation nuclei on the development of precipitation in convective clouds-a numerical study. *Atmospheric Research* 53: 91-116.

Zebiak, S. E. 1993. Air-sea interaction in the equatorial Atlantic region. *Journal of Climate* 6: 1567-1586.

Zhang, C. 1993. Large-scale variability of atmospheric deep convection in relation to sea surface temperatures in the tropics. *Journal of Climate* 10: 1898-1913.

Zheng, X. 1997. The response of moist zonally sysmetric atmosphere to subtropical surface temperature perturbation. Quarterly Journal of Royal Meteorological Society 125: 1209-1226.

Zorita, E., Kharim, V. and Storch, V. 1992. The atmospheric circulation and sea surface temperature in the north Atlantic area in winter: their interactions and relevance for Iberia precipitation. *Journal of Climate* 5: 1087-1108.