SOME ASPECTS OF THE SPATIAL AND TEMPORAL CHARACTERISTICS OF DROUGHT IN NIGERIA: A STATISTICAL APPROACH

By

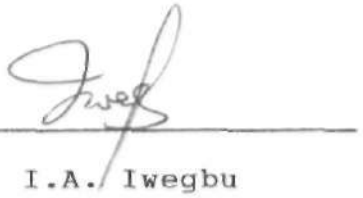
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A thesis submitted in conformity with the requirements
for the Degree of Doctor of Philosophy at
the Ahmadu Bello University
Zaria.

DEPARTMENT OF GEOGRAPHY
1993

DECLARATION

I declare that the information contained in this thesis is my work and that it has not been presented in any form for any award elsewhere. Information obtained from other litarary publications have been duly acknowledged.



I.A. Iwegbu

8.4.94

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CERTIFICATION

This thesis entitled "Some Aspects of the Spatial and Temporal Characteristics of Drought in Nigeria: A Statistical Approach" by I.A.Iwegbu meets the requirements of the regulations governing the award of the degree of Ph.D. Geography (Climatology) in the Department of Geography, Ahmadu Bello University, Zaria and is approved for its contribution to knowledge and literary presentation.

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DEDICATION

Dedicated to my daughters, Ego,
Natasha and Nkechi.

ABSTRACT

Mean rainy season (April-October) precipitation series for fourty one stations in Nigeria for a long period of years, are used to derive a Bhalme and Mooley-type drought index for the region. Various statistical analysis indicate that precipitation series are more variable Over the Sahelian, eastern Midland and western sub-sectors of the Guinea-Savanna and Coastal zones. The rainfall series for the Sahelian and northern Midland zones indicate a significant trend towards increasing aridity, especially in the last 30 years.

Map-pattern correlation method stratifies 46 per cent of the eighty two years, and produces comparatively good results as the more sophisticated eigenvector method, six moisture amomaly patterns, are identified by map-pattern correlation method indicating distinct differences between areas north of 10°N and the rest of the country. Maps of the mean seasonal areal coverage of drought reveal remarkable changes in size and position from year to year. Drought does not cover the entire country and no particular areas were persistently affected until the 1970s and 1980s. Since the beginning of the 1970s areas affected by drought have increased in extent with a consequent decrease in near normal and wet conditions particularly in the Sahelian and the Midland zones. The 1970s and 1980s were characterized by exceptionally high frequency of drought conditions with a presence of consecutive years of large-scale droughts since 1982, confirming persistence of drought in Nigeria.

All stations are subjected to Non-Integer method of spectral analysis and low-pass filter method in order to identify any trends and regular periodicites in the drought index. The low-pass filter results confirm negative trend for the Sahelian and Midland zones (from the mid 1950s and late 1960s respectively), indicating increasing aridity. The drought spectrum as indicated by the Non-integer method is dominated by peaks in the 2.1-7.9 frequency range (Coastal and western Guinea-Savanna zones) and longer frequencies with peaks between 11.0-11.9 and 13.0-13.9 years in the Sahelian and the Coastal zones. Based on the number of stations with significant quasi-11-year oscillations and their spatial distribution a linear relationship between drought index fluctuation and 11-year sunspot cycle may not be realistic, as it is often portrayed in the literature.

In general, drought in Nigeria is a complex random phenomenon, generated by non-linear atmospheric processes sporadic in spatial occurrence, making its prediction difficult.

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CHAPTER 1

INTRODUCTION

1.1 Motivation for This Study

The 1980s witnessed severe global climatic instability, which is now regarded as a possible response of the climatic system to an enhanced greenhouse effect caused by various human activities. Although atmospheric behaviour fluctuates naturally on different time scales, the present climatic fluctuation is rated as more rapid and destructive than has been experienced before. This fluctuation is characterised by an increased probability of the occurrence of such extreme events as summer heat waves, droughts and floods. The estimated increase of global mean surface air temperature by 0.3°C to 0.6°C over the last one hundred years, with the five warmest years occurring in the 1980s, could indicate that the global climatic system today is being pushed towards a new state (IPCC, 1990).

Already there are indications of the possible effects of global warming. In 1987, for example, an unprecedented tropical storm tore across Great Britain, uprooting century-old trees, disrupting power lines and destroying homes. In the same year a heat wave, as a result of an exceptionally hot and dry summer in Europe, killed 1,200 people in Greece. Hurricanes Gilbert (1988) and Hugo (1989) were particularly devastating in the Caribbean and Tuvalu, a small low-lying island group in the Pacific, was hit by twenty one tropical

cyclones in 1989 compared with an annual average of four. In August 1988, the Sudan, which had been affected by drought for more than two years, received its average yearly rainfall in three days. The resulting flood washed away topsoil and left more than 1.5 million people homeless (Grubb, 1989; Mintzer, 1990; Sinclair, 1991; Oladipo, 1991). The winter of 1991-1992 was recorded in the United States as the warmest in the last 97 years, while in the Middle East it was the wettest and coldest in 40 years with an unusual amount of snow in Amman, Damascus and Jerusalem. The same year the continent of Africa suffered from its worst drought in fifty years and floods affected Hong Kong, parts of Australia, southern Brazil, California and some West Europian cities (Time magazine, 15th June, 1992).

The above examples indicate the need to understand the degree of climatic variability over a period of time. Nigeria is no exception. Rainfall in Nigeria is highly variable from year to year and in recent years frequent dry spells culminated in severe and widespread droughts, capable of large-scale destruction of plants, animals and human life, such as the well known 1972-1973 drought. Although, alternating wet and dry years have always been observed in the region (e.g. Adefolalu, 1973; Oguntoyinbo and Richards, 1977; Ayoade, 1977a) the persistence of droughts since 1972-1973 suggests that a dangerous trend in mean conditions is taking place. These droughts have taken catastrophic dimensions in some years. For example, the 1972-1973 drought

in the Sahel reportedly claimed between 100,000 and 250,000 lives (Nicholson, 1989b). During this drought over 10.4 million people faced famine situation and estimated losses of guinea corn and millet were 1.675 million tonns in the worst affected areas of Nigeria (van Apeldoorn, 1981). An estimate of 3.5 million heads of cattle (approximately 25 per cent of the total) perished (Grainger, 1990) thus, ranking this drought among the greatest tragedies of the twentieth century. In 1983 and 1987 droughts returned and during the 1987 growing season about five million metric tonnes of grains valued at about $400 million were lost to drought (*New Nigerian, 12th January, 1988*).

These repeated droughts have contributed in no small way to the inability of semi-arid zone vegetation to cope with the rate of exploitation from increasing population. The detrimental impacts of drought on a wide spectrum of human activities and environmental processes can not be over-emphasized: desertification, damage to plants, animals and soil resources, depletion of ground water reserves, impaired water and air quality and increasing fire hazard are just a few of the environmental impacts of drought. The socio-economic impacts range from disruption of livelihood, migration and health related problems (due to the poor quality of water, poverty and hunger), to serious financial losses by farmers and cattle owners.

Despite its recorded history and the hardships imposed by drought, only a few studies have examined the nature,

areal coverage, duration and trends of drought occurrences in Nigeria. The present study, therefore, will focus investigation on these important aspects of drought.

1.2 Aims and Objectives of The Study

Rainfall is the most important climatic element affecting agriculture in Nigeria. Unfortunately, rainfall in Nigeria is extremely variable resulting in periodic droughts in various parts of the country. Drought, as an inevitable part of the climate of Nigeria, affects not only the arid and semi-arid areas of the north but also the wetter southern areas. While the arid and semi-arid areas of northern Nigeria are, by nature, prone to recurrent and persistent periods of drought separated by wet years, significant reduction in the amount of rainfall normally received in the south can also have a devastating effect.

Therefore, the purpose of this study is to describe the spatial and temporal scales of drought occurrence in Nigeria. The specific objectives are:

(i) To describe April-October precipitation climatology in the country.

(ii) To calculate and assess the performance of van Rooy (1965), Tabony (1977) and Bhalme and Mooley (1980) metheorological indices in depicting the intensity and severity of drought.

(iii) To examine the extent of spatial coherence in the occurrence of drought.

(iv) To detect any systematic and regular pattern in the temporal distribution of dry and wet periods.

Based on the foregoing objectives, two major hypotheses are formulated:

(i) The region of study does not consistently respond as a unique entity to any drought-related climatic mechanism.

(ii) Drought in Nigeria is a complex phenomenon, produced by mesoscale processes, that may persist to prolong its occurrence and expand its areal coverage.

1.3 Significance of The Study

Precipitation is the major source of available water in Nigeria and a valuable resource to be adequately managed, particularly in the drought prone areas of the country. Hence any variations in the timing or amount of rainfall can cause serious problems for rainfed agriculture and other economic activities. Although it is impossible to avert natural drought, it is nonetheless possible to minimise its effects by making adequate long-range mitigating plans. These plans must, inevitably, be based on a concrete knowledge of drought characteristics in the region of study. Therefore, a comprehensive description of spatial and temporal aspects of drought would provide a valuable input into the possible anti-drought strategies such as drought monitoring, disaster preparedness and management. Mapping the spatial scales of drought would help in identifying the particular drought-prone areas,

which require special attention. In this respect, knowledge gained may be useful for selecting appropriate crop and plant types for cultivation in a particular area.

In addition, a good understanding of the temporal chang in drought occurrence would be significant for understanding the predictability of the phenomenon. Although significant periodicities (e.g. 2, 8, 15, 27, 32 and 45 years) have been identified for the Sahel region (Bunting et al., 1976; Kraus, 1977; Nicholson, 1985; Adejuwon, 1991), it is still difficult to evaluate the reliability of these periods for Nigeria. Therefore, a comprehensive assessment of the temporal characteristics of drought in Nigeria, based on long-term precipitation records, will aid in better documentation of the scales of drought in the region.

Moreover, understanding the characteristics of drought in Nigeria would be a useful contribution to a future regional approach to the problem of drought. The concern about the future of food and water supply, especially in northern Nigeria, justifies continued interest in drought study. To analyze and explain the nature of drought will increase current knowledge about its variability, so that available water resources may be more meaningfully managed.

1.4 Scope of The Study

The study focuses on the climatological characteristics of April-October droughts in Nigera, particularly their spatial and temporal aspects. The major questions concern

the extent of drought and trends in drought patterns. These have received little attention in the past. Research dealing with some aspects of drought in Nigeria has been carried out by Oguntoyinbo and Richards (1977), Adefolalu (1986a,b) and Olaniran (1991). Oladipo (1989) investigated the spatial coherence of drought in northern Nigeria for the period 1916-1983 and established that there are remarkable seasonal changes in the drought patterns in northern Nigeria, with no consistent recurrent spatial patterns in the moisture anomalies.

The present study investigates the areal extent of drought for the whole of Nigeria. This is important because the whole of the country may not respond as an entity to any single climatic forcing. Secondly, for a scientific planning of agriculture and water resources it is important to know how often droughts occur and for how long they persist.

In order to assess these problems, some specific questions must be answered:

(i) What is the spatial distribution of seasonal (April-October) drought occurrences?

(ii) Are there any temporal and spatial coherences of drought in Nigeria?

(iii) Are droughts periodic? If so, on what time scale?

(iv) Are droughts persistent?

1.5 Organization of The Study

An important initial step in the investigation of drought is its accurate definition. Therefore, Chapter two reviews

both the historical and recent evidence of drought events. In the same chapter some important causal mechanisms of drought occurrence are discussed.

For tropics in general and Nigeria in particular, precipitation is the most important climatic input into any drought study. Therefore, detailed discussion of the precipitation data used in this study is followed by simple statistical tests and discussion of secular variation of April-October precipitation climatology in Nigeria. These discussions and tests constitute Chapter three.

Chapter four reviews the performance of some metheorological drought indices. In the same chapter, an equivalent of the Bhalme and Mooley (1980) drought index (BMDI) is derived for Nigeria and its performance in representing drought conditions is compared with the performance of the van Rooy (1965) rainfall anomaly index (RAI) and the Tabony (1977) drought index.

Chapter five discusses the spatial characteristics of BMDI using a statistical map-pattern correlation approach over time and space. The spatial patterns of BMDI are presented in order to determine the spatial and temporal coherence, if any, of droughts in Nigeria.

In Chapter six the low-pass filter technique and the non-integer technique of spectral analysis are applied to the BMDI series with the objective of identifying marked fluctuations and significant periodicities.

Chapter seven presents the findings of the study within the context of our present understanding of the drought phenomenon and its predictability.

CHAPTER 2

REVIEW OF LITERATURE

2.1 Introduction

The persistence of an unusually long run of dry years in semi-arid sub-tropical Africa since the late 1960's, has aroused the interest of the scientific community and prompted numerous research papers dealing with different aspects of drought. In spite of extensive literature, drought remains a scientific problem. This chapter provides a review of the literature on three important aspects of drought study. In the first section of this chapter the problems of drought definitions are discussed. Next, in order to put drought in its proper perspective, its historical trend over the region is reviewed, using evidence from the Sahel region as a whole. Lastly, the causal mechanisms of drought that have been proposed are discussed.

2.2 Drought Definitions

Whenever drought is discussed, the problem of its definition arises. What is drought? Although it has been with human societies since antiquity, there is still no universally acceptable definition of this elusive phenomenon. According to Yevjevich (1967) this lack of a generally accepted precise and objective definition of drought has been one of the principal obstacles to its investigation. Yet, a precise and objective definition of

drought in a specific situation is important, in order to eliminate confusion, wrong responses and to work out appropriate management strategies. Many reasons make drought very difficult to define. Firstly, drought is a complex phenomenon and a result of many factors acting on, and interacting with, the environment. Secondly, unlike other natural disasters (e.g. flood) it has neither a distinct start, nor finish. Recognizable after a period of time, it can be interrupted by one, two, or more wetter periods, giving the impression that it is over (Oladipo, 1985). Drought is referred to as a "creeping phenomenon", making an accurate prediction of either its onset, or its end, a difficult task. As Tannehill (1947), for example, noted:

> "We scarcely know a drought when we see one. We welcome the first clear day after a rainy spell. Rainless days continue for a time and we are pleased to have a long spell of such fine weather. It keeps on and we are a little worried. A few days more and we are in trouble. The first rainless day in a spell of fine weather contributes as much to the drought as the last, but no one knows how serious it will be until the last day is gone and the rains have come again .... We are not sure about it until the crops have withered and died"

There are other features of drought which make its definition difficult. Drought can occur in any part of the world, even in regions of copious rainfall. It is a condition relative to some long-term condition of balance between rainfall and evapotranspiration in a particular area, perceived as "normal" (Wilhite and Glantz, 1987). However, the meaning of "normal" will

be relative to the climate of an area and will therefore vary from place to place. For example, in a humid country like Britain, 15 days, none of which received as much as 0.25mm of rainfall, constitute drought (British Rainfall Organisation, 1936), while in persistently drier areas such as the Sahelian zone of West Africa droughts are only recognized after a year or more without adequate rain (Oladipo, 1985). Also, drought severity will depend not only on the duration, intensity and geographical extent of a specific drought episode, but also on the demands made by human activities and vegetation on a region's water supply (Wilhite and Glantz, 1987).

There is a distinction between the concept of drought and that of aridity. Aridity is a permanent climatic feature of a region, resulting from consistently low precipitation. Consequently, small communities living in those areas adapt themselves to the conditions of permanently low precipitation. Drought, on the other hand is a temporary feature of the climate of the region, when less than "normal" (adequate) precipitation occurs for a period of time. If the departure from the normal conditions is significant and prolonged, the stress on biological environments and human society could be disastrous.

The term "drought" is generally associated with the idea of a deficit in the supply of moisture for some

specific purpose. It is concerned with the description of precipitation, stream flow, run-off, evapotranspiration, groundwater levels, water supply and dryness of air over a season, a year, or several years and also of the spatial extent of the phenomenon. Its main characteristic is a decrease of water availability in a particular period over a particular area rather than a general decrease in water availability (Ojo and Oyebande, 1985).

Problems of drought definition were discussed by Subrahanyam (1967), Hounam et al. (1975), Matthai (1979), Dracup et al. (1980), Oladipo (1985), Ojo and Oyebande (1985) and Wilhite and Glantz (1987) among others. The general consensus is that the problem of drought definition is partly caused by the conflicting concepts held by a variety of academic fields of study and their specific interests. For example, hydrologists are concerned with drought in the context of a period of below normal stream flow and depleted reservoir storage, while meteorologists are concerned with drought in the context of a period of below normal rainfall. Agriculturists, on the other hand, are concerned with drought in the context of a period during which soil moisture is insufficient to support crops. The economist is concerned with drought as a period of low water supply, which affects society's productive and consumptive activities. The result of the diverse views about drought is over 150 definitions. All these definitions can be grouped

into four categories: (1) meteorological, (2) hydrological, (3) agricultural and (4) socioeconomic.

Meteorological drought occurs when precipitation is significantly below normal for a period of time. Most objective definitions of meteorological drought are site specific and developed for application in a particular country (Wilhite and Glantz, 1987). For example, drought in Libya is defined as occurring when annual rainfall is less than 180mm (Hounam et al. 1975).In India actual seasonal rainfall must be deficient by more than twice the mean deviation for drought to be recognized (Ramdas, 1950). As already stated, in Britain an absolute drought is defined as a period of at least 15 consecutive days with rainfall of less than 0.25mm per day (Hounam et al., 1975). However, a period of 29 days with mean rainfall not exceeding 0.25mm per day is regarded as a partial drought in the same country, and a dry spell is a period of 15 consecutive days none of which has received as much as 1.0mm of rain. According to the U.S. Weather Service a period of 21 days, or longer, with less than 30% of the average precipitation for the time and place constitute drought occurrence, with extreme drought occurring when rainfall fails to reach 10% of normal for 21 days,or more. Probably the best known meteorological drought definition is that proposed by W.C.Palmer (1965). He described drought as "an interval of time, generally of the order of months or

years in duration, during which the actual moisture supply at a given place rather consistently falls short *of climatologically appropriate moisture supply*".

Some meteorological definitions of drought may also include atmospheric parameters other than precipitation. Condra (1944) referred to atmospheric drought as a "period of strong wind, low precipitation, high temperature and usually low relative humidity".

Agricultural drought occurs during a period of time when the requirement of plants are not met either in terms of water demand to survive by avoiding wilting, or to produce maximum yield (Ojo and Oyebande, 1985). It does not begin on the day rainfall ceases, but when vegetation can no longer obtain water from the soil rapidly enough to replace that lost to the air by transpiration (Chin, 1977). Therefore, the timing of the rainfall and the length of the dry period between rainfall are the most important factors of agricultural drought.

Definitions of agricultural drought link various characteristics of meteorological drought to agricultural impacts. Hounam et al. (1975) list 59 indices of agricultural drought, based either on rainfall, or on the balance between moisture supply and demand. The thresholds for agricultural drought depend on demand, varies with time and may be affected by changes in the rural economy, such as a shift to different crop types.

Hydrological drought refers to periods of below normal stream flow and depleted reservoir storage. Ojo and Oyebande (1985) list six types of hydrological drought based upon variations in duration, season of the year, or severity, produced by UNESCO-WMO, 1985:

i) A three week to three month runoff deficit during the period of germination and plant growth. This could be catastrophic for farming that is dependent upon irrigation water from rivers.

ii) A minimum discharge significantly lower, or more prolonged than the normal minimum, but does not affect the germination period and thus is of less consequence for agriculture.

iii) A significant deficit in the total annual runoff which affects hydropower production and irrigation from large reservoirs.

iv) A below normal annual high water level of the river. This may create the need for pumping for irrigation.

v) Drought extending over several consecutive years. Discharge remains below a low threshold, or rivers dry up completely and remain dry for a very long time.

vi) A significant natural depletion of aquifers (groundwater storage). This is often complicated by over-exploitation of groundwater during drought.

As in the case of agricultural drought, hydrological drought is concerned with impacts and effects of the event rather than its causes. Other objective definitions of hydrological drought are given by Dracup et al. (1980), Linsley et al. (1975) and Wipple (1966) among others.

Economic drought is viewed by the socioeconomists from the point of view of different areas of human activities affected. Definitions of drought by social scientists and economists inevitably incorporate features of meteorological, agricultural and hydrological drought. Regarding drought as a supply and demand phenomenon, Sandford (1978) defined socioeconomic drought as "a rainfall-induced shortage of some economic goods, brought about by inadequate, or badly timed rainfall". Increasing demands by society in the form of agricultural, or industrial, activities can create drought situation, or make an existing one worse.

The foregoing summary of definitions demonstrates some of the problems in deriving an objective definition of drought. Drought is indeed an undefinable entity, but a prime mover which has attributes (Ojo and Oyebande, 1985). Agricultural drought may exist due to poor distribution of precipitation, before it is evident to the meteorologist, or hydrologist, because the seasonal totals could still be statistically speaking normal. On the other hand, agricultural drought may end at

least temporarily, by rainfall that replenishes soil moisture, but which is not sufficient enough to contribute to ground water, or stream flow. It is now accepted that a number of environmental factors, other than rain fall deficiency contribute to the overall intensity of drought. Meteorological drought is the focus of this research, since precipitation is the most important climatic input to water availability.

2.3 Climatic Fluctuations and Drought

Historically speaking droughts are not uncommon in the region of study. Even humid phases in the climatic history of the West African region contained many anomalous periods of dryness. Generally, information about climatic fluctuations and drought for arid and semi-arid northern hemisphere Africa from late Pleistocene (c. 35,000 years BP) is well known through studies of lake level fluctuations, radiocarbon dating of deposits, instrumental records and chronicles. Literature on conditions in more humid zones to the south, however, is not so plentiful due to the fact that conditions are more difficult to establish, interpret and precisely date (Nicholson, 1976). Therefore, most literature on past climatic variations and drought is concerned with the present day Sahel and Sudan zones of West Africa. The following information about the past climatic conditions in Nigeria draws heavily from studies made by

Livingstone (1975), Nicholson (1976, 1978, 1979, 1980a, b and 1981a), Grove (1972), van Apeldoorn (1981), Street - Perrot and Roberts (1983). An extensive review of the current knowledge of climatic history of West Africa is given by Oladipo (1991).

Generally four climatic episodes can be identified during the late - Pleistocene and Holocene, which are applicable to Nigeria. Summaries of these periods are presented below.

2.3.1 Period of the Late Pleistocene

The period of the late Pleistocene c. 30,000 to c. 20,000 BP was characterized by wetter conditions than present. Lakes were formed during the late Pleistocene throughout the southern Sahara and Sahel, in areas which today receive only about 50mm of rain per year (Butzer et al. 1972).

2.3.2 The Period of Aridity (c. 20,000 to 12,000 BP)

A period of c. 20,000 to 12,000 or 10,000 BP was characterized by extreme aridity and coincided with the last glacial maximum (Wisconsin or Weichsel) in midlatitudes. Termed as the Ogolien episode in West Africa, it was accompanied by the formation of dunes, ergs, aeolian deposits, diminishment and desiccation of lakes. Ogolien dunes reached as far south as 10°N in Nigeria, Cameroon and Sudan into regions which today receive as

much as 800mm of rain, representing nearly 300km southward expansion of the Sahara along its entire 4000km east-west extent (Grove, 1972; Talbot, 1980). In Nigeria the Ogolien dry period is represented by the Kano erg and deposits of pediment gravel (today found only in the Sahara and the Sahel) in the west and south-west. Lake Chad was at the time probably totally dry, as were numerous basins and depressions in Niger and other areas (Nicholson, 1980a). Figure 2.1 shows the lake level variations in Niger and Chad during the late Pleistocene and Holocene, with the dry Ogolien episode clearly visible. The explanation applied to this period is a shift towards the equator (with respect to the present) of climatic zones throughout tropical northern hemisphere Africa, due to intensification of the subtropical high over the Sahara and the weakening, or perhaps absence, of the summer "monsoon" rain system, so that Sahel and much of Sudan zones became desert. Rainforest retreated into the wetter areas such as coastal regions and highlands (Rognon and Williams, 1977; cited in Nicholson, 1980a).

2.3.3 Early Holocene Lacustrine Episode (c.12,000 to 8,000 BP) and the Neolithic Humid Phase (8,000 to c.4,000 BP)

At a time that varies slightly from place to place, but is commonly close to between 12,000 and 10,000 BP, the

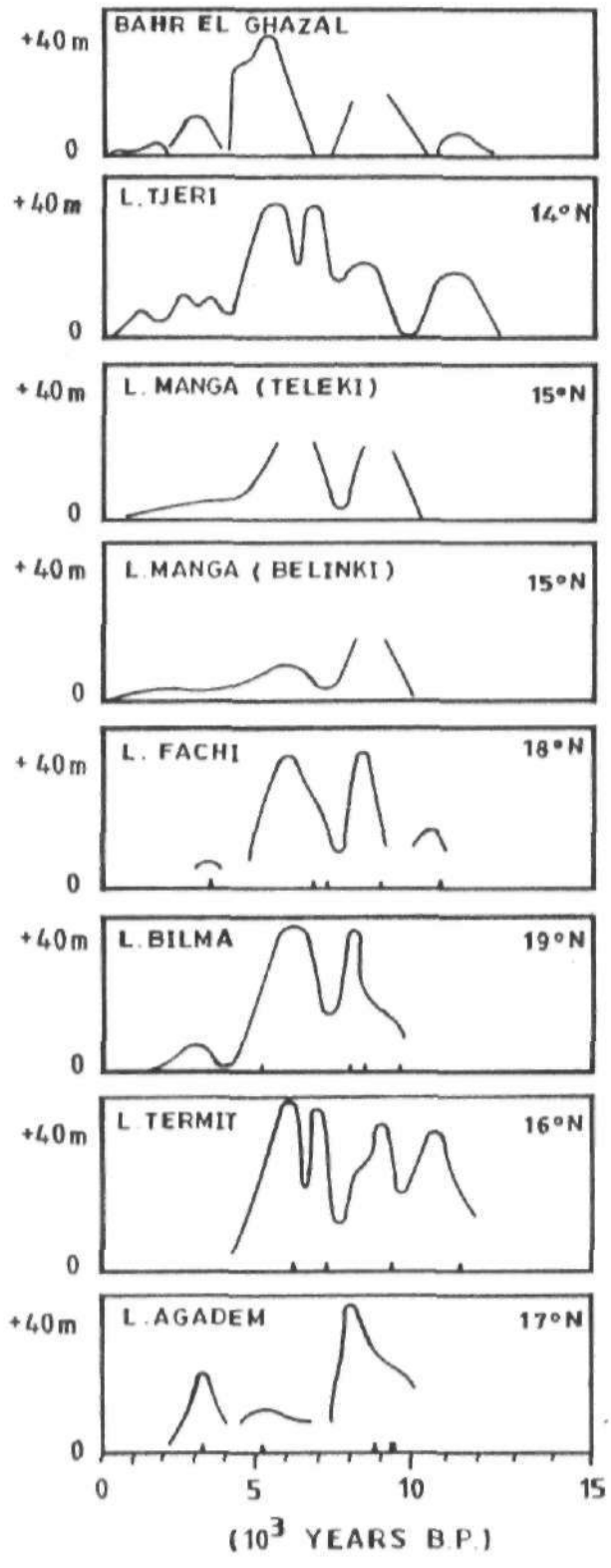


Fig. 2.1. Lake-level variations in Niger and Chad during Late-Pleistocene and Holocene
(From Servant, 1973, cited in Nicholson, 1980).

climate in Sahel became wet. A lacustrine episode beginning at about 12,000 BP and attaining its maximum from c. 10,000 to 8,000 BP left traces throughout the southern Sahara, Sahel and Sudan zones. Between 10,000 to 8,000 BP the surface of Lake Chad was 38 meters above its present level and the paleolake occupied an area of 320,000 sq. km (about sixteen times its present size). The lake's depth was about 340 meters in the Bodele depression in the north-east compared with the present mean depth of 3.8 meters (Servant, 1973). Rivers draining into the lake were much greater than today, their floods much higher and they built great deltas into the lake (Grove, 1972). This humid episode affected areas as far south as the Ivory Coast, Nigeria and Ghana, where Lake Bosumtwi reportedly rose some 40 meters above its present level (Talbot and Delibrias, 1977).

Though much of the period 8,000 to 4,000 is climatically similar to the period of wetter conditions from c. 12,000 to 8,000 BP., there are some drier oscillations within these latter four millenia, and the trend toward conditions resembling those of present began within this period. For example, the post-Ogolien dry interval lasting for a millenium towards c. 7,000 BP was evidenced by falls of lake and river levels and reworking of aeolian dunes in the Sahel. A drop in the level of Lake Chad took place between 8,000 to 7,000 BP with a minimum towards 4,500 BP. The dunes shaped during this period are generally oriented NNE-SSW, rather than NE-SW as in the Ogolien, indicating a change in wind

direction with respect to that time. This episode left traces in south-western Nigeria, where there is evidence of modification of the Ogolien gravel deposits (Michel, 1973; cited in Nicholson, 1980a). In general, the post-Ogolien dry oscillation in the Sahel was not as dry as the Ogolien, but was slightly drier than today. The dry episode towards 7,000 BP was followed by a rapid change to wetter conditions (second lacustrine phase), which occured from 6,500 to about 4,500 BP. Known as the Nouakchottien in West Africa and as Neolithic (or climatically Atlantic) on a much larger scale, it was characterized by contraction of the Saharan belt, higher lake levels and flourishing Neolithic civilisation, which spread over the Sahara. During this period (6,000 to 4,500 BP) climate and vegetation zones south of the Sahara were situated northward of their present position, but not as far northward as at about 10,000 to 8,000 BP (Nicholson, 1976). Lake Chad rose rapidly towards 6,000 BP maintaining its high level until c.4,000 BP, but it was smaller than during the humid phase in the early Holocene (Servant, 1973). By about 4,500 BP, however a change began towards the present day aridity in the Sahel.

2.3.4 The Last Four Millenia

Pre-Historical Times

Climatic fluctuations in the Sahel and Sudan within the last four thousand years have been minor compared with those of the late-Pleistocene to the Neolithic (see Nicholson, 1976, 1979 for details). However, a pronounced period of extreme aridity c.4,000 to 3,500 BP, followed by a humid phase can be clearly identified. Figure 2.2 shows a generalized curve for Lake Chad and the Niger lakes. In general, a marked trend towards increased aridity at the end of the Neolithic phase (c.5,000 to 4,000 BP) was followed in the region by a period of extreme aridity until about 3,500 BP. A return to humid conditions then commenced, but this humid phase ended towards 2,000 BP.

2.3.5 Historical Period

Historical information for climatic reconstruction for the Sahel and Sudan zones has been extensively documented by Nicholson (1976) and only a summary is presented below.

In general, there is good evidence that wetter conditions than today prevailed not only throughout the Sahel and Sudan zones but also as far south as the Guinea Coast until at least the 14th century. This resulted in the contraction of the Saharan belt along the southern margin. This humid phase began to decline after

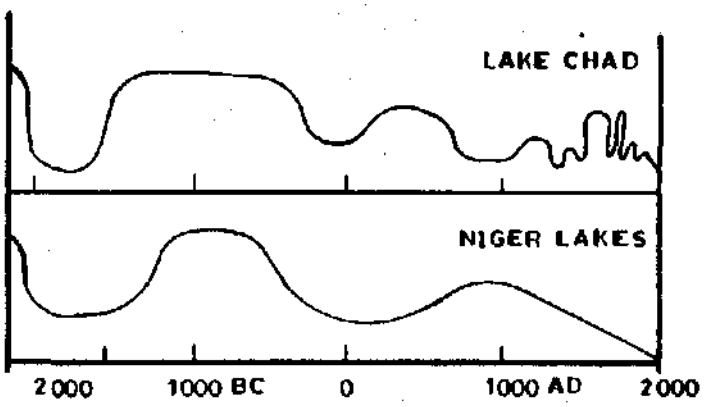


Fig. 2.2. Lake-level variations in Niger and Chad during the last 4000 years (From Nicholson, 1976)

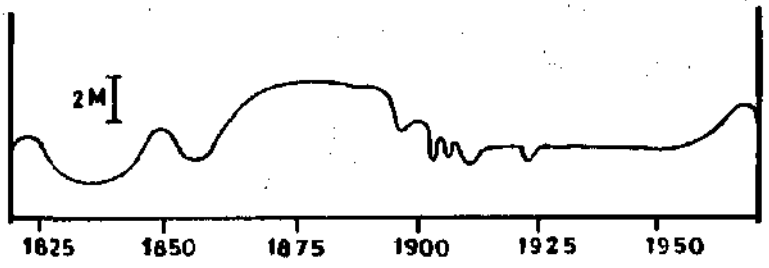


Fig. 2.3. Variation of the level of Lake Chad in 19th and 20th centuries
(From Maley , 1977 ; cited in Nicholson , 1978)

about the 14th century, with severe drought affecting Timbuktu c. 1445 - 1452 (see Nicholson, 1978). Another humid period occurred between the 16th and 18th centuries. Lake Chad, for instance, stood about 4 meters above the present mean level from about 1600 to some time in the 18th century, suggesting increased rainfall in the Chad basin (Maley, 1977). The occupation of Tibesti (1500 to 1700) and the presence of extensive African state - systems at the southern fringe of the Sahara are good indications of this humid period (Brooks, 1970; Plote, 1974).

This humid phase, which corresponds roughly to the so - called "Little Ice Age" in the mid - latitudes still features a series of anomalous drought periods starting from the 1680's. Extensive droughts affected the Sahel and Sudan zones during the periods c. 1681 - - 1687, c. 1738 - 1756 and c. 1828 - 1839, with more localized ones in the early 1760's, the 1770's and the 1790's. Great famines due to drought were reported at Bornu (north - eastern Nigeria) between 1657 - 1694. This drought reportedly affected areas as far south as Ghana and Benin Republic between 1682 and 1686. According to Nicholson (1978) this period seems to mark the beginning of a major change toward conditions resembling those of today. There is strong evidence that the drought of the 1680's covered the entire east - west zones of the Sahel. The period c. 1738 - 1756 features severe drought of

one or two decades in duration throughout the entire Sahel Sudan as well as the Guinea Coast. This drought reportedly killed 50 per cent of the population of Timbuktu and other parts of the Niger Bend and caused severe famine in the Benin Republic, Chad, Nigeria, Mali and Mauritania (Nicholson, 1978). Lovejoy and Baier (1975) suggest that famine at Bornu (north - eastern Nigeria) and Kano may have lasted for a few years within that period. There is further evidence of a lowering and partial desiccation of Lake Chad at this time (Maley, 1977).

The last decade of the 18th century was also characterized by severe drought that plagued Chad, Bornu and Kano. The surface of Lake Chad began to fall rapidly and in the years from 1795 the harmattan became more pronounced (Nicholson, 1989b; cited in Oladipo, 1991). The sharp decrease in lake levels since the beginning of the 19th century persisted until the 1930's and general drought conditions affected the entire Sahel - Sudan zone (Nicholson, 1989b). In northern Nigeria, the mid 19th century droughts are locally known as Darawa (1847) and Bamga - Bamga (1855) (Apeldoorn, 1981).

This aridity gave way to wetter conditions towards 1870, and from 1875 to 1895 rainfall was above the present normal. This manifested itself in increased discharge of the Senegal and Niger rivers and narrowing of the desert belt along both margins (Nicholson, 1978).

Figure 2.3 shows variation of the level of Lake Chad in the 19th and 20th centuries. Noteable in this curve is a tendency towards very high lake level in late 19th century, with a sharp decrease generally towards 1900, and a sharp rise again in the 1960's.

Thus, the trend since the beginning of the 20th century has been that of increased aridity. Fortunately, instrumental records are available since the beginning of this century. These will be used in Chapter 5 to describe the temporal changes of drought in Nigeria since the 1900's.

2.4 Explanations for Drought Occurrences

Various hypotheses have been outlined relating droughts to different scales of atmospheric behaviour and interaction (Hulme and Tosdevin, 1989) over the region of study. Obviously, the year to year precipitation variability and drought occurrence in Nigeria will depend on the controlling pressure systems and their effects on the strength and intensity of the rainy season. Therefore, causes of droughts in Nigeria could be better perceived through what is known about the systems producing West African rainfall in general.

2.4.1 Rainfall Producing Systems

Basic features of the lower atmosphere over West Africa during the rainy season are the thermally induced

low pressure cell over the Sahara with its axis located between latitudes 18°N and 22°N and the Subtropical Anticyclone over the south Atlantic. The gradient between these two pressure systems induces a warm, moist and convectively unstable tropical maritime air-mass (mT) from the south Atlantic to penetrate deep into West Africa, reaching latitudes 20°N - 22°N in August (Ojo, 1977; Kamara, 1986). The depth of the south-west monsoon decreases inland from the Guinea Coast from 3000 meters to 1000 meters (in July), forming a wedge under dry continental air (Figure 2.4). Consequently, the horizontal and vertical distribution of precipitable water decreases northwards and north-eastwards with nearly 40 to 50 per cent of the total vertically integrated moisture concentrated below 850mb (1500m) level (Anyadike, 1977). These two air-masses are separated by the quasi-frontal discontinuity zone called the Inter Tropical Discontinuity (ITD). Thus, the rainy season begins in a particular zone of Nigeria only after ITD and monsoon air have reached it, with the main rainfall zone some 500km behind the surface position of ITD. Five distinct weather zones associated with the ITD were adequately described by Ilesanmi (1970, 1972), Ojo (1977) and Olaniran (1991). Another important feature related to rainfall in the region is the position and intensities of two jet streams, known respectively as the African Easterly Jet (AEJ) and Easterly Tropical Jet (ETJ). AEJ originates

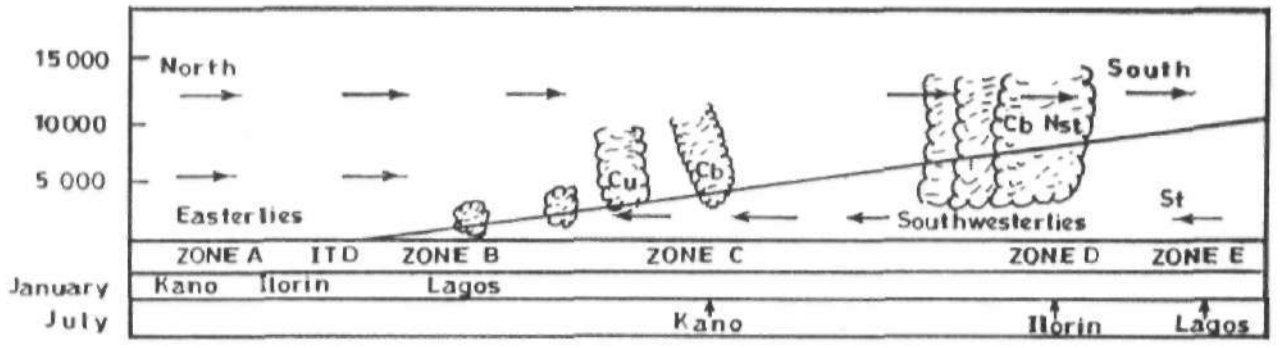


Fig. 2.4. The ITD and associated weather zones (From Ojo, 1977).

from the strong horizontal temperature gradient between the hot Saharan air and cooler equatorial air at the height of 3km. In the lower latitudes the jet stream appears twice during the year, once during the Early Squalls Season (April - May), and also during the Late Squalls (October - November). In central and northern parts of West Africa it appears only once coinciding with a single long squalls season from May to September (Kamara, 1986). Appearance of AEJ in the wind field coincides with periods of enhanced development of lower level synoptic disturbances such as easterly waves, line squalls and vortices. The ETJ originates over south east Asia and is located during the rainy season near latitude 10°N at the height of 12km. Air uplift with attendant widespread cloud development and rainfall occur on the right (northern) rear quadrant and front left (southern) quadrant of the jet axis (Koteswaram, 1958; Farmer and Wigley, 1985; Kamara, 1986). The position and strength of the two jet streams, therefore, have profound influence on small scale synoptic disturbances which play a major role in the production of rainfall. The most important of these disturbances are easterly waves, cyclonic vortices and linear thunderstorms. Easterly perturbations, originating between longitudes 15°E and 30°E on the southern side of the AEJ, move westwards across West Africa at a speed of 6° - 7° longitude per day. Within the wave troughs at the 850mb (1.5km)

level the wind is organized into a series of closed cyclonic vortices a hundred kilometers wide. Ahead of the wave trough there is strong surface convergence, cloud development, thunderstorms and rainfall, while to the rear there is subsidence and fine weather. A line squall is a 300km to 500km long belt of thunderstorms arranged in a north - south direction and propagating westwards against the surface wind at speeds exceeding 18m/s. Originating from west - central Africa and the highlands of eastern Cameroon, squall lines are responsible for over 70 percent of the total rainfall received in West Africa (Omotosho, 1985, 1990) and together with other small scale disturbances they account for the characteristic "patchiness", or "showery", nature of West African rainfall (Kamara, 1986). In general, therefore, drought will be due to either: (a) insufficient and /or late incursion of the ITD and monsoon air inland resulting in the absence of adequate moisture flux to the region, (b) change in atmospheric processes that promote the weakening of organized atmospheric rain generating systems (disturbances), (c) subsidence, (d) or a combination of these factors.

2.4.2 Mechanisms of Drought

Major works (e.g. Winstanley, 1973; Otterman, 1974; Charney, 1975; Kidson, 1977; Adedokun, 1978; Motha et al., 1980; Nicholson, 1983, 1986b, 1989a; Newell and Kidson, 1984; Dennett et al., 1985; Philander, 1986; Cadet and Nnoli, 1987; Folland and Owen, 1988; Adedoyin, 1989; Shinoda, 1989,1990; Druyan and Hastenrath, 1991 among others) have tried to explain drought in West Africa. Major hypotheses fall generally into four main categories. First are studies that explain Sahelian droughts as local manifestation of anomalies in the large scale atmospheric circulation. The second group of studies have concentrated on dynamic teleconnections between the dominant drought time scales and anomalies of global sea-surface temperatures (SST) and other tropical phenomena such as El Nino / Southern Oscillation (ENSO). The third group of studies relate drought to human land use feedback processes and an increase in greenhouse gases resulting from human activities. Fourthly, drought has been explained in terms of naturally occurring non-linear interactiosn in the Atmosphere.

Many studies (e.g. Winstanley, 1973; Bryson, 1973; Kraus, 1977; Lamb, 1978a,b; Motha et al., 1980; Hastenrath, 1984 and Philander, 1986) postulated that large-scale circulation anomalies are responsible for the suppression of the northward progression of the ITD that may result in drought. However, Nicholson (1981b, 1986b) postulated that the weakened intensity of the rainy season, independent of the ITD

position, is the most likely cause of drought in the Sahel. Also, some studies (e.g. Shinoda, 1989, 1990) have indicated that a combination of the two factors shoudl be considered as a plausible causal mechanism of West African rainfall anomalies.

Several workers have examined upper air data to identify and describe the differences in circulation between wet and dry episodes in the Sahel. Shupelius (1976), Kraus (1977), Kidson (1977), Kanamitsu and Krishnamurti (1978), Adedokun (1978), Kruger and Winston (1975), Newell and Kidson (1984), Hulme and Tosdevin (1989) among others showed that, in drier periods, the ETJ is weaker, while the AEJ is stronger than normal. Unusually strong AEJ would inhibit onshore flows of moisture from the Atlantic, allowing the dry Saharan air to spread over most of the region. Weaker ETJ would reduce the vertical wind shear over West Africa and consequently reduce the number of rain producing disturbances. Newell and Kidson (1984), in addition, found that the unusually strong AEJ expanded equatorward during August months of the dry 1970-1973 period owing to warmer temperatures in the lower troposphere at 10^{o} - 15^{o}N. This observation led to the speculation that lower rates of evapotranspiration from rain deprived soils may account for the elevated temperatures. Recent model results by Druyan and Hastenrath (1991) support this view.

The role of moisture advection was discussed by Shinoda (1986), Druyan (1987) and Cadet and Nnoli (1987). They attributed West African droughts to scarce water supply to the region. Conversely, Lamb (1983) established that the sub - Saharan drought of 1968 - 1973 was not associated with the northward supply of unusually low moisture across the West African coast from the tropical Atlantic. A plausible explanation of Lamb's finding is weak moisture convergence caused by a shallow monsoon trough over the western Sahel. This, according to Druyan and Koster (1989), would cause water vapour to drift eastward without building up humidities over the western and central Sahel sufficient for adequate rainfall.

The second group of researchers studied the teleconnections between the Sahel rainfall and the sea surface temperature (SST) anomalies and the El Nino / Southern Oscillation (ENSO)[1]. Lamb (1978a,b) and Lough (1986) for example, provided evidence that variations in tropical Atlantic SSTs may contribute to rainfall anomalies in the Sahel. Many studies have documented the relationship between SST anomalies in the Pacific and

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[1]Pronounced barometric pressure fluctuation between the Indian - Australian and South American regions known as the Southern Oscillation (SO) and aperiodic and widespread equatorial Pacific Ocean warming known as the El Nino (EN) have been statistically associated with drought. The term ENSO is used to depict an integrated aspect of both the El Nino and the Southern Oscillation.

Atlantic oceans and atmospheric circulation. Rasmusson and Wallace (1983), Glantz (1984), Hastenrath (1984,1990), Palmer (1986), Folland et al. (1986,1989), Druyan (1987, 1988), Kayano et al. (1989), Owen and Folland (1988), Parker et al. (1988), Semazzi et al. (1988), Adedoyin (1989), Owen and Ward (1989), Druyan and Hastenrath (1991) have established that persistently wet and dry periods in the Sahel are strongly related to SST anomalies on a near global scale. Adedoyin (1989) examined the relationship between global SST anomalies, ENSO and rainfall data from 1951 to 1985 for seven stations in northern Nigeria and established, in agreement with previous findings by Hastenrath (1984) and Lough (1986), that: (i) there is a statistically significant relationship between Sahel drought and a persistent assymetrical pattern of global SST characterized by colder than normal conditions in the North Atlantic and warmer than normal SST in the Southern Atlantic and (ii) the Pacific El Nino does not appreciably affect Sahelian rains in the first year of its occurrence, but there is a considerable reduction in the number of squalls and rain days in its second year. Adedoyin (1989) in agreement with Desbois et al. (1988), Klaus Wolter (1989), Nicholson (1989a), Nicholson and Nyenzi (1990) and Ropelewski (1990) concluded that the relationships between rainfall anomalies and ENSO are not as statistically significant as with SST anomalies. For example, it is noted that although

several major El Nino's coincided with dry periods in the Sahel (e.g. 1972, 1982), other El Nino events (e.g. 1957) have coincided with above normal rainfall in the region (see Druvan, 1989; Nicholson, 1989a). In summary, many questions still remain about the empirical links between SST anomalies, ENSO and Sahelian rainfall. Among these, as pointed out by Oladipo (1990) are whether SSTs may lag and therefore possibly respond to rainfall; whether SSTs may be more important as reinforcing factors for interannual rainfall anomalies than for persistant Sahelian drought; and whether rainfall and SSTs may respond to common atmospheric forcing which may be the more direct factor in rainfall variability. Tropical teleconnections (SST and ENSO) can only indicate the magnitude of ocean - atmosphere interactions in tropical circulation regimes, that may be resposible for observed rainfall anomalies without being able to explain their drought related dynamics.

A number of researchers have attributed the persistent Sahelian droughts to land surface feedback processes whereby human-induced changes in ground surface conditions inhibit convective rainfall (see review by Nicholson, 1988). Major papers in this field of research are by Otterman (1974), Charney (1975), Shnell (1976), MacLeod (1976), Walker and Rowntree (1977), Sud and Fennessy (1982), Otterman and Tucker (1985), Cunnington and Rowntree (1986) among others. Otterman (1974) proposed

that baring of high albedo soils by overgrazing, deforestation, or drought would increase the reflectivity of the land, producing radiative cooling of the air, thus, reducing convective activity and precipitation. One of the most significant drought causal hypothesis, related to human landuse, is due to Charney (1975). His biogeophysical feedback hypothesis was distinct from Otterman's. He reasoned that higher reflectivity (albedo) of bared soils leads to reduction in the net radiation balance at the top of the troposphere. Maintainance of thermal equilibrium over this region therefore, would require some form of increased heat supply. This was provided through subsidence of air aloft which warmed the lower layers of the troposphere through adiabatic heating. Such subsidence increased the upper level overturning of subtropical air, suppressed vertical uplift and thus, suppressed a key process in the generation of Sahelian rainfall. Thus, an initial change in land cover (increased albedo, or reduced soil moisture storage) is amplified through atmospheric processes which in turn suppress future rainfall. This increases moisture stress on vegetation and lowers soil moisture levels. Drought becomes self - perpetuating without the need to invoke atmospheric perturbation external to the region, although such external control may be necessary eventually to reverse the feedback.

Charney's mechanism came in for criticism mainly

because of the omission of soil moisture variations and of any discussion of latent /sensible heat ratios. Also, the magnitude of regional albedo increase from 14 to 35 percent proposed by Charney (1975) has not found any empirical support. Wendler and Eaton (1983) obtained only 10 percent difference between ground and protected land in the northern Sahel (southern Tunisia). Similarly, Gornitz (1985) found that the anthropogenic albedo increa over West Africa during the last 100 years was only 4 percent. Consequently, studies with dynamic models (Walker and Rowntree, 1977; Cunnington and Rowntree, 1986 Rind et al. 1991) have shown that rainfall anomalies are more sensitive to soil moisture and /or evaporation perturbations, than albedo changes. Many problems with the General Circulation Models (GCM) add to the problem (see Oladipo, 1990 for details). Parameterization of land surface processes in models is highly simplified and in many circumstances unrealistic and exaggerated surface parameters are necessary to obtain a significant drought signal. Moreover, the occurrence of a surface induced persistent drought suggests that the surface characteristics responsible for the feedback must be related to the rainfall in the previous year. This "memory" effect has not been found to be true for the observed interannual changes in either soil moisture, or albedo. As noted by Hulme (1989) and Oladipo (1990), the fact that rains often return after a period of drought indicates

that other atmospheric mechanisms may override the feedback effect of albedo and soil moisture changes.

MacLeod (1976) in his investigation of 1974 drought observed that large amounts of dust in the Sahelian atmosphere, from denuded bare soil, produce an isothermal condition in the south-west monsoon, which is the source of precipitation in the region. Such an isothermal character of the monsoon air will supress any significant uplift, cumulus cloud formation and rainfall. Furthermore, the pressure of the dusty atmosphere would be reduced relative to that of non-dusty air because of higher temperatures aloft. Consequently, the north-east harmattan winds will flow further south and the south-wes monsoon more gently north, with ITD located in a more southerly position to produce sporadic and scanty rainfall. Unlike the albedo /soil moisture feedbacks, this hypothesis has not been extensively modelled. Coakley and Cess (1985) have nonetheless, demonstrated that aerosols, through enhanced heating of the atmosphere, could produce rainfall and provide an additional feedback mechanism whereby drought is self-perpetuating.

Some recent studies (see review by Mabbutt, 1989) have suggested that anthropogenic changes in the concentration of greenhouse gases (e.g. carbon dioxide) in the atmosphere may be partly responsible for recent droughts in semi-arid areas. The unusually warm weather condi-

tions and frequent droughts in the 1980's over the globe, provide some empirical support. It is however, too early to be certain about the consequences of greenhouse warming because behaviour of feedbacks may amplify, or damp, the direct radiative effects of radiatively active trace gases on the climate system.

Some researchers (e.g. Gordon and Hunt, 1987; Hunt and Gordon, 1988) proposed a quite different causal drought mechanism (see review in Oladipo, 1990). They hypothesise that droughts may be naturally occurring random phenomena arising from local nonlinearly induced perturbations in the atmosphere that can result in abnormal circulation patterns. These abnormal patterns can result in areas of deficient rainfall alternating with those of wet conditions in a small region. At certain times external large scale atmospheric forcing may make such patterns very stable. This hypothesis must be explored further, before any conclusions can be made.

It is clear from the above discussion that drought mechanisms are complex. Broad scale causes of drought are generally global, since the major climatic parameter is the general circulation of the atmosphere. On a local scale, however, drought may be a result of the combined effect of many physical factors acting on and interacting with each other, in which case the relative contribution of local terrestrial factors and ocean forcing becomes critical. Hulme (1989) summarized it by

saying that the state of land cover becomes an internal regulator of Sahelian rainfall and creates a resiliency within a climatic regime. Once the environment is sufficiently destabilized through deforestation, or other degradation processes, it responds to external forcing, in this case ocean warming. A proper and systematic documentation of the complex dynamics of West African rainfall is a prerequisite for the understanding of drought's causal mechanisms.

CHAPTER 3

SOME ASPECTS OF APRIL-OCTOBER PRECIPITATION CLIMATOLOGY

3.1 Introduction

Rainfall in Nigeria exhibits more pronounced seasonal and interannual variability than any other meteorological element. In addition, precipitation in Nigeria displays great spatial contrasts. Because the persistent droughts, that have been affecting some areas of the country, can be attributed to reduced rainfall activity, the study of precipitation series is very important.

Different aspects of the rainfall anomalies in West Africa were investigated by Adedokun (1978), Nicholson (1980c, 1983, 1985), Lamb (1982), Gregory (1982, 1983) and Dennet et al. (1985). Some aspects of Nigerian rainfall were investigated by Oguntoyinbo (1967), Ilesanmi (1972), Obasi (1972), Kowal and Adeoye (1973), Adefolalu (1973, 1982b, 1983a, 1986a,b), Ojo (1977), Obasi et al. (1977, 1980), Adefolalu and Oguntoyinbo (1985), Olaniran (1987, 1988a,b), Olaniran and Sumner (1989, 1990) and Adejuwon et al. (1990) among others.

Adefolalu (1986a) investigated long term precipitation trends in Nigeria in order to illustrate the changing patterns in annual precipitation, number of rainy days and length of the rainy season. Recently Olaniran and Sumner

(1989) examined the variability of precipitation in Nigeria, in terms of the onset, retreat and length of the rainy season, using data for the period 1919-1985 and mean rainfall intensity as an index of rainfall activity. Adejuwon et al. (1990), using the data for sixteen stations in Nigeria for the period 1922-1985, examined patterns of rainfall fluctuations in terms of their trends and periodicities.

This study goes further and investigates some important statistical aspects of precipitation in Nigeria such as normality, homogeneity, randomness and secular changes.

3.1.1 Data Base, Initial Data Survey and Station Distribution

Daily rainfall data for about seventy meteorological stations were obtained from the Nigerian meteorological services, Oshodi, Lagos, with the assistance of Dr. E.O. Oladipo. From these daily series, monthly, seasonal and annual rainfall values were obtained.

The initial data survey revealed four major problems affecting the data. First, some stations records contained a large number of missing monthly values. Such stations would produce poor matrix, making the analysis biased and, in extreme cases, impossible. The majority of such cases were found in the south-eastern part of the country for the period 1967-1971. The other problems were: clustering of stations, particularly in the south-western subregion, frequent changes in some station's locations and short record

length compounded by missing values. The task was, therefore, to select stations with good long series of data that were well distributed throughout the country. Only stations with a minimum of thirty years of continuous data were first selected. Then, only the best series were selected to eliminate the problem of clustering in the south-western subregion. Stations with evidence of frequent changes in location were rejected. In four cases, however, data from two series of the same station were combined to form a single long term record. Also, the procedure described in section 3.1.2 was applied wherever possible to stations in the south-western part of the country to fill the gaps in the series.

On the basis of these criteria, fourty one stations were selected out of the original seventy for the present study. The length of the period of observation ranges from 30 to 90 years, with an average station record length of 64 years. Figure 3.1 shows the spatial distribution of the selected stations and their classification into four zones reflecting areas of differing physical control on climate, using the criteria of Olaniran (1987). It is important to note that this classification does not include vegetation in the region of study.

The present study uses the time series of precipitation totals for the months of April to October for the drought assessment. These are the months during which stations in the region receive over 80 per cent of their annual rainfall totals.



Fig. 3.1. Data collecting stations and geographical regions.

3.1.2 Missing Data

Of the total 18,178 seasonal monthly precipitation values, about 714 were missing for 36 stations, representing 3.9 per cent of the data. There are climatological techniques available for the interpolation of missing data. These are given by Conrad and Pollak (1950) and Bradley (1976). The ratio-of-means method by Conrad and Pollak (1950) involves using two records, one from station A with missing data, and the other from station B with a complete record. The ratio of the means for corresponding periods of record (A/B) is first calculated. Then the missing monthly precipitation value at station A is estimated to be the product of the corresponding monthly value at station B and the ratio of the means. Bradley's (1976) procedure is similar to the ratio-of-means method and it involves two precipitation series for station A and B. From the series:

$$\begin{aligned} &A_1, A_2, A_3, A_4 \ldots\ldots\ldots\ldots A_n \\ &B_1, B_2, B_3, B_4 \ldots\ldots\ldots\ldots B_n \end{aligned} \qquad (3.1)$$

a new series of ratios can be computed such that

$$Z_1 = A_1/B_2,\ Z_2 = A_2/B_2 \ldots\ldots Z_n = A_n/B_n \qquad (3.2)$$

Since the Z-series will be positively skewed, the median

ratio is used to compute an estimate of missing precipitation values (X) in the same way as the ratio-of-means method:

$$X_i = B_i \times Z \text{ (median)} \qquad (3.3)$$

The median approach of Bradley (1976) was used wherever possible for filling the gaps in the record. In addition, some isolated missing monthly values were substituted by the long-term mean value for the month. Some stations in the south-eastern part of Nigeria, however, were left with missing values, since finding the nearest station with complete record was not possible in every case.

3.2 Analysis of Precipitation Series

In the remainder of this chapter, some of the features of precipitation fluctuations in Nigeria are investigated by various statistical methods. The homogeneity of the April-October precipitation series is tested, followed by the test for randomness and the frequency distribution of the series is analysed. Lastly, to present some aspects of secular changes of the seasonal precipitation in Nigeria, both the 30-year and 10-year sub-periods are analysed.

3.2.1 Homogeneity of Seasonal Precipitation Series

Homogeneity of the seasonal precipitation series was tested using Swed and Eisenhart's runs test. In this test runs above and below the median values of the precipitation series were computed. As Thom (1966) pointed out, too many runs would indicate oscillation, while too few runs would be an indication of a trend, or a shift in the median during the sample record. For the precipitation series to be homogeneous at a specified significance level (e.g. 5 per cent) the number of calculated runs must lie within the range of U obtained from statistical tables of Rohlf and Sokal (1969). Table 3.1 shows the results of the runs test. The results confirm homogeneity for all stations at a 5 per cent significance level.

3.2.2 Randomness of Series

Randomness of the precipitation series could be tested by a variety of tests. The runs test as applied in section 3.2.1 is a common and simple test to ascertain the randomness of the sample under investigation. However, this test is not as efficient as the Mann-Kendall and Spearman's rank statistic (Mitchell et al., 1966). This study, therefore, uses the Spearman's rank statistic to test for trends against randomness of the rainfall series. The Spearman's rank statistic (r_s) is given as:

TABLE 3.1

RUN'S TEST FOR HOMOGENEITY OF SEASONAL PRECIPITATION

Zone	Station	Median	U (runs)	N of cases	Upper & lower limits of runs (P) .025	.975
Sahelian zone	Sokoto	684.6	27	71	27	45
	Birnin-Kebbi	764.4	23	56	21	37
	Gusau	877.3	17	45	16	30
	Katsina	690.8	26	56	21	37
	Kano	811.3	37	83	32	52
	Hadejia	623.5	23	55	20	36
	Potiskum	706.9	20	47	16	32
	Maiduguri	633.4	31	72	28	46
Midland area	Yelwa	976.9	21	41	14	28
	Kaduna	1225.0	30	65	24	42
	Zaria	1047.4	28	59	22	30
	Jos	1352.1	27	66	25	43
	Bauchi	1061.9	44	73	28	46
	Gombe	876.6	14	30	9	21
	Numan	805.4	16	26	8	20
	Yola	935.3	36	71	27	45
	Pankshin	1037.0	24	49	17	33
	Lafia B-B	1204.4	22	52	19	35
	Minna	1246.6	42	72	28	46
	Bida	1189.7	32	60	22	40
	Lafiagi	1103.3	20	31	10	22

TABLE 3.1 (cont.)

Zone	Station	Median	U (runs)	N of cases	Upper limits .025	lower of runs (P) .975
Guinea-Savanna zone	Ilorin	1135.6	32	71	27	45
	Ibadan	1052.6	41	81	31	51
	Oyo	996.0	34	70	27	45
	Offa	1061.2	31	61	22	40
	Lokoja	1071.7	31	63	23	41
	Auchi	1168.1	19	35	11	25
	Makurdi	1222.9	25	44	16	30
	Yandev	1232.9	24	44	16	30
	Ado-Ekiti	1066.3	18	41	14	28
Coastal area	Lagos	1541.7	37	73	28	46
	Ondo	1299.1	34	78	30	50
	Benin	1819.4	37	82	32	52
	Agbor	1742.2	24	59	22	38
	Kwale	1972.2	29	50	18	34
	Owerri	2136.2	25	67	25	43
	P-Harcourt	2045.8	36	68	26	44
	Uyo	1963.8	19	38	13	27
	Afikpo	1831.3	26	65	24	42
	Calabar	2471.3	39	86	34	54
	Enugu	1588.1	33	62	23	41

$$r_s = \left[1 - \frac{6 \sum_{i=1}^{N} d_i^2}{N\,(N^2 - 1)}\right] \qquad (3.4)$$

where d_i is the difference between the rank of the i^{th} term and the number i itself of that term (where i ranges from 1 to N). For large samples, when N is ten, or larger, the significance of r_s under the H_o may be tested by:

$$t = r_s \sqrt{\frac{N-2}{1-r_s^2}} \qquad (3.5)$$

with the probability points of Student's t with df=N-2 (see Siegel, 1956 and Mitchell et al., 1966 for details). The results of Spearman's randomness test is presented in Table 3.2. Fourteen stations, 11 of them in the Sahelian and Midland zones and 3 in the Coastal area show some tendency toward negative trends at a 5 per cent significance level. Only in one case is the trend positive. Negative trends for most of the stations is a good indication of a possible shift in the rainfall regime toward increased aridity.

3.2.3 Nature of Frequency Distribution

To determine the nature of frequency distribution seasonal precipitation series were tested for normality. Fisher's statistical procedure based on the first four sample moments (e.g. mean, variance, skewness, kurtosis) was applied. The moments were then used to obtain

TABLE 3.2

SPEARMAN'S (r_s) RANK STATISTICS FOR RANDOMNESS

Zone	Station	r_s	N	P
Sahelian zone	Sokoto	-.2963	72	.006*
	Birnin--Kebbi	-.3336	56	.006*
	Gusau	-.3018	45	.022*
	Katsina	-.3438	56	.005*
	Kano	-.3192	83	.002*
	Hadejia	.0941	56	.245
	Maiduguri	-.2706	72	.011*
	Potiskum	-.3850	46	.004*
Midland area	Yelwa	-.2979	46	.022*
	Zaria	-.3401	59	.004*
	Kaduna	-.1158	64	.181
	Jos	-.2410	65	.027
	Bauchi	-.2237	73	.029
	Gombe	-.2261	30	.119
	Numan	.3313	30	.049
	Yola	-.0872	72	.233
	Pankshin	-.3221	46	.015*
	Lafia B-B	-.1621	51	.128
	Minna	-.3545	72	.001*
	Bida	-.1444	59	.138
	Lafiagi	.0770	31	.340

*non-random significant cases at 5 per cent level

TABLE 3.2 (cont.)

Zone	Station	r_s	N	P
Guinea-Savanna zone	Ilorin	-.0608	71	.307
	Ibadan	.0259	81	.409
	Oyo	.1117	71	.177
	Offa	.1114	62	.194
	Lokoja	-.1510	63	.119
	Auchi	-.0296	30	.440
	Makurdi	-.0449	43	.388
	Yandev	-.0751	44	.314
	Ado-Ekiti	.1472	41	.179
Coastal area	Lagos	-.0165	73	.445
	Ondo	.0999	78	.192
	Benin	.0323	78	.412
	Agbor	.3693	59	.002*
	Kwale	-.1048	42	.255
	Owerri	-.0400	67	.374
	P-Harcourt	.0171	68	.445
	Uyo	.0769	34	.333
	Afikpo	-.2014	65	.054
	Calabar	-.2754	85	.005*
	Enugu	-.1156	65	.185

coefficients that indicate the degree of skewness and kurtosis in the data. The first moment (m_1) about the zero as origin is obtained from:

$$m_1 = \bar{x} = \frac{1}{n}\Sigma x_i \qquad (3.6)$$

The second (m_2), third (m_3) and fourth (m_4) moments around the mean are given respectively by:

$$m_2 = \frac{1}{n}\Sigma x_i^2 - \bar{x}^2 \qquad (3.7)$$

$$m_3 = \frac{1}{n}\Sigma x_i^3 - \frac{3}{n}\bar{x}\Sigma x_i^2 + 2x^3 \qquad (3.8)$$

$$m_4 = \frac{1}{n}\Sigma x_i^4 - \frac{4}{n}\bar{x}\Sigma x_i^3 + \frac{6}{n}\bar{x}^2\Sigma x_i^2 - 3\bar{x}^4 \qquad (3.9)$$

The computing formulae of skewness (γ_1) and kurtosis (γ_2) used by the Statistical Analysis System (SAS) in this study are given as:

$$\gamma_1 = \frac{\left\{\left[\sum_{i=1}^{N} x_i^3 - 3\bar{x}\left(\sum_{i=1}^{N} x_i^2\right) + 3x^2\left(\sum_{i=1}^{N} x_i\right)\right] / N - x^3\right\}}{\left\{\left[\left(\sum_{i=1}^{N} x_i^2\right) - N\bar{x}^2\right] / (N-1)\right\}^{3/2}} \qquad (3.10)$$

$$\gamma_2 = \frac{\left\{\left[\sum_{i=1}^{N} x_i^4 - 4\bar{x}\left(\sum_{i=1}^{N} x_i^3\right) + 6\bar{x}^2\left(\sum_{i=1}^{N} x_i^2\right) - 4x^3\left(\sum_{i=1}^{N} x_i\right)\right] / N + \bar{x}^4\right\}}{\left\{\left[\left(\sum_{i=1}^{N} x_i^2\right) - Nx^2\right] / (N-1)\right\}^2} - 3 \qquad (3.11)$$

For normally distributed series, the values of (γ_1) and (γ_2) are equal to zero. The statistics

$$Z_1 = \gamma_1 / SE_1 = \gamma_1 / (6/N)^{\frac{1}{2}} \quad (3.12)$$

and

$$Z_2 = \gamma_2 / SE_2 = \gamma_2 / (24/N)^{\frac{1}{2}} \quad (3.13)$$

were used to test the H_o that the sample series came from a normal population. The test for normality at 95 per cent confidence level requires that the values of Z_1 and Z_2 should be less than 1.96. Table 3.3 shows the values of different parameters for each series. As can be seen from the Table, ten stations (24 per cent) have seasonal precipitation series that possess a degree of skewness, indicating non-normality. All the significant cases are positively skewed. On the other hand, the kurtosis values show that the departure from normality was noted in twelve (29 per cent) cases. Most of the significant cases of non-normality are from the Coastal and Guinea-Savanna zones. This reflects marked differences in the precipitation characteristics of the country. The seasonal (April-October) rainfall series used in the study may not be totally representative of the annual rainfall in the southern parts of Nigeria (Coastal and Guinea-Savanna zones), where there is a significant amount of rainfall in the months of November-March. Moreover, the sub-region is characterised by a bymodal

Table 3.3.

Statistics of Seasonal (April - October) Precipitation of 41 stations in Nigeria

Sahelian Zone

Station	Seasonal $\bar{X}$(mm)	S	CV(%)	γ_1	γ_2	Z1	Z2
Sokoto	685.5	149.6	21.8	-0.099	-0.213	0.348	-0.375
Birnin-Kebbi	775.9	147.4	19.0	-0.261	-0.738	0.839	-1.186
Gusau	913.1	183.2	20.1	-0.240	-2.060	-0.631	2.852*
Katsina	687.6	145.2	21.1	-0.115	-0.757	-0.368	-1.207
Kano	818.3	176.9	21.6	-0.028	-0.369	-0.103	-0.686
Hadejia	623.2	123.6	19.8	-0.523	-0.556	1.596	0.850
Potiskum	743.7	183.8	24.7	0.070	-0.714	0.193	-0.988
Maiduguri	625.6	152.9	24.4	-0.159	-0.571	-0.551	-0.989

Midland Area

Yelwa	964.4	166.7	17.3	-0.179	0.143	-0.467	0.186
Kaduna	1264.2	197.6	15.6	0.376	0.579	1.226	-0.953
Zaria	1056.4	175.2	16.6	0.185	0.068	0.580	1.568
Bauchi	1063.9	168.5	15.8	0.196	0.248	0.683	0.433
Gombe	923.3	157.9	17.1	1.164	3.345	2.558*	3.677*
Numan	815.8	196.2	24.0	0.093	-0.742	0.194	-0.773
Yola	936.9	172.9	18.4	0.211	-0.302	0.726	-0.520
Jos	1341.8	173.0	12.9	0.008	-0.389	-0.025	-0.640
Pankshin	1047.2	239.0	22.8	1.926	9.471	5.333*	13.112*
Minna	1261.6	214.4	17.0	0.266	-0.123	0.921	-0.213
Bida	1171.6	180.0	15.4	0.096	0.150	0.301	0.236
Lafiagi	1126.6	253.4	22.5	0.680	0.777	1.546	0.883
Lafia Beri-Beri	1222.2	206.1	16.9	0.394	1.423	1.149	2.074*

Table 3.3. (Cont.)

Guinea - Savanna Zone

Station	Seasonal $\bar{X}$(mm)	S	CV(%)	γ_1	γ_2	Z_1	Z_2
Ilorin	1153.8	215.3	18.7	0.133	-0.663	0.458	1.141
Oyo	1020.7	255.1	25.0	0.196	-0.003	0.675	0.005
Offa	1130.9	278.8	24.7	1.538	3.195	4.943*	5.136*
Lokoja	1081.9	254.7	23.5	0.380	1.268	1.231	2.054*
Makurdi	1224.0	202.0	16.5	0.177	0.075	0.475	0.101
Yandev	1240.3	189.3	15.3	0.070	-0.810	0.189	1.097
Ibadan	1082.2	258.0	23.8	0.634	0.435	2.331*	0.799
Ado-Ekiti	1081.4	172.7	16.0	0.534	0.463	1.390	0.605
Auchi	1150.1	304.0	26.4	0.119	0.217	0.262	0.238

Coastal Area

Lagos	1567.5	380.8	24.3	1.201	3.225	4.190*	5.624*
Ondo	1356.7	308.4	22.7	1.013	1.153	3.650*	2.079*
Benin	1781.8	323.0	18.1	0.275	0.892	1.015	1.650
Agbor	1742.9	299.8	17.2	1.000	2.579	3.135*	4.044*
Kwale	1962.8	294.2	15.0	-0,111	-0.901	-0.294	-1.191
Enugu	1591.0	253.0	15.9	1.228	4.417	3.947*	7.099*
Afikpo	1749.5	321.8	14.9	0.325	0.151	1.087	0.253
Owerri	2159.5	321.8	14.8	0.325	0.151	1.087	0.253
Uyo	2086.8	348.5	16.7	0.746	0.363	1.776	0.432
Calabar	2482.7	488.0	19.7	0.608	1.524	2.303*	2.885*
P-Harcourt	2103.7	335.8	16.0	0.758	1.741	2.552*	2.930*

* significant at 5 per cent level

rainfall distribution pattern. The rainfall regime of the Sahelian zone, on the other hand, is characterised by high concentration in the months of April-October, when over 90 per cent of the annual precipitation is expected. This, therefore, can explain cases of non-normality in the southern stations.

There are a number of techniques that could be used to normalize the series, e.g. square and cube root transformations (Stidd, 1970), log transformation and Lambda transformation of Box and Cox (1964). Since at least 75 per cent of the stations are normally distributed and the degree of non-normality is slight for most of the non-normal cases, the precipitation series could be taken to be generally normal. Therefore, no transformation was applied.

In order to produce a comprehensive description of precipitation climatology in Nigeria, the mean, variance, standard deviation and coefficients of variation for the series were calculated. Figure 3.2 shows the spatial distribution of the seasonal rainfall. Mean seasonal precipitation values vary considerably over the region, from about 615mm and 690mm in the Sahelian zone, to over 2400mm in the extreme south-eastern part of the Coastal area. Figure 3.2 shows an east-west asymmetry of the southern rainfall maximum, a secondary maximum around the Jos Plateau and relative rainfall minimum running almost east-west between latitudes 7°N and 8°N. North of the secondary maxima of Jos Plateau, zonal

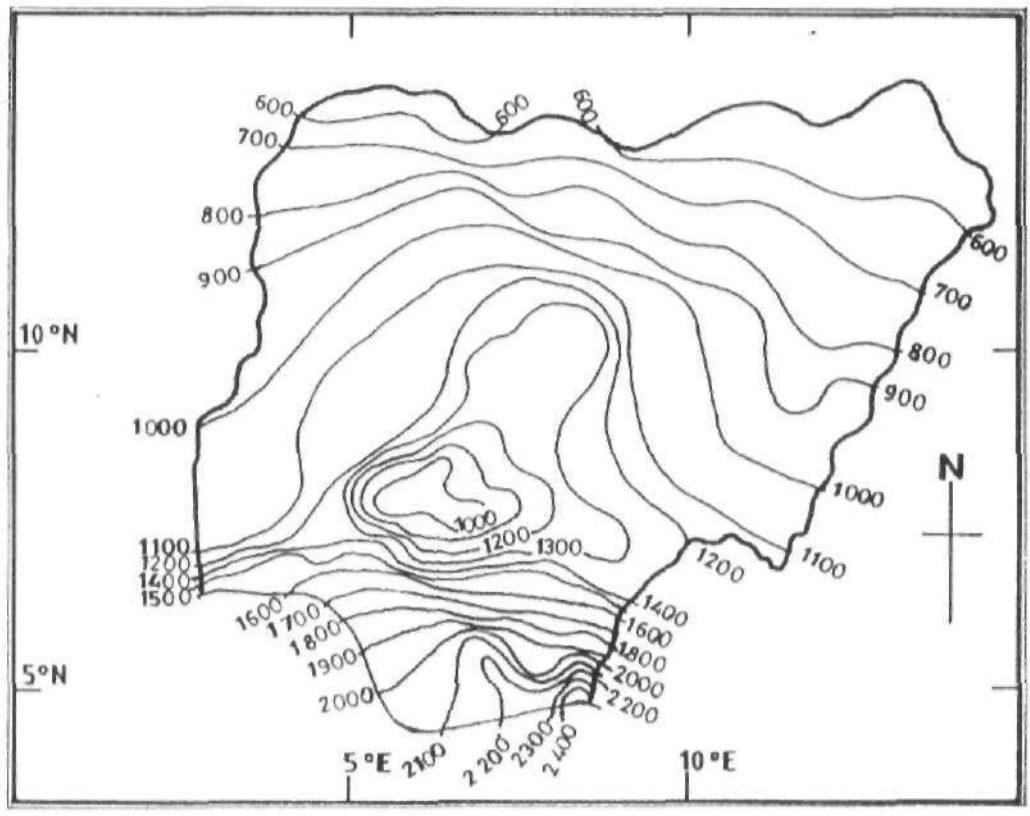


Fig. 3.2. Mean seasonal (April - October) precipitation (mm).

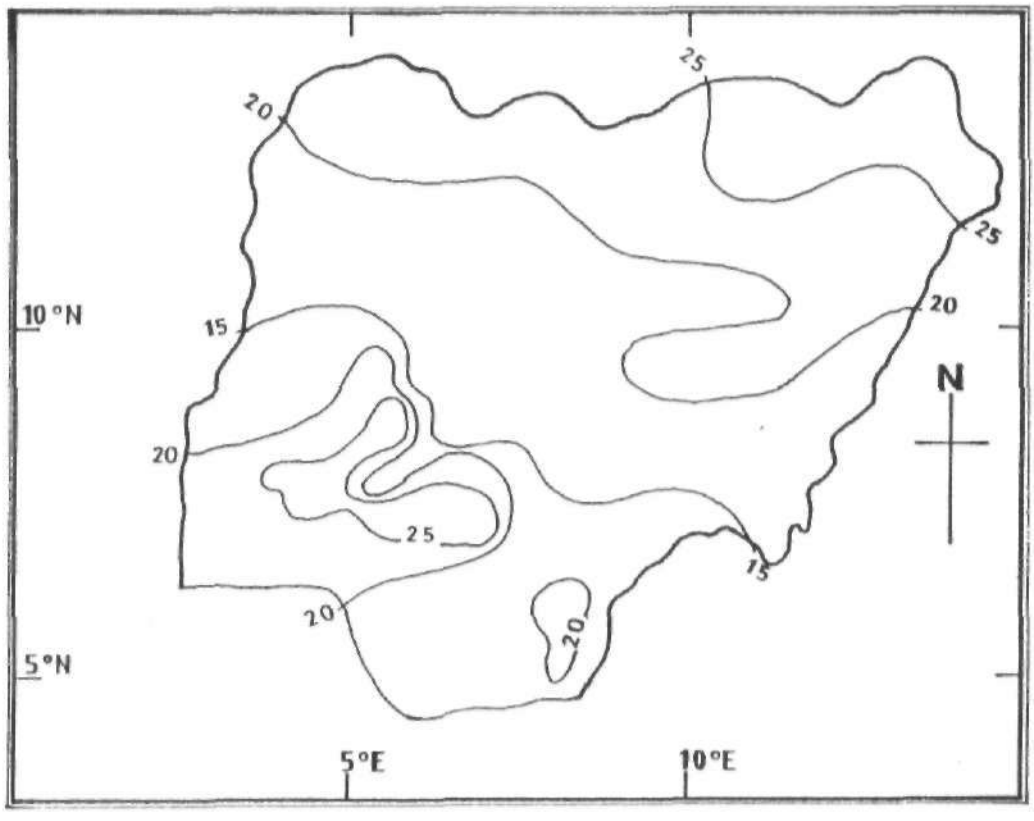


Fig. 3.3. Coefficient of variation of seasonal precipitation (%).

symmetry of the isohyets is conspicuous. Western sub-sector of the Guinea-Savanna and the Coastal zones of the country show rather low seasonal rainfall values (around 1100mm to 1500mm). This rather abnormal feature of rainfall distribution in Nigeria was extensively discussed by Ireland (1962), Adefolalu (1983a, 1986a), Omotosho (1988) and Olaniran (1987, 1988a, 1991). The existence of this dry region is a result of the combination of different factors. Stagnation of zone E weather with drizzle, or light rain, from mostly stratocumulus clouds, which occurs during July and August, general mid-tropospheric subsidence, frequent inversions due to upwelling of cold waters at the Gulf of Guinea, stabilization of the south-westerlies, caused by southern Africa subtropical high pressure - all can reinforce dryness over south-western Nigeria. An interesting feature of the pattern of seasonal rainfall is the position of the secondary minima, which lies between $7^{o}N$ and $8^{o}N$, and longitudes $5^{o}E$ and $7^{o}E$. This secondary minima is south of its latitudinal position noted by Adefolalu (1986a) and it could be an indication of a southward shift in areas which depend largely on April - October precipitation.

The seasonal CV for Nigeria range from about 15 per cent to over 25 per cent (Figure 3.3). The spatial pattern of CV shows two major areas with relatively high values of between 20 per cent and 25 per cent. The first is in the Sahelian zone, extending into the eastern Midland area. The second area of high variability

(CV> 20 per cent) is in the western sub-sector of the Guinea-Savanna and Coastal zone. These two areas are separated by a broad belt of low variability (CV< 20 per cent) extending from the west, through the Midland area, into the eastern Guinea-Savanna zone. This pattern is not surprising and can be explained in the light of atmospheric controls of precipitation over the country. As pointed out by many scholars (Kidson, 1977; Lamb, 1980; Nicholson, 1981b; Achempong, 1982, 1987) the amount of rainfall north of 10°N is the product of many factors such as northward excursion of the ITD, position and strength of the two jet streams, that control the number of squall lines and thunder storms, and the large-scale surface convergence that promotes widespread instability within monsoon air. These features possess a high degree of interseasonal variability, thus, producing high variability of rainfall north of 10°N. The rainfall regime over the south-western part of Nigeria (described above) with a high occurrence of inversions at the height of the monsoon season is responsible for high standard deviation values. Another source of variability might be the highly variable rains of November to March, which are not included in this study.

3.2.4 Blocks of Thirty Years

In order to investigate some aspects of secular changes in the rainfall series, they were sub- divided into 30 year periods corresponding to climatic normals as recommended by the World Meteorological Organization. The 82 (1906-1987) year period examined includes six over-lapping 30 year blocks (1911-1940, 1921-1950, 1931-1960, 1941-1970 1951-1980 and 1961-1987). Also, in order to capture recent changes in the rainfall regimes, the two recent 30 year non-overlapping periods (1931-1960 and 1961-1987) were examined. The period 1931-1960 represents the WMO climatic normal, while 1961-1987 constitutes the most recent nearly 30 year period. Period 1921-1987 was used in this chapter as a base because most stations have data for those years. The use of 1911 as a starting year for the overlapping periods is due to an attempt to maximize use of the available data.

All blocks were tested for normality by the procedure outlined in section 3.2.3 and the variances of (γ_1) and (γ_2) for each block were calculated as:

$$V(\gamma_1) = \left[6N(N - 1) \right] / \left[(N-2)\ (N-1)\ (N+3) \right] \qquad (3.14)$$

and

$$V(\gamma_2) = \left[24N(N-1)^2 \right] / \left[(N-3)\ (N-2)\ (N+3)\ (N+5) \right] \qquad (3.15)$$

where N is the total number of years under investigation.

Twelve to 27 per cent of the stations showed significant departure from normality at the 5 per cent level for the overlapping blocks. About 27 per cent and 19.5 per cent of the stations showed departure from normality as tested by skewness for the non-overlapping blocks of 1931-1960 and 1961-1987 respectively. Most of the cases came from the Coastal and Guinea-Savanna zones.

The sub-periods were also compared to the whole period 1921-1987. For this purpose, Cramer's test (see Mitchell et al., 1966 for details) was applied. This test is preferred when the stability of a long record, in terms of a comparison between the overall mean of an entire record, must be examined. The purpose of the test is to determine whether the differences of the means are no larger than would be compatible with th "null" hypothesis of randomness. The Cramer's test is given as:

$$t_k = \left\{ \frac{n(N-2)}{N - n(1 + \tau_k^2)} \right\}^{\frac{1}{2}} \tau_k \qquad (3.16)$$

where τ_k is a standardized measure of the difference between means given as:

$$\tau_k = (\bar{x}_k - \bar{x}) / s \qquad (3.17)$$

where $\bar{x}$ and s are the mean and the standard deviation of the entire record of N years respectively, $\bar{x}_k$ is the mean of the sub-period of *n* years and t_k is the value

of Student's t with N-2 degrees of freedom. Comparisons between the non-overlapping sub-periods 1931-1960 and 1961-1987 were carried out for each station and their significance tested by the Student's test (t_d) with n_1+n_2-2 degrees of freedom:

$$t_d=(\bar{x}_1-\bar{x}_2) \Big/ \left[(n_1 s_1^2 + n_2 s_2^2)/(n_1 + n_2 - 2)\left(\frac{1}{n_1} + \frac{1}{n_2}\right)\right]^{\frac{1}{2}} \quad (3.18)$$

where $\bar{x}_1$ and $\bar{x}_2$ are the means of the different sub-periods (n_1 and n_2) of the record and s_1 and s_2 their respective standard deviations.

The results of Cramer's test and Student's t test are presented in Tables 3.4 and 3.5. The results of Cramer's test indicate that on the whole, for all the overlapping periods, about 60 per cent of the stations show statistically significant t_k values at the 5 per cent level. The number of significant cases for each of the overlapping blocks (except 1961-1987 period) is between 3 and 8 cases. This dramatically increased to 16 cases for 1961-1987 period. There is a coherent pattern in the spatial distribution of significant cases. In general, most of the significant cases for all subperiods come from the Sahelian and Midland zones and western Guinea-Savanna zone. All significant cases in the Sahelian and Midland zones for all the overlapping periods (except 1961-1987 period) are positive, indicating wetter than long-term conditions. For the non-overlapping blocks of 1931-1960

TABLE 3.4

RESULTS OF CRAMER'S TEST FOR BLOCKS OF 30 YEARS

Sahelian zone

Station	1911-40 t_k	1921-50 t_k	1931-60 t_k	1941-70 t_k	1951-80 t_k	1961-87 t_k
Sokoto	1.315	1.855	2.712*	2.060*	0.094	-4.167*
Gusau	-	-0.495	1.768	2.270*	2.054*	-2.228*
Birnin-K	2.776*	1.726	1.155	-0.103	-1.726	-3.285*
Katsina	1.625	1.404	3.181*	1.966*	0.183	-5.286*
Kano	3.045*	1.981*	2.928*	1.522	0.356	-4.575*
Hadejia	-0.735	-0.179	2.615*	0.825	0.181	-1.398
Potiskum	1.839	0.471	2.503*	1.891	0.460	-2.360*
Maiduguri	1.732	0.532	1.781	1.852	2.131*	-3.204*

Midland Area

Yelwa	-	-1.675	0.571	2.585*	3.461*	-0.677
Zaria	1.236	0.870	1.758	1.530	0.803	-3.142*
Kaduna	1.409	-0.675	-1.469	-0.205	0.415	-0.609
Jos	0.626	1.308	1.703	1.100	-0.112	-2.370*
Bauchi	1.247	1.812	2.094*	0.637	-0.566	-2.317*
Gombe	-	-	0.996	1.442	1.050	-4.349*
Numan	-	-0.157	-1.888	-1.010	-0.698	2.082*
Yola	0.316	0.816	-0.485	0.116	0.118	-1.217
Pankshin	1.478	1.135	0.581	-0.861	-0.262	-1.440
Lafia B-B	0.233	0.429	0.599	1.199	0.715	-1.249
Minna	2.814*	1.867	-0.609	-0.208	-0.306	-2.316*
Bida	2.937*	0.648	-0.179	-1.358	0.327	-0.558
Lafiagi	-	-1.065	0.235	0.190	0.481	0.077

* significant at 5 per cent level

TABLE 3.4 (cont.)

Guinea-Savanna Zone

Station	1911-40 t_k	1921-50 t_k	1931-60 t_k	1941-70 t_k	1951-80 t_k	1961-87 t_k
Ilorin	0.0291	-0.499	-0.512	0.109	0.601	-0.569
Ibadan	0.201	-0.839	-2.173*	-0.544	0.847	1.366
Oyo	-1.633	-1.061	-1.934	-1.291	0.496	1.988*
Offa	-0.420	-1.355	-2.652*	-1.573	1.355	-1.216
Lokoja	2.804*	0.658	-1.436	-1.533	-0.379	-0.408
Ado-Ekiti	-3.029*	-1.889	-0.623	3.477*	2.234*	-0.637
Makurdi	-	-0.627	-0.304	1.304	1.359	-0.119
Yandev	-	0.534	-0.671	-0.554	-0.135	0.202

Coastal Area

Lagos	-0.619	-0.884	-0.645	2.229*	1.886	-0.070
Ondo	-1.074	-0.748	0.297	0.302	0.692	0.648
Benin	-1.167	-1.770	-0.549	1.417	3.615*	0.124
Auchi	-	-1.182	0.722	0.666	1.212	-0.873
Agbor	-2.088*	-3.533*	1.215	0.267	3.536*	3.002*
Kwale	-0.090	0.092	0.845	0.749	0.674	-1.285
Enugu	1.366	1.419	0.931	0.153	0.570	-0.746
Owerri	-	-0.892	-1.252	0.422	1.597	0.276
Afikpo	0.704	2.917*	1.426	-0.019	-2.080*	-3.947*
Port-Harcourt	-0.181	-0.905	-0.445	0.177	1.377	0.198
Calabar	1.349	0.893	0.416	0.231	-1.301	-1.349
Uyo	-	-0.157	-1.045	-0.019	0.169	0.912

TABLE 3.5

Student's t-test for 1931-1960 and 1961-1987 periods

Station	t_d	Station	t_d
Sahelian Zone			
Sokoto	3.793*	Kano	3.984*
Birnin-Kebbi	2.707*	Hadejia	2.090*
Gusau	2.027*	Potiskum	2.541*
Katsina	4.638*	Maiduguri	3.035*
Midland Area			
Yelwa	0.716	Gombe	1.983*
Kaduna	-0.310	Numan	1.890
Zaria	2.592*	Yola	0.384
Bauchi	3.027*	Lafia B-B	0.791
Jos	2.243*	Minna	1.089
Pankshin	1.098	Bida	0.174
Lafiagi	0.121		
Guinea-Savanna Zone			
Ilorin	0.065	Lokoja	-0.599
Offa	-0.383	Makurdi	-0.133
Oyo	-2.102*	Yandev	-0.440
Ibadan	-1.782	Ado-Ekiti	0.191
Coastal Area			
Lagos	-0.402	Owerri	-0.660
Ondo	-0.020	P-Harcourt	-0.338
Benin	-0.371	Uyo	1.199
Agbor	-2.466*	Calabar	1.149
Kwale	1.167	Afikpo	3.889*
Enugu	0.872		

* significant at 5 per cent level

and 1961-1987 eight (six positive cases in the Sahelian and Midland zones) and sixteen (twelve negative cases in the Sahelian and Midland zones) stations respectively show statistically significant values at the 5 per cent level.

The result of Student's test indicates that fifteen stations show statistically significant values at the 5 per cent level. All the Sahelian stations and four northern Midland stations indicate significant positive values.

In summary, the above results indicate that the 30 year block 1931-1960 was wetter in the Sahelian zone than the long-term period. Period 1961-1987 was significantly drier than the long-term period and the 1931-1960 block (as indicated by the distribution of significant t_k and t_d values) in the Sahelian and northern Midland zones. The number of stations with statistically significant values of t_k and t_d in other zones are too few for any meaningful generalisations about significant change in precipitation conditions there. On the basis of the above results it is possible to generalise that there was a significant change toward drier conditions in the Sahelian and northern Midland zones in the last 30 year period.

3.2.5 Decadal Analysis

The data were also considered for the decadal periods, that is 1911-1920, 1921-1930, 1931-1940, 1941-1951-1960, 1961-1970, 1971-1980 and 1981-1987. Decadal averages of the seasonal precipitation series for each station were examined to see if they differ from the mean of the entire period by using Cramer's test as described in section 3.3.1. The result of the test is presented in Table 3.6. From the Table it is evident that:

(a) In the Sahelian zone there is a strong pattern of significant cases. All stations (except Hadejia) show significant negative departures at 5 per cent level for either both, or one, of the 1971-1980 and 1981-1987 periods. Six out of eight, Sahelian stations indicate significant positive departures for 1951-1960 at 5 per cent level. These results indicate that while the decade 1951-1960 was generally wetter than the long-term, the decades of 1971-1980 and 1981-1987 were generally significantly drier than the whole period in the Sahelian zone.

(b) In the Midland area six stations show significant negative departures at 5 per cent level for the 1981-1987 period. These stations are at the northern fringes of the Midland area.

(c) In the Guinea-Savanna and Coastal zones there are generally fewer cases of departures from the

TABLE 3.6

RESULTS OF CRAMER'S TEST FOR DIFFERENT DECADES

Sahelian Zone

Station	1911-20	1921-30	1931-40	1941-50	1951-60	1961-70	1971-80	1981-87
Sokoto	-0.101	C.944	0.951	0.836	2.071*	-0.098	-2.035*	-3.345*
Birnin-K	-0.242	2.395*	1.221	-1.239	-1.597	-0.467	-3.737*	-
Gusau	-	-	-	-0.496	2.717*	0.474	0.726	-2.162*
Katsina	-	0.012	1.882	-0.025	2.351*	0.047	-2.557*	-3.383*
Kano	0.968	0.842	2.294*	-0.568	1.974*	0.433	-1.883	-4.165*
Hadejia	-0.928	-2.285*	1.855	0.847	1.443	0.097	-1.399	-
Potiskum	-	-	1.840	-0.666	2.469*	0.946	-2.465*	-1.456
Maiduguri	1.192	0.606	0.771	-0.648	2.396*	0.862	-0.294	-4.872*

Midland Area

Yelwa	-	-	-	-1.677	2.305*	2.572*	-0.879	-2.403*
Kaduna	-0.613	2.622*	-0.310	-0.432	-0.233	1.522	-0.846	-1.723
Zaria	-	1.241	0.794	-0.114	1.986*	0.847	-1.182	-2.848*
Jos	-	0.285	0.514	0.990	0.800	0.301	0.666	-2.072*
Bauchi	0.356	0.554	1.258	1.008	0.985	-0.894	-0.943	-2.600*
Numan	-	-	-	-0.153	-1.845	1.087	1.004	1.626
Yola	-1.573	2.347*	-0.549	-0.553	0.475	0.369	-0.578	-1.542
Pankshin	-	-	1.478	-0.321	-0.589	-0.443	1.572	-1.756
Lafia B-B	-	-	0.582	0.551	0.462	1.010	-0.330	-1.971*
Minna	1.289	3.519*	-0.278	-0.119	-0.429	0.357	-0.485	-2.777*
Bida	-	0.939	2.624*	-2.281*	-0.534	0.946	-0.032	-1.509
Lafiagi	-	-	-	-1.064	1.150	-0.436	-0.864	0.941
Gombe	-	-	-	-	1.053	0.666	-1.044	-1.052

TABLE 3.6 (cont.)

Guinea-Savanna Zone

Station	1911-20	1921-30	1931-40	1941-50	1951-60	1961-70	1971-80	1981-87
Ilorin	-0.701	1.538	-0.711	-1.366	1.522	-0.100	-0.430	-0.221
Offa	-1.143	2.169*	-1.772	-2.099*	0.865	-0.864	1.949	-2.057*
Oyo	-2.062*	0.665	-0.875	-1.357	-0.271	-0.199	1.537	1.445
Ibadan	-0.222	1.004	-1.101	-1.133	-0.994	1.229	0.987	-0.053
Ado-Ekiti	-	-	-3.031*	0.190	1.321	1.784	-0.982	-0.864
Lokoja	1.724	2.553*	-0.549	-1.296	-0.248	-0.663	0.315	-0.843
Makurdi	-	-	-	-0.596	0.248	1.877	-0.569	-1.070
Yandev	-	-	1.785	-0.285	-1.395	1.075	0.264	-0.592

Coastal Area

Lagos	-0.518	0.432	-0.699	-0.764	0.672	3.506*	-1.724	-1.594
Ondo	-0.826	-1.594	0.074	-1.484	-0.363	0.716	0.162	0.079
Benin	-0.604	-0.379	-0.635	-1.439	1.277	2.164*	1.242	-2.689*
Auchi	-	-	-	-1.180	1.755	-0.672	-0.986	-0.553
Agbor	-0.967	-1.746	-0.411	-1.972*	1.268	2.146*	2.159*	-
Kwale	-	-	-0.235	0.439	0.832	-0.587	0.349	-1.524
Enugu	1.877	-0.365	0.957	1.784	-0.735	-0.555	0.112	-1.004
Owerri	1.050	0.748	-0.137	-1.764	0.661	2.236*	-0.373	-1.234
Afikpo	-3.133*	2.466*	1.550	-0.299	0.588	-0.451	-3.834*	-1.605
P-Harcourt	-0.567	-0.473	0.648	-1.050	0.210	1.190	0.343	-0.950
Calabar	-0.257	1.694	0.339	-0.363	-0.597	0.094	-2.568*	0.146
Uyo	-	-	-	-0.311	-0.567	1.451	0.120	-0.187

the long-term mean values. Therefore, in general in these zones there is no indication of decades significantly wetter, or drier, than the long-term condition. The number of stations with significant t_k values are too few for any meaningful generalization.

Overall, the 1920s and 1950s were wetter than normal over most of the country, with below average rainfall over only isolated areas. The decade 1941-1950 shows below average rainfall for about 80 per cent of the stations and can be considered to be a dry decade. The decade 1971-1980 indicates below average precipitation for all the Sahelian and 77 per cent of the Midland stations. The period of 1981-1987 was also drier than average over most of the country, with about 88 per cent of the stations recording below average precipitation.

Seasonal average precipitation for each station for each decade was also compared with the 1961-1987 subperiod and the difference expressed as a deviation from the 1961-1987 mean in form of:

$$Z = (\bar{x}d_i - \bar{x}_p) / s_p \qquad (3.19)$$

where $\bar{x}d_i$ is the decadal seasonal average precipitation for decade i, $\bar{x}_p$ is the seasonal average for the period 1961-1987 and s_p is the seasonal standard deviation for the reference period (1961-1987). The period of 1961-1987 was chosen because it is the most recent near 30 year

period. The results are presented in Figure 3.4 (a-h) in form of above, or below, normal (1961-1987) values. From the analysis of Figure 3.4 the following are observable:

(a) The 1961-1987 period was drier than the preceeding 30 year period in the Sahelian, Midland and eastern Coastal zones.

(b) On a decadal basis, both the above and below 1961-1987 mean conditions exhibit no coherent geographical pattern, although the first four decades exhibit below normal rainfall over western Midland, Guinea-Savanna and Coastal zones. For the decade 1941-1950 over 55 per cent of the stations registered below average precipitation.

(c) The decadal period of 1971-1980 exhibits drier than average conditions over the northern part of the Sahelian, western Midland and parts of the Coatal zones, indicating a new pattern of below normal conditions.

(d) The 1981-1987 period exhibits countrywide occurrence of below average precipitation as compared with the reference period of 1961-1987. For this period over 90 per cent of the stations registered below average precipitation. Thus the 1970s and the 1980s exhibit a tendency for below normal conditions north of $10^{o}N$.

Although there is an indication of significantly drier than average conditions over parts of the Sahelian, Midland and Coastal zones since the beginning of the 1970s,

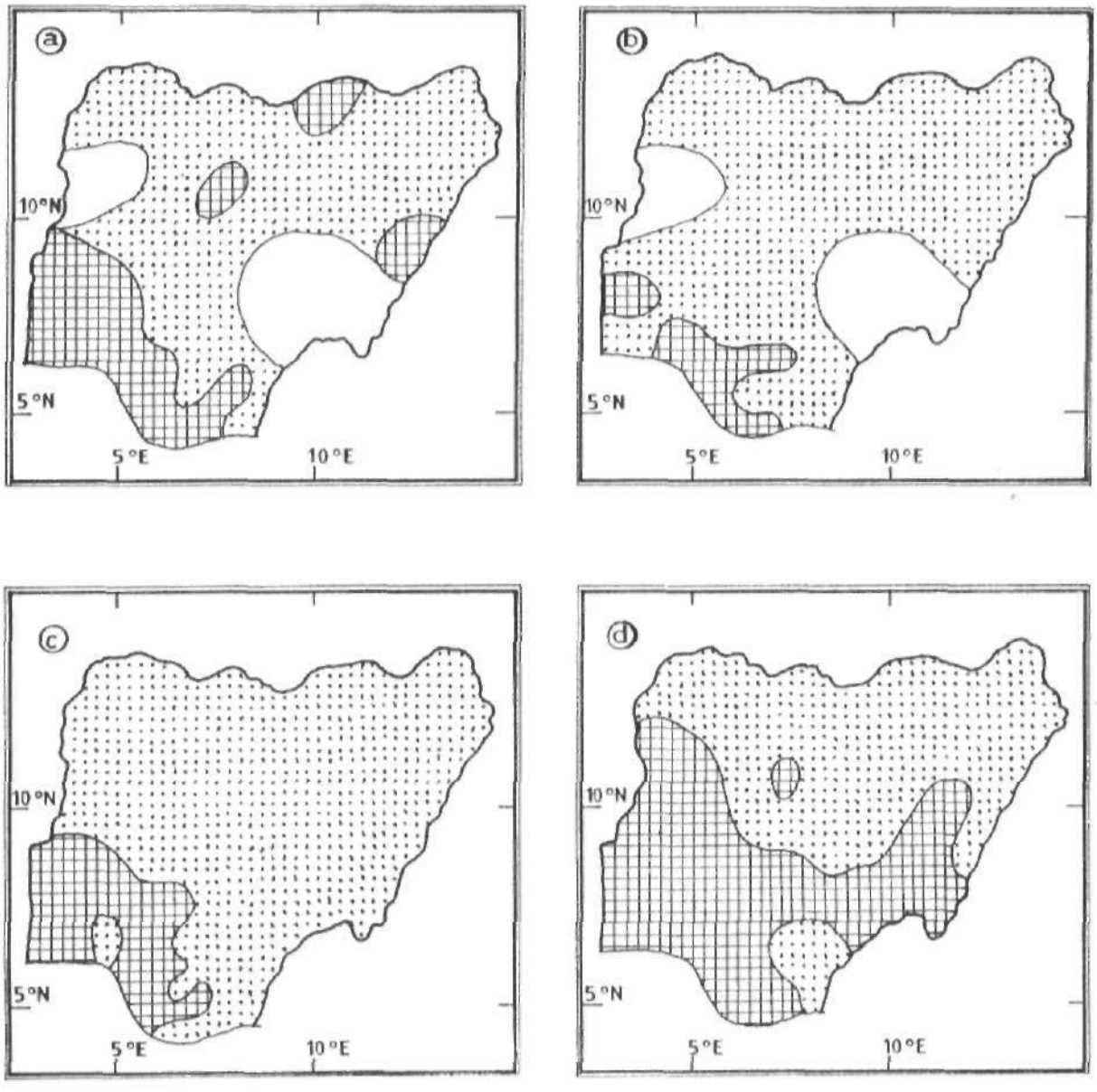


Fig. 3.7: Deviation of mean seasonal precipitation
ⓐ 1911 - 1920 ⓑ 1921-1930 ⓒ 1931 1940 ⓓ 1941-1950
from 1961 - 1987 averages.

LEGEND

Above average

Below average

No data

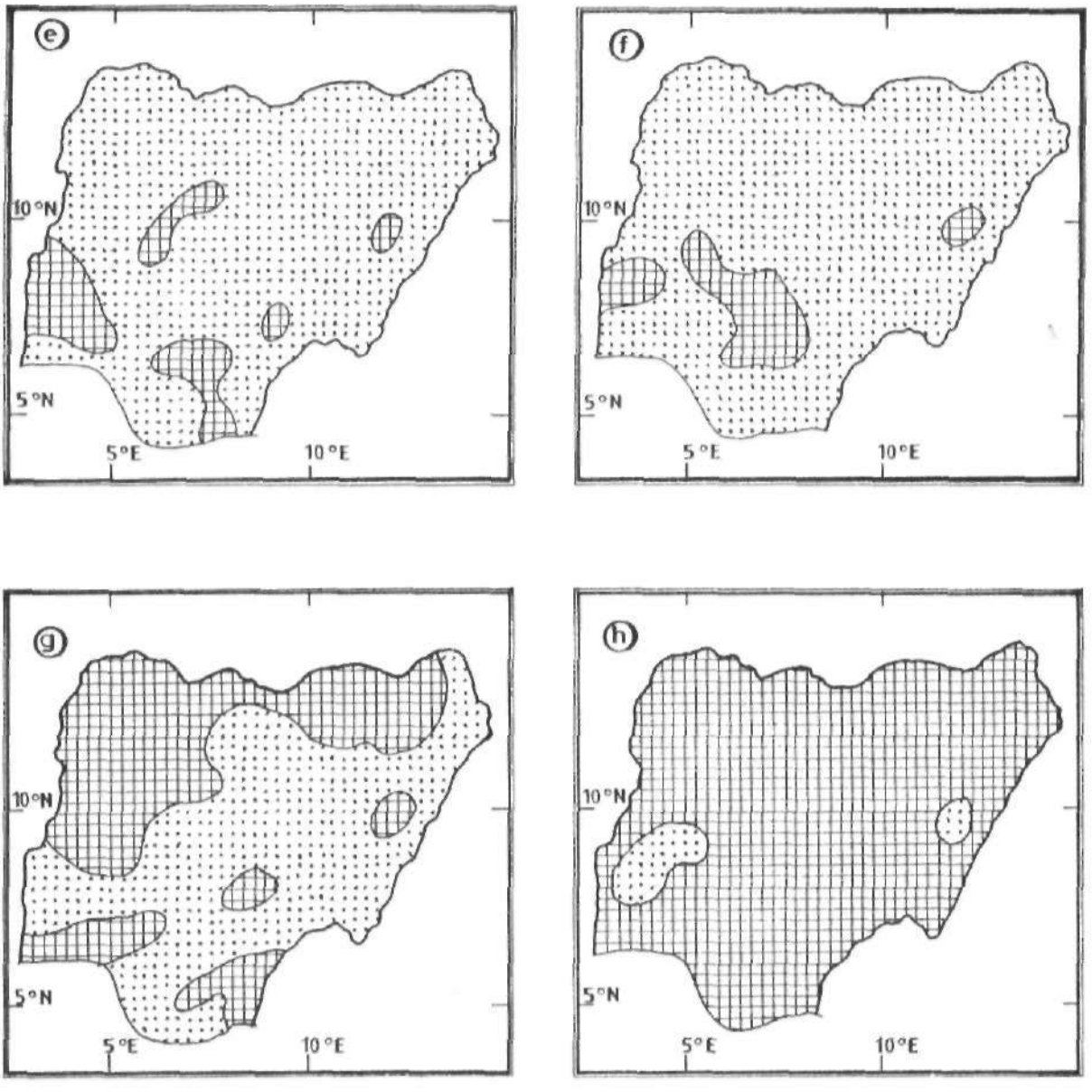


Fig. 3.7.(cont.) for ⓔ 1951-1960 ⓕ 1961-1970 ⓖ 1971-1980 ⓗ 1981 1987 .

CHAPTER 4

DROUGHT INDEXING

4.1 Introduction

Various indices have been developed to give a quantitative description of drought. The widely known meteorological drought indices are those of Foley (1957), Palmer (1965), Herbst et al. (1966) and Bhalme and Mooley (1980). Other indices of interest are the rainfall anomaly index of van Rooy (1965), the rainfall deciles of Gibbs and Maher (1967), percentiles of standardized seasonal precipitation series of Pinkayan (1966) and Tabony's drought index (1977). Extensive review of meteorological drought indices is given in Oladipo (1985). This chapter discusses the merits and demerits of each index in some detail. Then, the index that is derived for Nigeria is presented and its performance is compared with the van Rooy and Tabony drought indices.

4.2 A Review of Some Drought Indices

One of the earliest quantitative expressions of drought was developed by Foley (1957). He integrated the deviations of rainfall from long-term averages to produce a graph of cumulated departures (residual mass curve). The steepness of the fall is a measure of the drought intensity, with deficiencies expressed in thousands of the annual rainfall termed units. Units are cumulated as a method of assessing a drought period

and may also be divided by the number of months in a period to give an index of drought severity. The advantage of the graph is that it clearly shows both the duration and the amount of the rainfall deficiency. The mass curve approach, however, has the disadvantage that all deficiencies carry the same weight, irrespective of the seasonal water requirements of vegetation. Moreover, because the severity index relies only on average rainfall, a comparison of drought incidence and severity between stations having significant differences in average rainfall is not reliable. For example, two stations may have received the same amount of rainfall in a particular month, but one, being normally wetter than the other, would indicate a greater deficit. This can be overcome to some extent by dividing all values of monthly rainfall by the average annual rainfall.

Similar to Foley's method, van Rooy (1965) derived a rainfall anomaly index (RAI), given by an empirical equation:

$$RAI = 3\left[(P - \bar{p})/(\bar{m} - \bar{p})\right] \qquad (4.1)$$

for positive anomalies and

$$RAI = -3\left[(P - \bar{p})/(\bar{x} - \bar{p})\right] \qquad (4.2)$$

for negative anomalies, where P is the actual precipitation, $\bar{p}$ is the long term average precipitation, $\bar{m}$ is the

mean of ten highest values of P on record and $\bar{x}$ is the mean of ten lowest values of P on record. It should be noted that the threshold values of +3 and -3 are arbitrarily assigned to respective means of the 10 wet years and 10 dry years. Nine classes, ranging from extremely wet to extremely dry, were given against a scale of numerical values of the relative RAI. Like Foley's severity index, RAI is not weighted for seasonal fluctuations of rainfall. It thus expresses the degree of rainfall anomaly for the relevant period in relation to the long-term mean rainfall for the specific period. This might incorrectly indicate that an index of, say, -3.0 for a month with an average rainfall of 200mm and that of -3.0 for another month with an average rainfall of 80mm have the same degree of relative dryness. This inference neglects the apparent differences in the seasonal fluctuations of water needs.

Gibbs and Maher (1967) used deciles to describe rainfall distribution. The decile range into which a particular rainfall amount falls is used to indicate the departure of annual,or seasonal precipitation from the average. For example, the first decile is that rainfall amount which is not exceeded by the lowest 10 per cent of the totals, the seventh decile is the amount not exceeded by 70 per cent and so on.

In a similar approach to that of Gibbs and Maher, Pinkayan (1966) treated a standardized annual precipi-

tation series (X) as a random variable with a distribution function F(x) defined as:

$$F(x) = P\ (X \leq x) \qquad (4.3)$$

The method of Gibbs and Maher, and Pinkayan do not consider the role of antecedent moisture conditions and therefore fail to take into account seasonal changes in plant's water requirements. Being based on average rainfall, these methods also make comparisons between stations with different amounts of rainfall difficult.

Herbst et al. (1966) developed another technique, incorporating the carry-over effect of preceding moisture conditions, thus making the method more realistic than those discussed already. Also, the use of monthly values is an obvious advantage. The disadvantages of the method are the laborious calculations involved to determine the drought onset and termination and the lack of a numerical indication of the degree of drought severity. The derivation of the weighting factor is rather empirical.

Tabony (1977) proposed two simple methods of drought analysis. A simple measure of meteorological drought ($Dm,_1$) is given as:

$$Dm,_1 = (R - \bar{R}) / \bar{R} \qquad (4.4)$$

where R is the actual rainfall and $\bar{R}$ the average rainfall. Modifying Equation (4.4) to include the effects of potential evaporation (Ep), the drought equation becomes:

$$Dm,_2 = (R - Ep) - (R - Ep) / R \qquad (4.5)$$

where $\bar{E}p$ is the average potential evaporation. The potential evaporation (Ep) takes place when water supply is unrestricted and is dependent only upon meteorological elements like humidity, temperature and radiation. Tabony's methods are simple, and because they represent fractional deviations from long term mean conditions they are attractive. For the purpose of this study Equation 4.4 is used to compute Tabony's drought index for all the stations.

Perhaps the best known of all the drought indices is that developed by Palmer (1965). It describes drought severity on a regional basis. Although the index was developed for assessing drought, it demonstrates extremely wet situations equally well. Therefore, Palmer's index is at times referred to as simply the Palmer Index (PI) and at other times as the Palmer Drought Severity Index (PDSI). Palmer's index combines precipitation and temperature data as predictor variables and addresses two of the most elusive properties of droughts: their intensity, and beginning and ending times. It is based on the balance between the quantity of available soil moisture and the water requirements of crops, or natural vegetation. It uses the water balance computation of Thornthwaite (1948). Using climatological analysis of long records of temperature and precipitation, five constants are calculated which define the moisture characteristics of the region. The factors are then combined

in various forms to derive an index.

It is clear that Palmer (1965) treats drought severity empirically as a function of accumulated weighted differences between actual precipitation and the precipitation requirement, where the requirement depends on the carry-over of previous rainfall as well as evapotranspiration, moisture recharge, and runoff that would be climatically appropriate for the particular time and place being investigated. Thus, the average requirement is for normal rainfall, but individual periods may require much more below normal rainfall depending on the character of the preceding weather and the temperature of the period.

Palmer's method has been analysed by many users and its weaknesses clearly pointed out. Alley (1984) offered an extensive discussion of the properties and assumptions of the method. There are several limitations embedded in water balance models, which assume that the capacities of the two soil layers are independent of seasonal, or annual changes in vegetation cover. Other serious deficiencies in water balance computations are related to the estimation of potential evapotranspiration and runoff. Most water balance models assume that evapotranspiration for a period is equal to the potential evapotranspiration whenever P PE. This is not always correct.

Also, the universal designation of 25mm as the moisture capacity of the surface layer from which evapotranspiration takes place at the potential rate seems rather arbitrary.

Palmer's runoff term includes both recharge to ground water and overland runoff and the model assumes that runoff does not occur until the soil moisture capacity of the upper and lower layers is filled. Such simplified treatment of runoff results in end-of-month soil-moisture storage (simulated by the Palmer's model) being more often than not at its capacity (AWC) for many regions. This is an unrealistic approximation. Other serious limitations of PDI are the arbitrary designation of drought severity classes, and the term such as "severe" and "extreme" may be rather loosely defined by the PDI. Other limitations are the excessive emphasis on antecedent precipitation and the laborious calculations involved. Because of the considerable emphasis placed on antecedent precipitation, PDI will in some instances bear little relation to the amount of rain that actually falls within a given month. This was pointed out by Bhalme and Mooley (1979), when calculating PDI for some states in India. They discovered that the great drought of 1918 for the Indian subcontinent was shown as a normal year by PDI. In addition, Palmer (1965) combined records of climatic divisions. Thus, the derived PDI values are valid only for the geographical units used in the analysis and not individual stations (Oladipo, 1985).

Inspite of these limitations, the consideration of total water budget of an area and explicit use of

duration in assessing severity are clearly basic advantages. Also PDI should show more internal consistency through time and space when compared to simpler drought measures. With an adequate modification of the weighing factor, K, the PDI can be adapted to suit local conditions (Oladipo, 1985).

4.3 Drought Index Used In This Study

Bhalme and Mooley (1980) modified Palmer's drought index, because it failed to give a realistic picture of drought conditions in arid, semi-arid and dry subhumid regions in India. They developed a drought indicator referred to as the Bholme and Mooley drought index (BMDI) which produces more realistic results in tropical weather conditions. Palmer's index has been modified in respect of the weighing factor and drought severity equation. The development of BMDI is based on the assumption that plant life and established human activities in a region are adjusted to average rainfall conditions

Because of its realistic approach, the procedure of Bhalme and Mooley was followed closely to derive a drought index for drought conditions in Nigeria. The procedure consists of the following steps:

(a) For 41 stations used in this study, the long-term mean monthly precipitation ($\bar{x}$), the standard deviation (s) and coefficient of varia-

tion (cv) for each of the seven months (April -October) were derived.

(b) The percentage departure of monthly rainfall from the long-term mean was used to compute a moisture index MI given by:

$$MI = 100 \cdot (x - \bar{x})/s \qquad (4.6)$$

This moisture index, within reasonable limits permits comparison of the rainfall anomaly in space and time.

(c) Averages of the highest accumulated values of the negative moisture index during various intervals of months were obtained for all 41 stations. This was done in the following steps:

i. Moisture indices for April-October were accumulated.

ii. Highest negatives were picked for each station, and called "highest accumulated negatives".

iii. Monthly highest accumulated negative indices were averaged.

This aspect of index development provides a relative measure of regional moisture anomalies because it permits numerical designation of extreme droughts in various parts of the study region. This data is presen-

ted in Table 4.1

TABLE 4.1 MONTHLY STATISTICS FOR THE HIGHEST ACCUMULATED NEGATIVE MOISTURE INDEX

	April	May	June	July	August	Sept.	October
x	-138.6	-262.1	-375.7	-453.7	-519.8	-597.2	-646.8
s	55.1	60.8	75.0	86.5	93.0	109.3	142.6
CV	-39.7	-23.2	-20.0	-19.1	-17.9	-18.3	-22.0

This data is also presented in Figure 4.1, where the extreme drought conditions (I = -4.0) is shown with the solid line given by the least-squares equation:

$$\sum_{t=1}^{k} M_t = -93.54 - 83.53k, \qquad (4.7)$$

where $\sum_{t=1}^{k} M_t$ is the accumulated moisture index over a duration of k months. The ordinate distance between solid line (extreme drought) and the top of the graph was divided into four equal lengths and the body of the graph was correspondingly divided by three more lines which are arbitrarily labelled as severe, moderate and mild drought . Numerical values are assigned to these lines namely -4.0 for extreme drought, -3.0 for severe drought, -2.0 for moderate drought and -1.0 for mild drought.

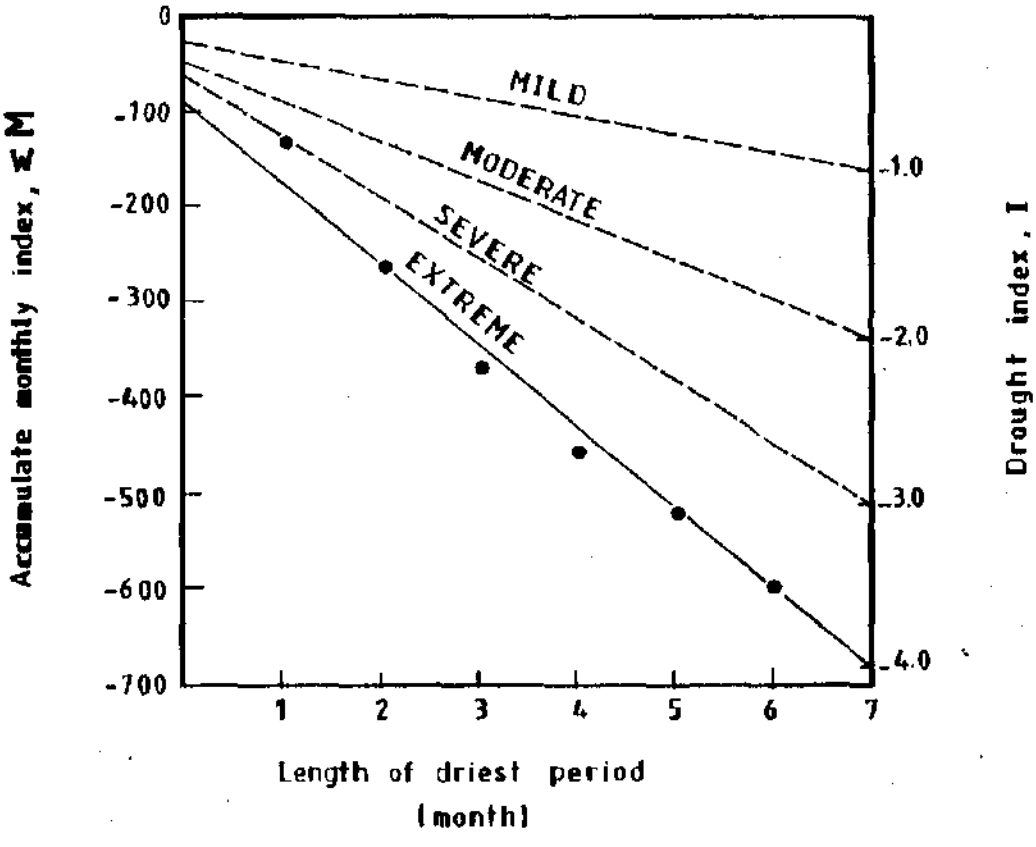


Fig. 4.1. Accumulated moisture index during the rainy season (April - October).

(d) Next let $\sum_{t=1}^{k} \tilde{M}_t$ be an arbitrary value of the accumulated moisture index over a duration of k months. Then the drought intensity I_k for the k^{th} month is given as:

$$I_k = \sum_{t=1}^{k} \tilde{M}_t \; / \; (23.39 + 20.88k) \qquad (4.8)$$

The matter of drought intensity was then considered on an incremental basis such that each successive month is evaluated in terms of its contribution to the severity of drought. Thus the contribution of each month can be obtained by setting k = 1 in equation (4.8) and we get:

$$I_1 = \tilde{M}_1 \; / \; 44.27 \qquad (4.9)$$

For an initial month under investigation $I_o = 0$, hence

$$I_1 - I_o = \Delta I_1 = \tilde{M}_1 \; / \; 44.27 \qquad (4.10)$$

In successive months a negative value of M is required to maintain an existing dry spell. The rate at which the moisture index ($\tilde{M}$) must increase to maintain a constant value of intensity (I) depends on the value of (I) that is to be maintained. This suggests that for all months following the initial dry month there should be an additional term to account for any carry - over effect. Then the equation (4.10) becomes

$$\Delta I_k = (\tilde{M}_k / 44.27) + CI_{k-1} \qquad (4.11)$$

where,

$\Delta I_k = I_k - I_{k-1}$ and C is constant.

To determine C we use the equation (4.8) to compute M_k for two consecutive months for a constant value of I. We may set k = 5, $I_5 = -1$ in equation (4.8), then $\sum_{t=1}^{5} \tilde{M}_t = -127.79$ and likewise setting k = 4, $I_4 = -1$ we have $\sum_{t=1}^{4} \tilde{M}_t = -106.91$. Hence

$$\tilde{M}_5 = \Delta\tilde{M}_s = \sum_{t=1}^{5} \tilde{M}_t - \sum_{t=1}^{4} \tilde{M}_t = -20.88 \qquad (4.12)$$

Substituting this value of $\tilde{M}_5$ into equation (4.11) we get

$$\Delta I = 0 = (-20.88 / 44.27) - 1.0C$$

$$\text{i.e. } C = -0.472$$

Even with other values of intensity, the resulting value of C will come out to be the same (Bhalme and Mooley, 1980). For example, take k = 4 and 3 and I_4 and $I_3 = -2$, then

$$\sum_{t=1}^{4} \tilde{M}_t = -214.80 \text{ and } \sum_{t=1}^{3} \tilde{M}_t = -173.04$$

Therefore

$$M_4 = \sum_{t=1}^{4} \tilde{M}_t - \sum_{t=1}^{3} M_t = -41.76$$

$$\Delta I = 0 = (-41.76 / 44.27) -2.0C$$

$$C = -0.472$$

From the above considerations the final drought index for Nigeria for the months April - October from equation (4.11) is:

$$\Delta I_k = (\tilde{M}_k / 44.27) - 0.472\ I_{k-1} = I_k - I_{k-1}$$

$$I_k = I_{k-1} - 0.472\ I_{k-1} + (\tilde{M}_k / 44.27)$$

$$= 0.528\ I_{k-1} + (\tilde{M}_k / 44.27) \qquad (4.13)$$

Equation (4.13) is the modified drought severity equation, where M is the moisture anomaly index and the subscript k refers to the current month. The modified drought severity equation was used to carry out computations to determine monthly drought severity indices for the month of April - October for each year for 41 stations. From these monthly values the mean index was calculated for each of the years on record for all stations. Equation (4.13) gives either negative or positive values of the index and therefore provides the measure of dryness or wetness. Seasonal mean index values are then classified into nine classes using the same numerical values as Palmer. This classification is shown in Table 4.2.

TABLE 4.2 DESCRIPTION OF DROUGHT (-) / FLOOD (+) INDEX

Index	Character of Weather
$\geq$ 4.00	Extremely wet
3.00 to 3.99	Very wet
2.00 to 2.99	Moderately wet
1.00 to 1.99	Slightly wet
0.99 to -0.99	Near normal
-1.00 to -1.99	Mild drought
-2.00 to -2.99	Moderate drought
-3.00 to -3.99	Severe drought
$\leq$ -4.00	Extreme drought

For the purpose of discussing seasonal characteristics of drought, however, because extreme values are rare on seasonal basis, modified BMDI classification was adopted. This classification is presented in Table 4.3.

TABLE 4.3 MODIFIED BMDI CLASSIFICATION

Index	Character of Weather
$\geq$ 3.00	Very wet
1.00 to 2.99	Wet
-1.00 to 0.99	Near normal
-2.99 to -0.99	Moderate drought
$\leq$ -3.00	Severe drought

4.4 A Comparison of Three Drought Indices: BMDI, RAI and Tabony

In order to assess the performance of the modified BMDI, the mean seasonal (April - October) values of RAI and Tabony indices for all 41 stations were calculated and plotted together with seasonal BMDI.

Stations selected to be presented for this chapter come from four climatic zones in Nigeria as defined by Olaniran (1987) and described in Chapter 3. Single - station, rather than areally averaged indices were plotted since, as pointed out by Oladipo (1985), Bhalme and Mooley's method is especially useful for single station analysis of meteorological droughts.

The results of the three drought indices for 16 stations for long periods are presented in Figures 4.2 - 4.5 and Appendix A. A comparative analysis of the three reveal the following:

(a) Generally the three drought indices agree well on the number of years with drought / flood conditions. Therefore, they appear to be effective and suitable in describing meteorological drought in Nigeria. All previously identified drought periods in different parts of the country are presented by the three indices. For example, droughts of 1932 and 1936-1938 that affected Coastal and south-western Guinea-Savanna zones (described by Olaniran, 1991) are depicted

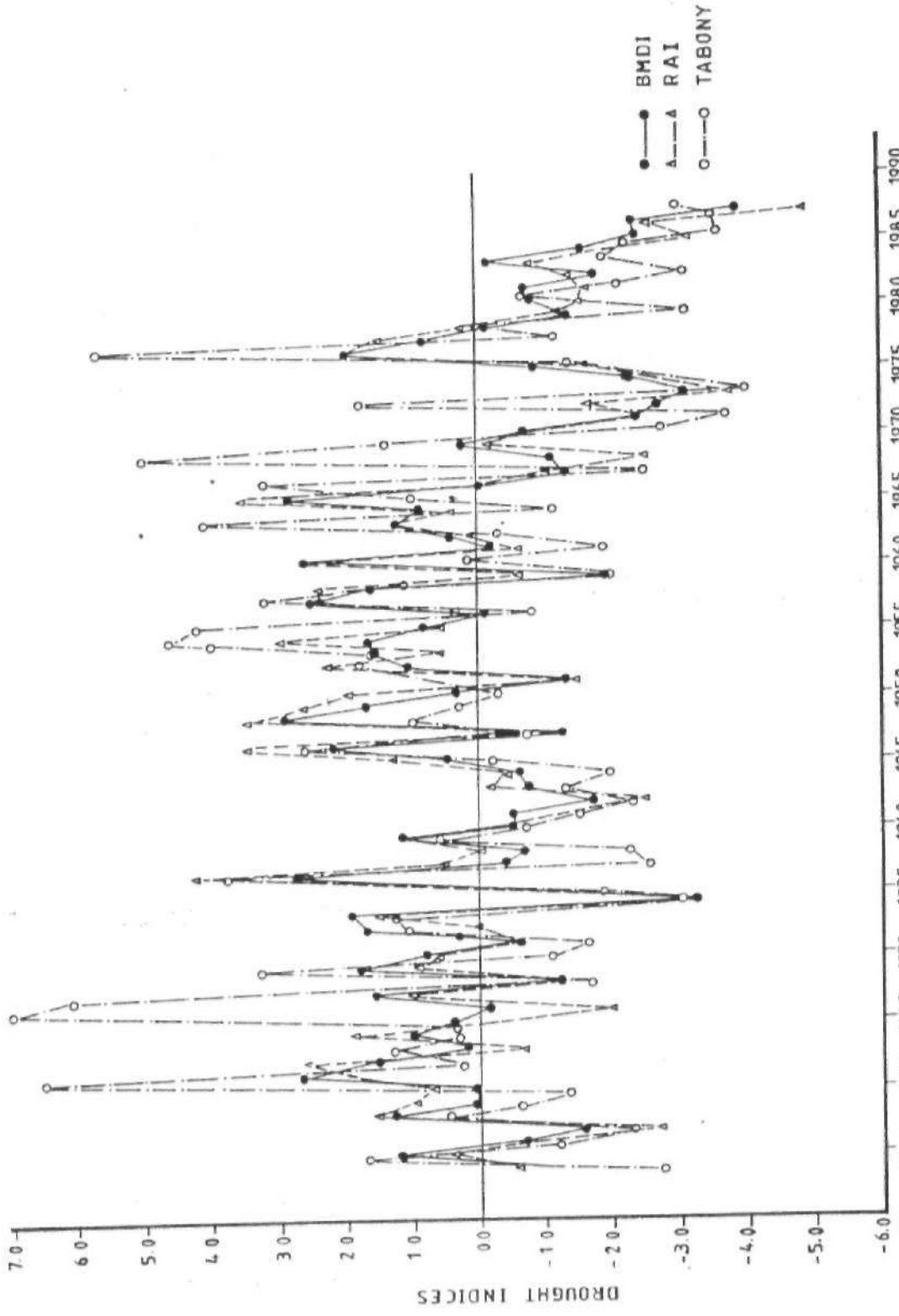


Fig. 4.2. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Sokoto (SAHELIAN ZONE)

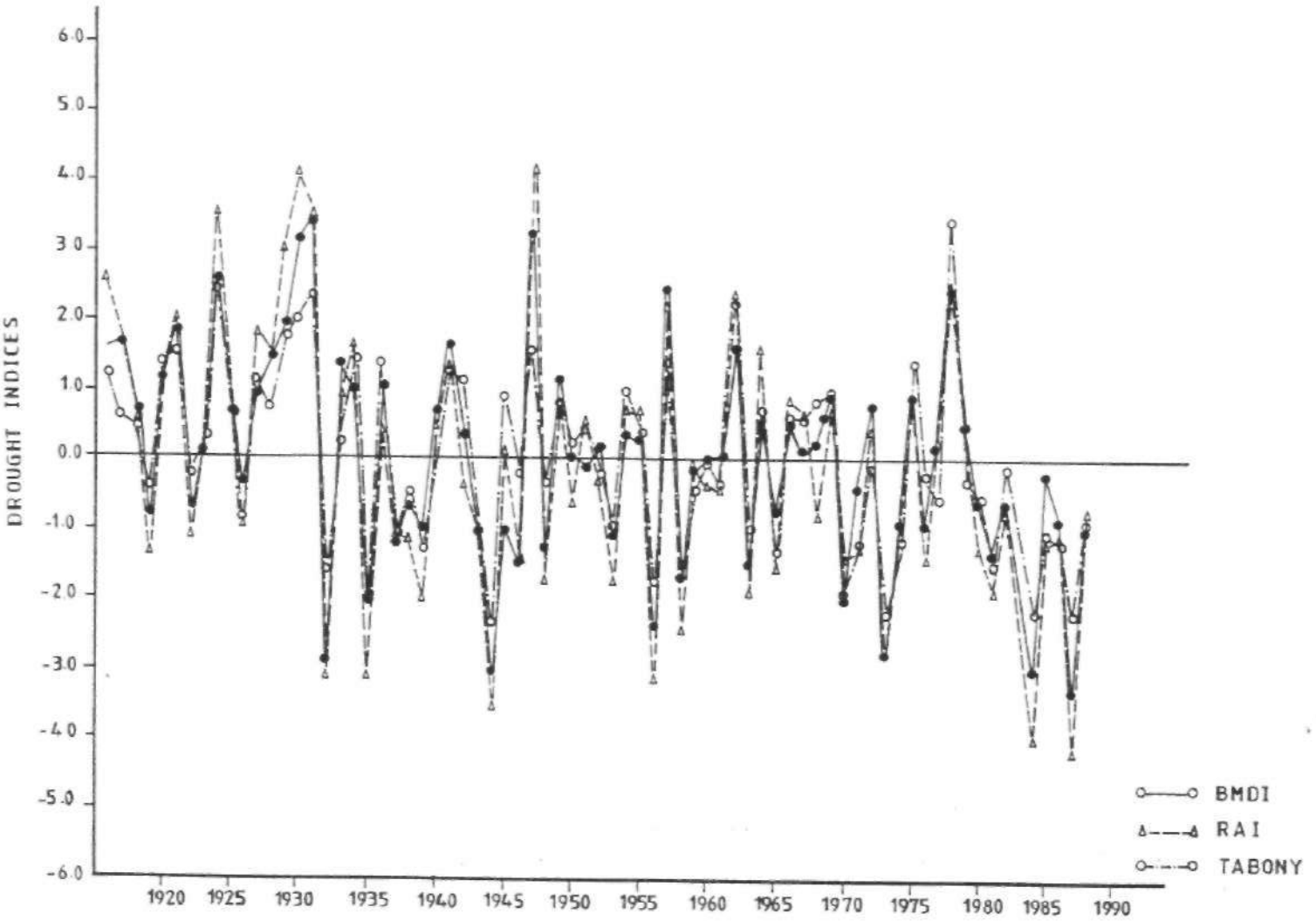


Fig. 4.3 Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Minna (MIDLAND AREA).

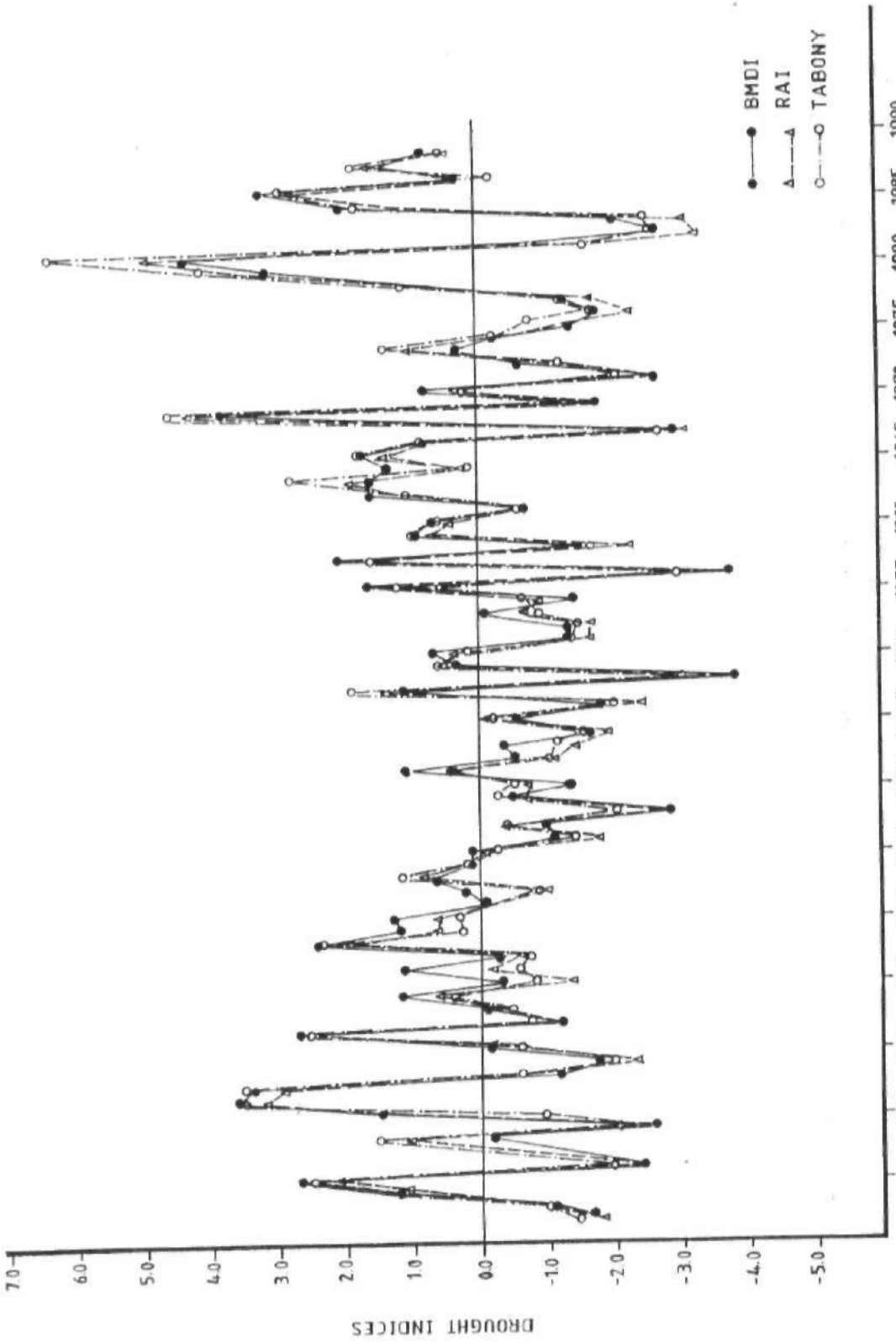


Fig. 4.4 Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Ibadan (GUINEA SAVANNA ZONE).

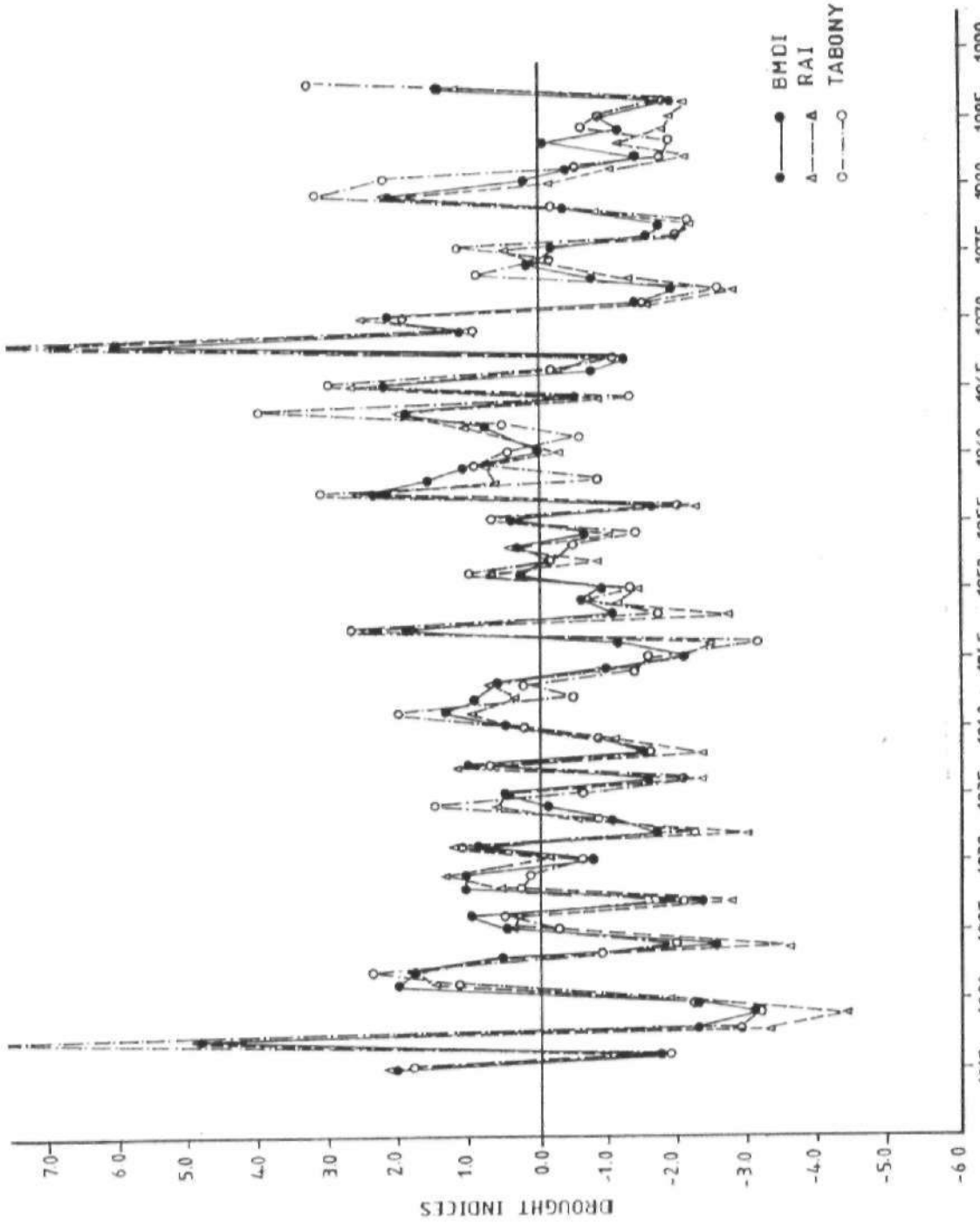


Fig. 4.5. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Lagos (COASTAL ZONE).

by all the three indices. Table 4.4 shows an example, for 1932, of the indices for some stations in the southern part of the country.

TABLE 4.4 VALUES OF BMDI, RAI and TABONY INDICES FOR SOME SOUTHERN STATIONS FOR 1932

Stations	BMDI	RAI	Tabony
Calabar	-1.6	-1.6	-2.3
Port-Harcourt	-1.5	-1.4	-3.0
Lagos	-1.8	-3.0	-1.8
Ilorin	-2.1	-2.7	-2.0

The droughts of the early 1940's that were described by Apeldoorn (1981) and Oladipo (1989) are also presented by all three indices. A typical example for 1944 for some stations is shown in Table 4.5.

TABLE 4.5 VALUES OF BMDI, RAI and TABONY INDICES FOR 1944 FOR SOME STATIONS

Station	BMDI	RAI	Tabony
Katsina	-2.5	-3.0	-1.2
Kano	-3.5	-3.3	-3.4
Minna	-3.1	-3.6	-2.4

(b) Although the three indices agree well on drought / flood conditions, certain differences in

their behaviour still exists. For example, by comparison, seasonal Tabony's index is more sensitive to rainfall deficiency and wetness in the Sahelian zone, accentuating dryness and wetness there, whereas BMDI and RAI appear to respond more slowly. Conversely, in the Midland, Guinea - Savanna and Coastal zones, it is BMDI and RAI that depict strongly extremely dry and wet conditions. These differences could be attributed to the fact that seasonal Tabony's index is based on the actual values of rainfall, whereas both the BMDI and RAI are influenced by antecedent moisture conditions.

(c) Drought index values of less than -3.0 (severe drought) for the mean seasonal drought indices are very infrequent throughout the period of analysis. Since seasonal mean values are cumulative from monthly index values this could indicate that the probability of getting seven consecutive months of severe drought is very small in the region of study.

(d) The difference between drought severity estimated by the Bhalme and Mooley procedure and the simple method of van Rooy is small, thus making RAI as useful as the more elaborate BMDI in estimating drought / flood conditions. However, BMDI responds more slowly to an increase in

precipitation than the van Rooy's index. This is in good agreement with Oladipo's (1985) observation.

Therefore, based on the above performance of BMDI, as compared to RAI, Tabony and detailed description of PDI, certain advantages of BMDI can be pointed out:

(a) BMDI is a modified PSDI. Bhalme and Mooley modified Palmer's procedure by making it suitable for the description of meteorological drought in tropical conditions. This makes it more useful to the Nigerian conditions than others.

(b) RAI and Tabony are simple indices that express the degree of rainfall anomaly for a period in relation to the long - term mean rainfall. These indices do not take into account antecedent moisture conditions. BMDI, on the other hand, using the standard deviation and coefficient of variation, takes into account the seasonal variation in precipitation and therefore is considered superior to the other two indices. Consequently, the BMDI index is used in the following chapters to depict drought situation in Nigeria.

4.5 Bhalme and Mooley Index Performance Some Examples

In order to further assess the performance of BMDI, seasonal index values for Kano, Sokoto and Maiduguri for long period of years are presented in Table 4.6 and Figures 4.6-4.8 and compared with the historical accounts of droughts and famines in the area that were discussed in Apeldoorn (1981).

TABLE 4.6 BMDI VALUES FOR KANO, SOKOTO AND MAIDUGURI

Year(s)	Kano	Sokoto	Maiduguri
1907	-1.52	-	-
1913-1914	-2.88	-1.05	-
1926-1928	-1.01	-1.16	-1.96
1942-1944	-3.49	-1.75	-2.11
1971-1973	-4.05	-3.14	-2.75
1983-1987	-3.41	-3.90	-4.16

From the Table 4.6 and Figures 4.6-4.8 it is evident that different stations in the Sahelian zone respond similarly to drought conditions with BMDI values clearly indicating this. The description of consequences and effects of droughts/famines shows clearly the following: Late and scanty rains in Kano in 1907 led to widespread scarsity and famine. The drought of 1913 and famine known as Malali was responsible for over 100,000 deaths in Kano province alone. In Bedde area of

Borno State (north-eastern part of the study region), parents gave out their children in trust in order to get food for the rest of the family. Livestock population in western Borno numbered 88,000 in 1913 reduced to only 26,000 in 1914 (Apeldoorn,1981).

Due to poor rains of 1926-1927 famine known as Mai Buhu was recorded in the region. Famine known as 'Yar Gusau or 'Yar Balange due to drought of 1942 was aggravated by wartime economy. Catastrophic droughts of early 1970s culminated in famine of 1973. Between 100,000 and 250,000 people died in the Sahel and millions became victims of famine. In Kano, for example, an overall loss of 50 to 85 per cent of all agricultural production was recorded in 1973.

The 1980s were characterized by more severe and persistent droughts than the previous decades. These droughts were characterised by migration of people in search of food and cattle grazing and massive losses of agricultural production.

The above examples of historical droughts/famines and the responce of BMDI indicates that BMDI performs well and presents weather conditions realistically.

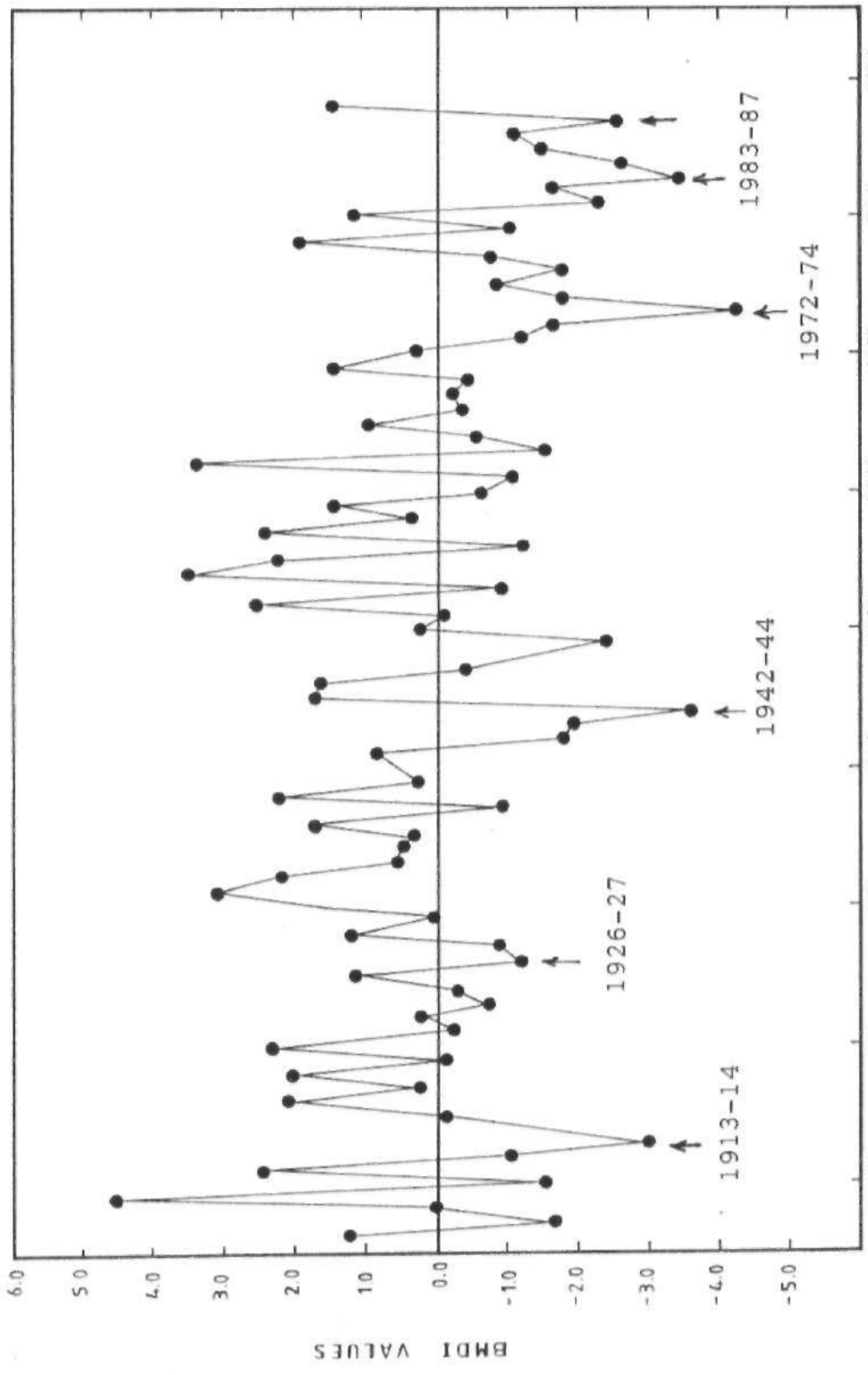


Fig. 4.6 Seasonal means of Bhalme and Mooley drought indices (BMDI) for Kano

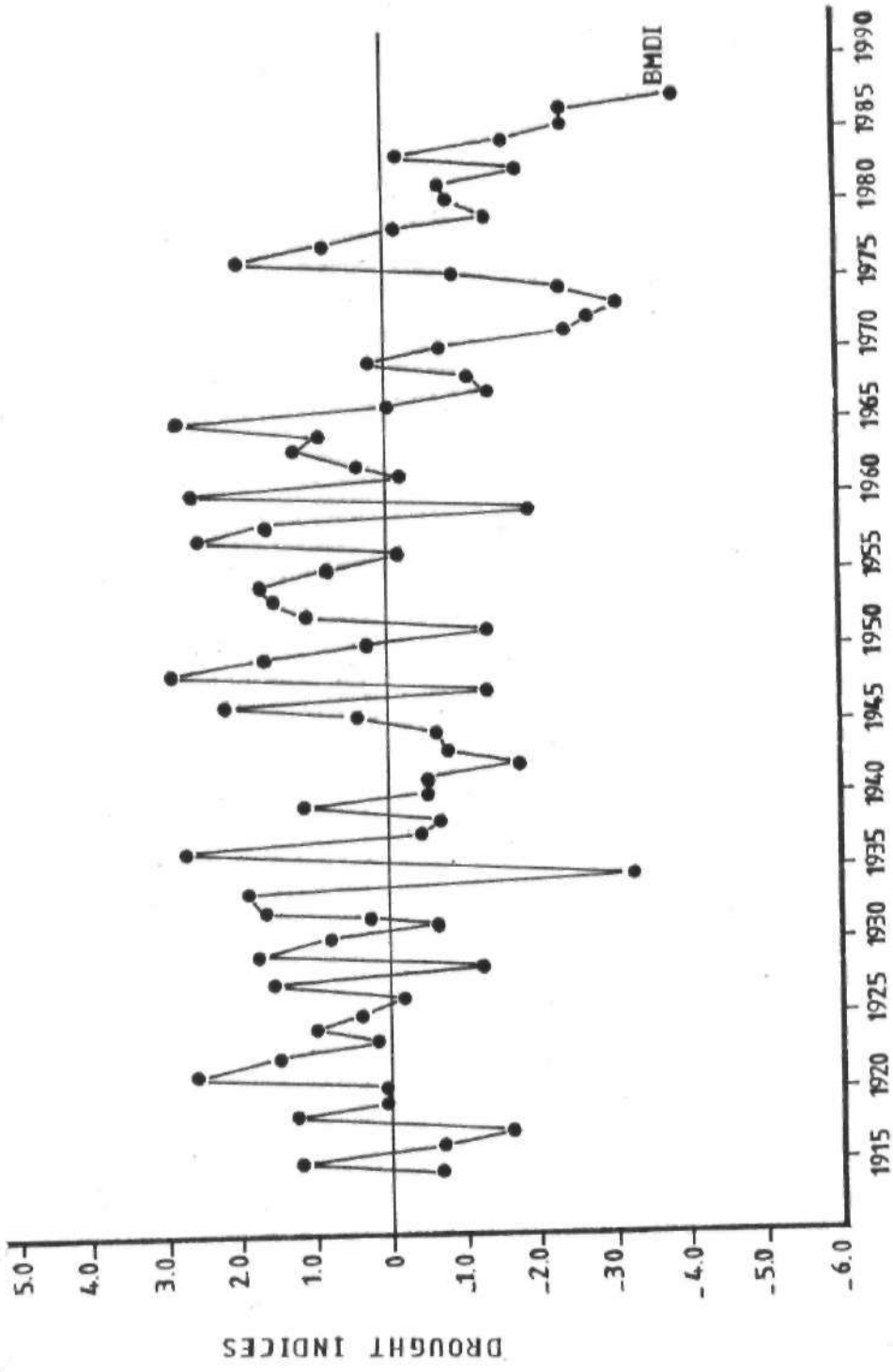


Fig. 4.7. Seasonal means of Bhame and Mooley (BMDI) drought index for Sokoto (SAHELIAN ZONE).

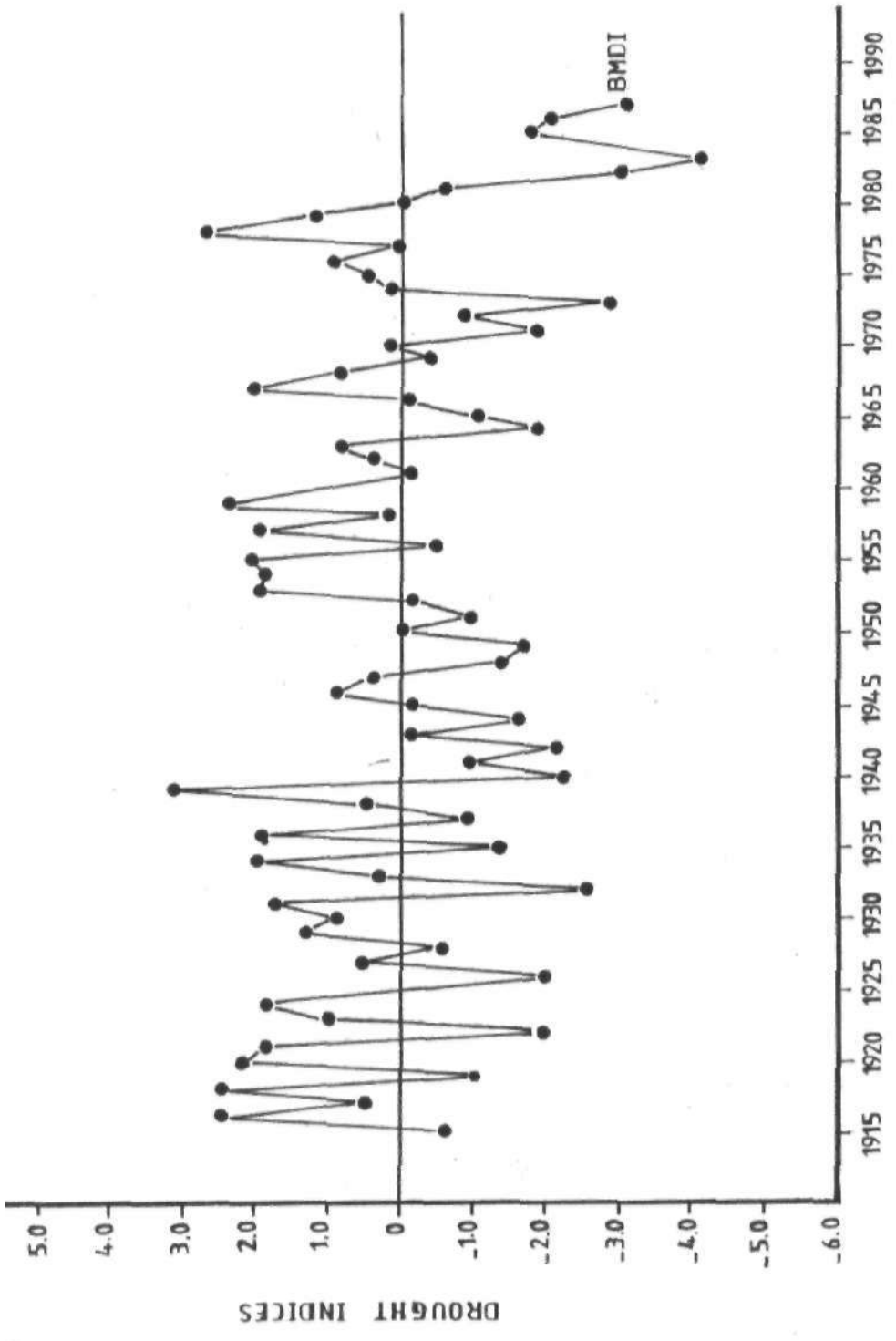


Fig. 4.8. Seasonal means of Bhalme and Mooley (BMDI) drought index for Maiduguri (SAHELIAN ZONE).

CHAPTER 5

SPATIAL CHARACTERISTICS OF DROUGHT IN NIGERIA

5.1 Introduction

This chapter focuses on the spatial distribution of drought in Nigeria, which has received little attention in the past. Such knowledge is of tremendous importance for agriculture, water resource management and development of meaningful strategies for coping with drought. Only a few investigations have approached this relevant aspect of drought study. Kowal and Adeoye (1973) examined the extent and severity of the rainfall deficit for the 1972 drought year, delineating the boundary separating areas with the maximum requirements for arable farming from those not useful for agriculture. Oguntoyinbo and Richards (1977), Adefolalu (1985, 1986a,b) and recently Olaniran (1991) examined rainfall anomaly patterns for selected wet and dry years (or groups of years) in terms of deviations from long-term mean conditions. Olaniran (1991) mapped spatial anomaly patterns and then classified them according to configurations into five spatial anomaly types. The studies of Ojo and Oyebande (1987) and Oladipo (1989), on the other hand, were different in the sense that they used drought indices to present some aspects of drought distribution in Nigeria. Oladipo (1989) established, among other things,

that patterns of moisture anomaly vary greatly from year to year with no consistent recurrent spatial pattern. All these point to the need to further evaluate the extent of spatial coherence in the occurrence of drought in Nigeria.

This chapter, therefore, examines the spatial-temporal characteristics of drought in Nigeria by using the map-pattern correlation method to establish patterns of moisture anomalies. In the following, some of the methods relevant to the present work and their relationship to the map-pattern correlation are discussed first, This is followed by a discussion of the map-pattern correlation method. The map-pattern correlation method is then applied to the BMDI series and the anomaly maps, depicting the drought type patterns, are presented and discussed. In order to establish the degree of continuity of drought in space and the degree to which wetter conditions in some areas coincide with droughts in other areas, the maps of the mean drought patterns for each year are presented and analyzed. Finally, the probabilities of drought occurrence in some sub-areas,given that it occurred in other areas, are calculated.

5.2 Review of Commonly Used Methods

Different methods of joint spatial-temporal analysis are extensively used in climatology. These include: elementary linkage analysis, factor analysis, principal

component analysis (PCA) and the map-pattern correlation method among others. The basic feature of all these procedures is the use of the matrix of similarity coefficients derived from the raw data, usually summarized in an n x m data matrix, in which n is the number of objects and m is the number of attributes (variables) by which each object is described (Ayoade, 1977). Eigenvector analysis [i.e. empirical orthogonal function analysis (EOF), principal component analysis (PCA) and common factor analysis (CFA)] are immensely popular diagnostic tools for examining patterns of climatic variations (Weare, 1990). These procedures have been employed to solve a wide variety of problems including the identification of groups of intercorrelated variables, or cases, under consideration and the identification of "causes" for the observed variation (Legates, 1991). Some of the well known applications of eigenvector analysis for representing patterns of variance of meteorological elements have been performed by Stidd (1967), Sellers (1968), Kutzbach (1967, 1970), Kidson (1975), Blasing (1975, 1978, 1981), Dyer (1977), Klaus (1978), Richman (1981, 1983, 1986), Ogallo (1980, 1989), Eklundh and Pilesjo (1990) among others. As the climatological applications for these eigenvector techniques have grown, a number of methodological discussions have appeared regarding the correct application of the techniques (e.g. Richman, 1986; Weare, 1990; Legates, 1991).

Also, the advantages and disadvantages of the method have been extensively discussed by Blasing (1975, 1978, 1981), Richman (1983, 1986) and Legates (1991). Blasing (1975, 1978, 1981) has shown that eigenvector analysis, as a representative of joint spatial-temporal variation of climate, contain certain mathematical constraints. The orthogonality constraint can be viewed as both a limitation and an advantage. Orthogonality is a desirable property when building prediction equations, since it precludes partial correlations between the derived variables, but it is a limitation, since atmosphere is not constrained to have orthogonal modes of behaviour. Another limitation is related to its "reflection" property. This simply means that an eigenvector pattern is fixed but the sign can assume both positive and negative values. In order to reduce some ambiguities, which often accompany the direct solutions, several methods can be used to adjust the frames of reference of the eigenvectors through rotations (Richman, 1986). However, determination of the number of eigenvectors to be rotated is one of the major problems of the rotation of eigenvectors (Ogallo, 1989). Richman (1986) also criticised PCA for not reflecting the spatial covariance structure observed in the data, owing to an overdependence on the shape of the domain. Richman further criticised traditional (unrotated) solutions for (i) not retaining similar eigenvector loading patterns when smaller portions of the domain are evaluated, (ii) in-

accurate representation of data interrelationships as described by the dispersion matrix and (iii) exhibiting large sampling errors when two, or more, eigenvalues are nearly identical.

An alternative technique of representing spatial-temporal variation of climatic variables is based on the use of the correlation coefficient as an indicator of similarity between two anomaly types. The method was first developed by Lund (1963), then used by Kutzbach (1966) and in modified form by Berryman (1969). Later Blasing (1975, 1978, 1980) and Nicholson (1979, 1980) applied the spatial correlation method to pressure and precipitation departure maps, respectively. Oladipo (1987) used the map-pattern correlation method to investigate spatial patterns of drought in the Interior Plains of North America.

The map-pattern correlation technique appears to have distinct advantages over PCA and is free from the constraints embedded in eigenvector analysis. Generally, for descriptive, as opposed to predictive purposes, the anomaly types determined by spatial correlation are found to be more representative of observed climatic variation and can be more readily interpreted in synoptic terms. Therefore, the present study employs the map-pattern correlation method. The next section provides a summary of the method and its application to the BMDI series.

5.3 Map-Pattern Correlation Analysis

The following description of the map-pattern correlation method is based on the works of Blasing (1975, 1978, 1980), Nicholson (1979, 1980) and Oladipo (1987). The procedure is analogous to the usual correlation of time series of two, or more, variables. In the map-pattern correlation method, however, the variables correspond to points in time (i.e. years) and the individual observations of each variable j (seasonal mean BMDI values for each station i in the year j) are functions of space. Thus, the procedure consists of linear correlation of BMDI series for the years 1906 to 1987 for all forty one stations and classification of data into generalized "moisture anomaly types". The year 1906 is the first year with data available for at least six stations which theoretically could form a type.

The map-pattern correlation method uses the least-square technique of correlating two variables. The linear correlation coefficient for two variables x_1 and x_2, as given by Panofsky and Brier (1968), is:

$$r_{12} = \frac{\overline{x_1 x_2} - \bar{x}_1 \bar{x}_2}{\left[(\bar{x}_1^2 - \bar{x}_1^2)^{\frac{1}{2}} \right] \left[(\bar{x}_2^2 - \bar{x}_2^2)^{\frac{1}{2}} \right]} \qquad (5.2)$$

where $\overline{x_1 x_2}$ is covariance value between $\bar{x}_1$ and $\bar{x}_2$. Substituting X_{ij} and X_{ik} for x_1 and x_2, and expanding the average in (5.2), the coefficient of linear correlation between the years j and k is given as:

$$r_{jk} = \frac{N_{jk}^{-1}(\sum_i X_{ij}X_{ik}) - (N_{jk}^{-1}\sum_i X_{ij})\ (N_{jk}^{-1}\sum_i X_{ik})}{\left[N_{jk}^{-1}\sum_i X_{ij}^2 - (N_{jk}^{-1}\sum_i X_{ij})^2\right]^{\frac{1}{2}} \left[N_{jk}^{-1}\sum_i X_{ik}^2 - (N_{jk}^{-1}\sum_i X_{ik})\right]^{\frac{1}{2}}} \quad (5.3)$$

where Σ_i denotes the summation over all the stations available in the years j and k. X_{ij} and X_{ik} are respectively the seasonal BMDI values for station i in the years j and k, and N is the number of years.

The computed correlation coefficients measure the similarity between two drought patterns, with a high positive value implying that the degree of moisture anomaly is similar in both years. BMDI values for each year were correlated with those of other years. This produced a symmetric matrix of correlation coefficients with both rows and columns corresponding to years.

5.3.1 Determining Drought Patterns

In order to determine which of the years is highly correlated with the greatest number of other years, the correlation matrix was examined. For this purpose a high correlation coefficient (one equal to, or greater, than a cutoff value r_c) is determined from the matrix (Blasing, 1978). There is still no general agreement on the choice of the threshold correlation coefficient (r) to determine what is highly correlated. The selection of (r) depends upon the pattern being correlated, the size of the region and the consistency in the number of

years being correlated. Because of the missing data and the variable length of the drought series, the number of stations N_{jk} involved in each correlation varies greatly. Therefore, according to Nicholson (1979) significance level, rather than a critical minimum value of the correlation coefficient, must define "highly correlated". The method of Nicholson (1979), whereby (r) was defined in terms of the correlation coefficient reaching a specified significance level, was used in this study.

The year j, with the most correlation coefficients at .001 significance level was first selected. This year was averaged with the moisture anomaly patterns for three years best correlated with year j to produce "Anomaly Type Pattern 1" (recorded in the group A). Blasing (1975, 1978) noted that the choice of the number of years to be averaged into one type is arbitrary and generally should depend on the number of observations and degree of generality (spatial resolution) desired in the anomaly types. The averaging process was intended to smooth out the "noise" of the individual moisture anomaly patterns, while retaining the common features in the Type 1 pattern.

Correlation coefficients between the Type 1 pattern and each year's drought map pattern were then computed. Each year which correlated positively with the Type 1 pattern to the .001 significance level was then removed from the correlation matrix to form group B.

In order to determine higher order types, the procedure was repeated using the remaining correlation data (years not classified already as moisture anomaly Type 1). The level c_i defining "highly correlated" had to be lowered from .001 in successive iterations, once no other year in the remaining data correlated with at least three other years at that significance level. Significance level c_i defining "highly correlated", was lowered first to .005, then to .01, .02, .03, .05 in successive experiments, each with $N_{j,k-1}$ degrees of freedom. Those observations which have high negative correlations with Type pattern were identified from the matrix to form a group C of anomaly type. According to Blasing (1975) a large number of such observations would indicate that a "reflection" of the Type pattern is realistic. Thus, these observations, having high negative correlations with the Type pattern will be positively well correlated with each other.

After all types were identified, as described above, each remaining unclassified year, was searched for the highest positive correlation it had with any type and classified as this type. Other years which had not yet been classified were also examined for significant negative correlations with the identified type patterns.

5.3.2 Pattern Maps

Using the above procedure six moisture anomaly patterns were identified. These Types classify thirty-eight out of eighty-two years (46%). Results of the analysis are given in Table 5.1 and in the maps in Figures 5.1 to 5.6.

The first anomaly type (Figure 5.1) depicts a generalized situation whereby areas north of about 10°N (Sahelian and northern part of the Midland zones) experience near normal and wet conditions, while the area between latitudes 7°N and 10°N show moderate drought conditions. Normal moisture conditions prevail along the coastal areas. A possible explanation for this moisture anomaly type is that there is a strong northward incursion of the ITD, such that major zones of rainfall activity (zones C and D) have a long residence time over northern Nigeria, while zone E with its stratiform clouds stagnates over southern Nigeria. There is, therefore, a general weakening of convective activity behind the ITD within the monsoon in the southern part of the country (southern Midland area, Guinea-Savanna and northern Coastal zones). This weakening, or inhibition, of convective activity due to strong inversion, or large scale subsidence has been discussed earlier by Kidson (1977), Ojo (1977), Lamb (1978a), Nicholson (1981), Achempong (1982, 1987) and Olaniran (1991). Coastal

TABLE 5.1

RESULTS OF DROUGHT PATTERN TYPING

A: four years averaged together to form anomaly

B: additional years classified as the type

C: years negatively correlated with the type

c: significance level

Anomaly type	A	B	C	c
1	1936, 1946 1956, 1958	-	1928, 1949, 1973	.001
2	1932, 1933, 1945, 1954	1950	-	.005
3	1941, 1968 1980, 1985	1917	1920, 1925	.01
4	1974, 1971, 1972, 1987	1942	1977	.02
5	1951, 1923, 1926, 1969	1908,1943	-	.03
6	1961, 1965 1975, 1976	-	1986, 1907 1927	.05

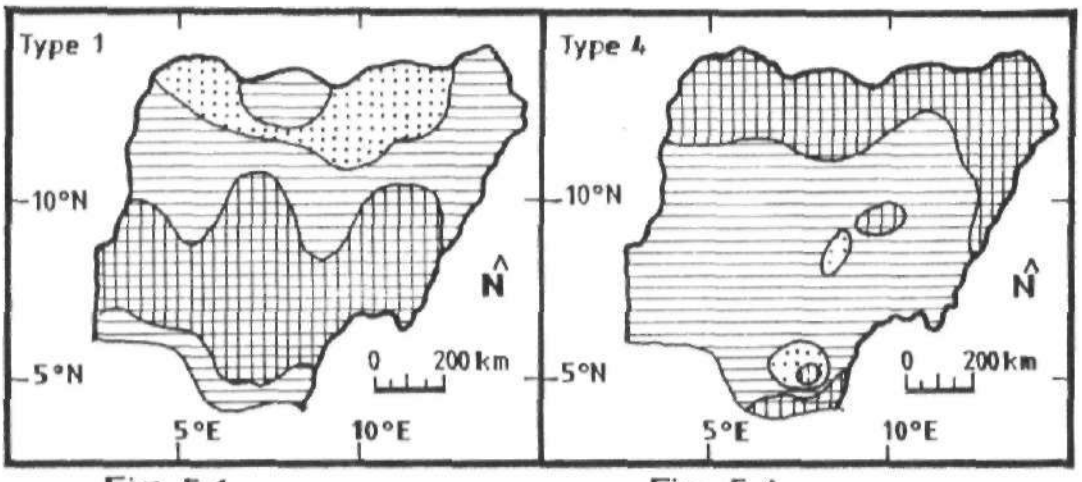


Fig. 5.1.

Fig. 5.4.

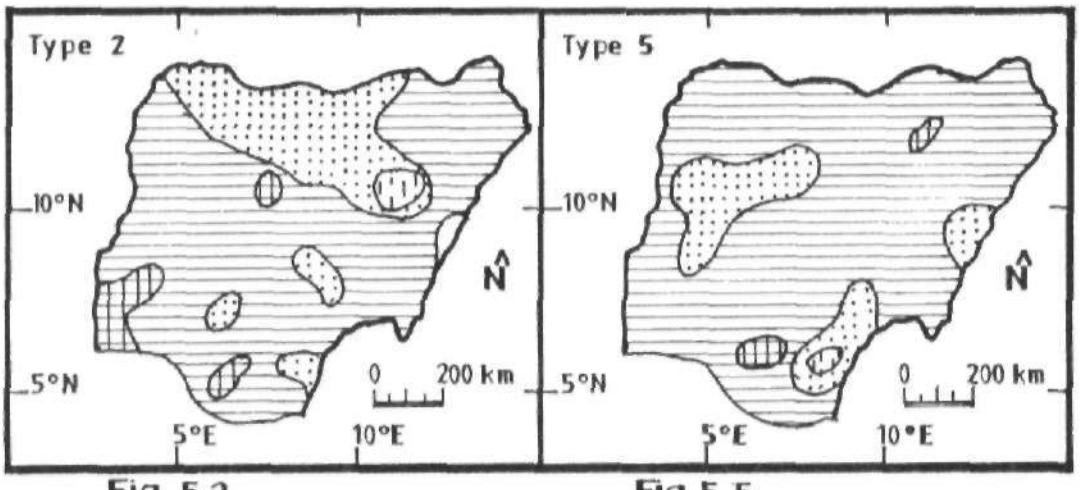


Fig. 5.2.

Fig. 5.5.

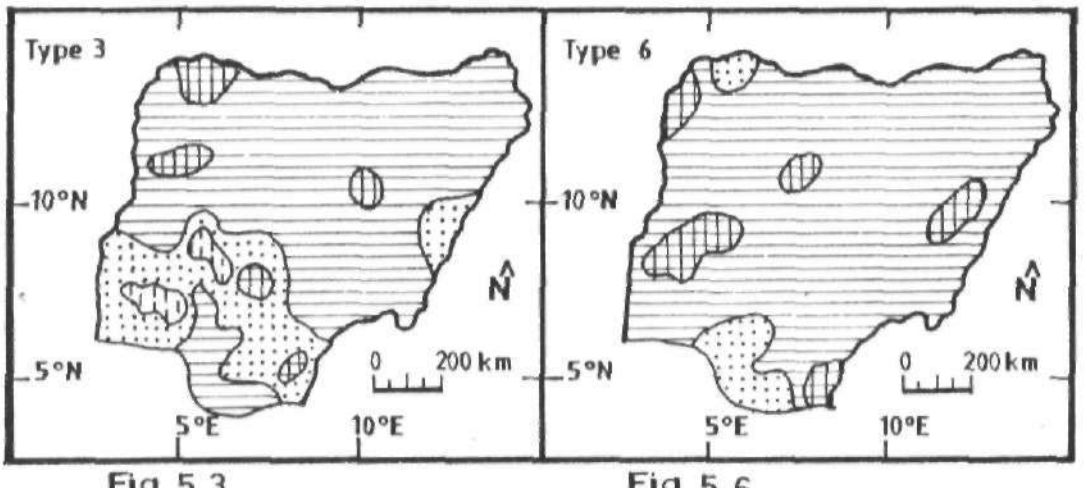


Fig. 5.3.

Fig. 5.6.

Figures 5.1 - 5.6: Mean moisture anomaly patterns (Types 1-6).

LEGEND

- Very wet (BMDI > 3.0)
- Moderately wet (3.0 ≥ BMDI > 1.0)
- Near normal (1.0 ≥ BMDI > -1.0)
- Moderate to severe drought (-1.0 ≥ BMDI > -3.0)

areas in this moisture anomaly type still show near normal moisture conditions, probably due to local coastal configurations and mesoscale features. Olaniran (1991) identified a similar moisture condition in Nigeria as Type 3 in his study.

Type 2 (Figure 5.2) can be viewed as a variant of Type 1. It depicts near normal, to wet, to very wet conditions north of latitude 10°N (Sahelian zone and northern Midland area). Unlike the first type there is only a small area of mild drought conditions over the south-western and coastal subarea. Similar to Type 1, the basic explanation is the sufficient northward incursion of the ITD with long residence of zones C and D over the northern subregion. Even though there is a small inversion over the south-western part, generally normal convective activities are likely to occur south of the ITD.

Figure 5.3 shows the third moisture anomaly pattern (Type 3) for Nigeria. In this situation conditions are predominantly wet south of latitude 9°N with some localized extremely wet conditions over the western Guinea -Savanna and Coastal areas. Mild drought affects some areas north of latitude 10°N. Consistent with the ITD model, if the ITD advances sufficiently, but not too far, northward and the convective activity is not inhibited south of the ITD, the situation presented by Type 3 will occur.

Type 4 (Figure 5.4) features extensive drought

conditions north of latitude 11°N and localized drought situations south of latitude 5°N along the eastern part of the Coastal area. The rest of the country is dominated by near normal conditions. This type corresponds to a situation when the ITD has a restricted northward advance and short residence time. In this situation the major rainfall producing zones (zones C and D) may stagnate over the central area of Nigeria, thereby permitting zone E weather to move a short distance inland from the Gulf of Guinea. This produces drought conditions in the Sahelian zone, northern Midland and parts of the Coastal areas of Nigeria, but near normal conditions elsewhere. This corresponds to Olaniran's (1991) moisture anomaly Type 1.

Type 5, as depicted in Figure 5.5, is dominated by near normal conditions with isolated moderately wet areas. One such area is the western part of the Midland zone, the other two moderately wet areas are in the extreme eastern parts of the Midland and Guinea-Savanna and Coastal zones. Moderate drought is found only in insignificant proportions.

Type 6 (Figure 5.6) also shows near normal conditions dominating the country. Instead of isolated areas of moderately wet conditions, Type 6 features isolated areas of drought in all zones. The central Coastal region features an extensive area of moderately wet conditions, while wet conditions are also present around So-

koto. Types 5 and 6 thus depict a situation that shows normal northward advance and southward retreat of the ITD over the country.

In general, the above results indicate four basic modes of moisture anomaly consistent with the ITD model in Nigeria. The first mode involves periods of increased precipitation north of latitude 10°N and drought and near normal conditions southwards (Types 1 and 2). While the second is an opposing mode, which depicts periods of drought north of approximately 10°N to 11°N and along the eastern coast with near normal conditions elsewhere. This is represented by the Type 4 pattern. The third mode of moisture anomaly, represented by the Type 3 pattern, involves periods of wet to very wet conditions south of approximately latitude 10°N and near normal to dry conditions north of that latitude. The fourth mode involves moisture patterns featuring predominantly near normal moisture conditions in the whole region, but with localized areas of moderate to severe drought and wet to extremely wet conditions (Types 5 and 6).

Based on the analysis of type patterns certain observations can be put forward:

(i) There is a marked spatial variability of moisture conditions in Nigeria as depicted by Type patterns.

(ii) Areas north of latitude 10°N (the Sahelian zone and the northern Midland area) appear to show

more homogeneous moisture patterns than areas south of latitude 10°N.

(iii) There is a marked tendency for departures of opposing signs to occur north and south of latitude 10°N. This confirms the findings of Klaus (1978), and Nicholson (1979), who identified latitude 10°N as a major zone of discontinuity in Africa.

(iv) Finally, western subareas of all zones and extreme eastern subareas of the Coastal and Midland zones show greater spatial-temporal variability of moisture conditions than other regions.

In order to judge the validity of the typing scheme used in the study, and to show that the anomaly types are non-random combinations of four years, but representative of moisture anomaly conditions in Nigeria, the correlation between individual years and the types, the degree of correlation among years averaged to form each type and the correlation between various types were computed.

Figure 5.7 shows the correlation of each year from 1906 to 1987 with each of the types. Years constituting the Type show high degree of correlation with the type and the years falling into groups B and C. However, as is shown in Figure 5.7 some years possess high correlation coefficients, but were not classified into the type. The explanation for this could be in the mode of choosing years that are "highly correlated". As pointed out by

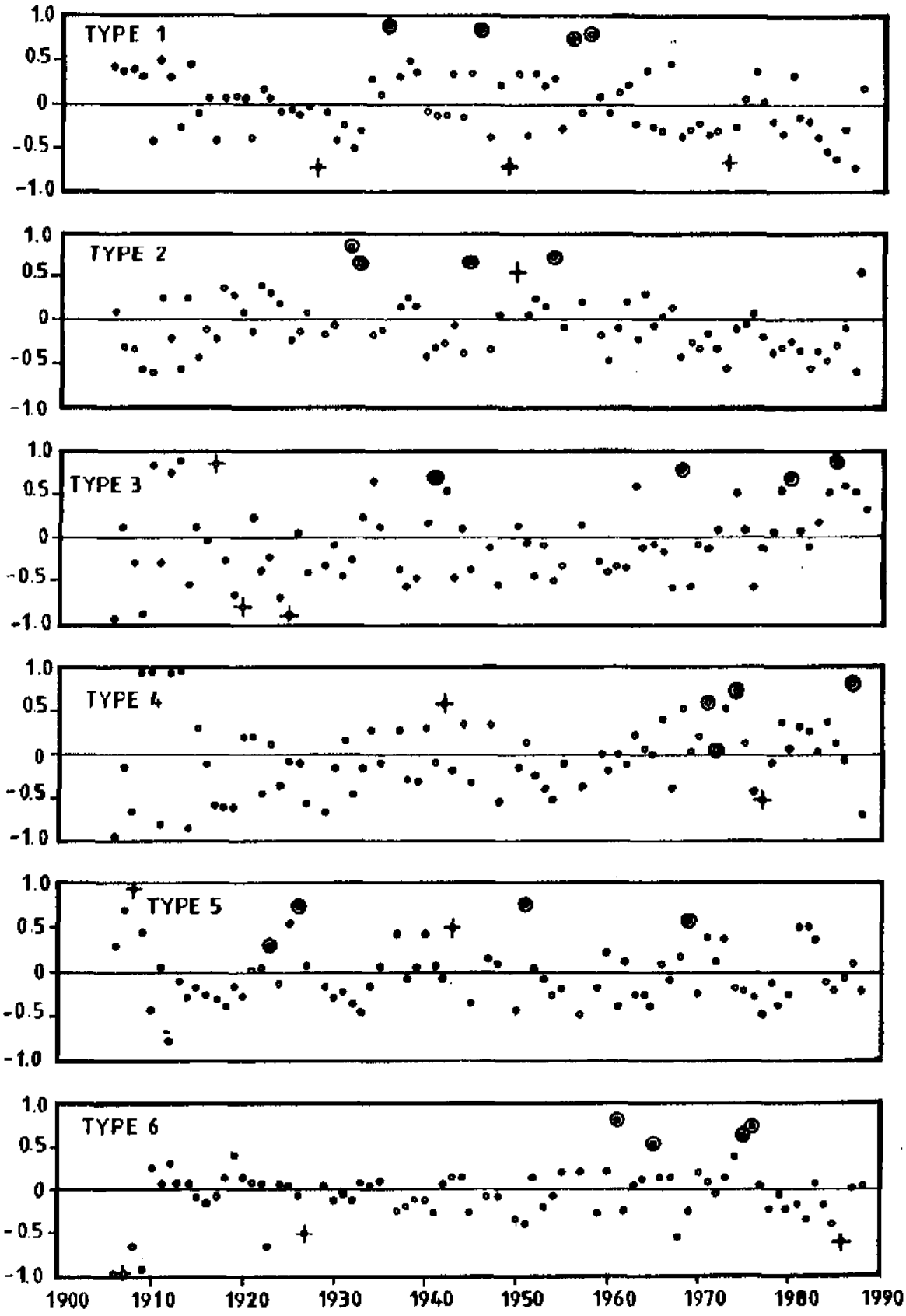


Fig. 5.7. Linear correlation of seasonal moisture anomaly patterns for the years 1906-1987 with anomaly types 1-6; ● denotes year classified as type, + denotes year classified as groups B & C in each case.

Nicholson (1979), types are formed according to correlation coefficients above a particular significance level calculated on the basis of N-1 degrees of freedom, where N is the number of stations involved in a particular correlation. If the number of degrees of freedom is reduced,or increased, the correlation between the type and the years classified as type would be more or less significant.

Table 5.2 shows the intra-type correlation for each of the six drought pattern types and the type correlation with each year that constituted the type. From the Table it is evident that principal years of each type are highly correlated with the remaining years of the type, but these years do not always correlate well with each other. Type correlation with each year constituting the type correlate well in nearly all cases. The intra-type correlation coefficients for the first three types are high and positive suggesting that they are validly classified. The intra-type correlations for the remaining types 4, 5 and 6 are generally slightly lower, but the difference, especially in Type 4 is not large enough to suggest they are invalid. There is an occurrence of negative values only in two cases (Types 4 and 6).

Table 5.3 shows the inter-type correlation coefficients between the types. It is clear that there is no significant correlation between any two types. Also positive correlations are few which indicates that types are independent and therefore valid.

TABLE 5.2 INTRA - TYPE CORRELATION COEFFICIENTS

TYPE 1

	1936	1946	1956	1958
1936*	1.0	0.727	0.580	0.654
1946		1.0	0.363	0.488
1956			1.0	0.410
1958				1.0
Type 1	0.905	0.866	0.767	0.785

TYPE 2

	1932	1933	1945	1954
1932*	1.0	0.486	0.577	0.542
1933		1.0	0.146	0.289
1945			1.0	0.320
1954				1.0
Type 2	0.896	0.668	0.690	0.723

TYPE 3

	1941	1968	1980	1985
1941	1.0	0.216	0.215	0.549
1968		1.0	0.280	0.598
1980			1.0	0.539
1985*				1.0
Type 3	0.721	0.822	0.727	0.932

* principal year

TABLE 5.2 (cont.)

TYPE 4

	1971	1972	1974	1987
1971	1.0	0.425	0.431	0.107
1972		1.0	0.390	-0.173
1974*			1.0	0.587
1987				1.0
Type 4	0.615	0.036	0.760	0.713

TYPE 5

	1923	1926	1951	1969
1923	1.0	0.212	0.473	0.018
1926		1.0	0.610	0.153
1951*			1.0	0.452
1969				1.0
Type 5	0.319	0.770	0.796	0.618

TYPE 6

	1961	1965	1975	1976
1961*	1.0	0.436	0.382	0.369
1965		1.0	-0.152	0.010
1975			1.0	0.533
1976				1.0
Type 6	0.821	0.535	0.661	0.742

TABLE 5.3

INTER-TYPE CORRELATION COEFFICIENTS

	Types					
	1	2	3	4	5	6
1	1.0	0.271	-0.403	-0.407	-0.123	0.087
2		1.0	-0.167	-0.164	-0.005	0.096
3			1.0	-0.065	-0.117	-0.323
4				1.0	0.286	-0.116
5					1.0	-0.508
6						1.0

As discussed by Nicholson (1979) and Oladipo (1987) when judging the validity of types the coherency of departure patterns must be taken into account. Areas of negative, or positive, departures, generally uninterrupted by areas of the opposite sign, suggest a valid anomaly type, rather than a random combination of four years. Large areas of departures well above, or well below normal might also suggest a valid typing procedure. Conversely, a map dominated by departures in the normal to near normal range may be invalid, as departures are relatively small and negative and positive values may cancel out, leaving the positive correlations among years to be determined mainly by small regions with large departures. These criteria suggest that Types 5

and 6 may not be valid. This assumption, however, may not be the case here as will be shown later.

The arbitrariness involved in selecting the number of years to be averaged to form the type and the number of years classified could be viewed as a shortcoming of the typing scheme, or an indication of unrepresentativeness of the patterns. The present study used groups of four years, because of the inability to form bigger groups at very high significance levels. As discussed by Blasing (1975) and Nicholson (1979) setting a lower limit for "significant" correlations, particularly for the first two anomaly types, might have helped to increase types representativeness, but it would risk losing potentially meaningful anomaly types. The six anomaly types in this study classified 46 per cent of the years with group B classifying only few years. The explanation for this must be sought in the nature of the data correlated. Precipitation data, as pointed out by Namias (1968) and Nicholson (1979) is more difficult to classify (no matter by what method) than temperature,or pressure data, due to its high variability in space. Namias (1968) showed, for example, that whereas three eigenvectors account for 81 per cent of the variance of winter temperature over the U.S. (from 1899 to 1939), twenty eigenvectors were needed to explain the same amount of precipitation variance (five eigenvectors explained only 45 per cent of the variance of precipitation).

Nicholson (1979) classified 56 per cent of the years in her study of African rainfall. Oladipo (1987) classified 61 per cent of the years in his study of spatial coherence of drought in Interior Plains of North America.

The analysis, therefore, is fairly successfull in classifying moisture anomaly patterns and determining various frequently occurring precipitation anomaly types. Types 1, 2, 3 and 4 could be considered reasonably coherent, statistically valid and acceptable as a realistic presentation of actual climatic conditions in Nigeria. Predominantly near-normal moisture conditions indicated by types 5 and 6 is also believed to be a realistic representation of the actual climatic conditions in Nigeria.

5.3.3 Interannual Variations

In order to delineate the major features of moisture anomaly further, maps of different moisture conditions for each year from 1906 to 1987 were produced. Maps for 1906 to 1912 were not included here because a large number of the stations have missing data for these years. Their use, therefore, could lead to a misrepresentation of the actual situation. Interannual variations for some selected years are presented in Figure 5.8 and Appendix B.

An examination of the seasonal maps of moisture anomaly provides evidence of considerable changes in the patterns, size and position of areas of dryness and

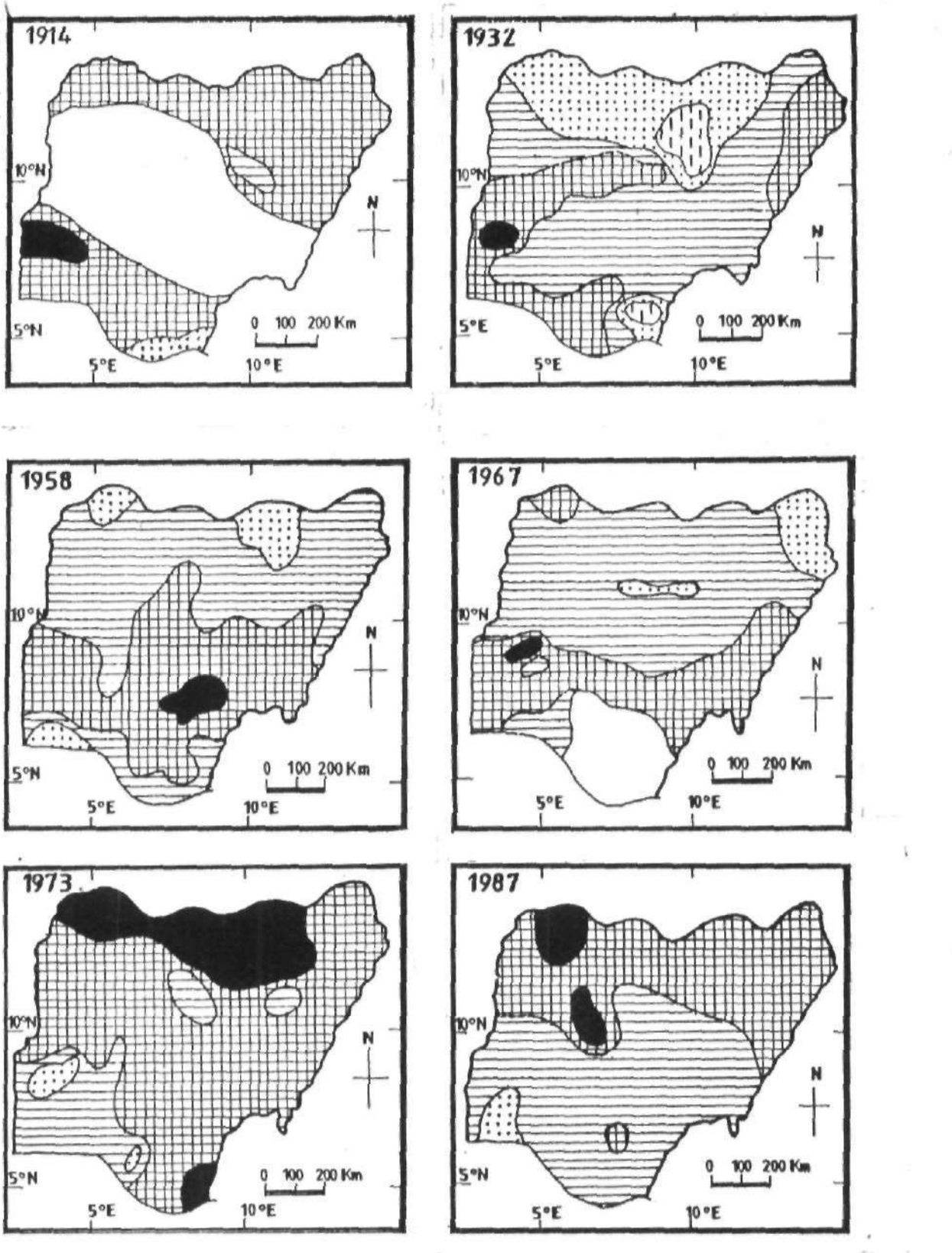


Fig. 5.8. Interannual variations of moisture anomaly patterns for some selected years.

wetness from year to year. Even for the years of country -wide occurrence of normal and above normal moisture conditions there are always localized areas having at least mild drought. In fact, areas of mild and severe drought covered at least 11 per cent of Nigeria in about 43 per cent of the years. The prominent periods of drought in Nigeria, as shown on the maps, were 1913-1914, 1922, the 1930s (1932, 1936-1938), the 1940s (1942-1946, 1948,1949), 1950, 1956,1958,1961, 1967, the 1970s (1970--1973, 1975-1977) and the 1980s (1981-1987). In general, the moisture anomaly patterns are very different in the configurations, position and areal extent.

The droughts of 1913-1914 were a culmination of the trend towards increased aridity which began in the last five years of the 19th century (Apeldoorn, 1981). Unfortunately, only a few stations have data for that period and this makes spatial resolution of drought occurrence very difficult . Drought index values of selected stations for 1913-1914 are shown in Table 5.4 and they clearly indicate the occurrence of drought.

In 1914 moderate to severe drought conditions existed over most of the country. These droughts and consequent famines, are well documented by Mortimore (1973, 1989), Grove (1973), Apeldoorn (1981) and Oladipo (1991). According to these sources famine due to drought, known in Hausaland as Malali raged as wide and devastating as the disaster of the early 1970s which drove thousands of people from the countryside into the cities. Deaths during

this famine in Kano province alone may have been of the order of 100,000 (Apeldoorn, 1981). Nomadic herders were particularly badly hit. One such group in western Borno numbered 10,000 with 88,000 cattle in 1913, but by 1914 only 5,500 remained with 26,ooo cattle (Grove, 1973).

TABLE 5.4 DROUGHT INDICES FOR SOME SELECTED STATIONS IN NIGERA FOR 1913-1914

Stations	1913	1914
Kano	-2.9	-1.6
Sokoto	-	-1.1
Yola	-	-2.1
Ondo	-0.9	-2.5
Bauchi	-1.7	0.0
Owerri	-1.7	-1.5
Ibadan	-	-2.2
Afikpo	-1.4	-2.1

The 1920s were generally of normal to wet conditions although some small localities, particularly in the western Guinea-Savanna zone and Coastal area of the country (with the exception of 1929), were always affected by mild drought.

The droughts of the 1930s were similarly localized and mild in nature with the exception of 1932, 1936 to 1938, when extensive areas of moderate drought occurred south of latitude 10°N. Thus, drought conditions affected about 25 per cent of Nigeria in 1932, 18 per cent, 17 per cent and 16 per cent in 1936, 1937 and 1938 respectively.

Droughts of the 1940s were more serious in terms of their areal extent and severity. Starting from 1942 to the end of the decade, with only a little respite in 1947, moderate to severe droughts affected large areas of the country. The drought of 1942 manifested itself in famine, known in Hausaland as 'Yar Gusau, or 'Yar Balange and affected around 30 per cent and most of the Sahelian zone of Nigeria. In 1943 the western parts of the Midland and Guinea-Savanna zones and parts of the Coastal area (approximately 28 per cent) were affected by drought of moderate to severe intensity. Another major event of countrywide (around 70 per cent) occurrence of moderate to severe drought took place in 1944. This drought was the most extensive of the decade. The other droughts during the 1940s include those that occurred in 1945, 1946 and 1948 which were similar in spatial extent and intensity to that of 1943. In 1949 moderate to severe drought again affected over 30 per cent of the Sahelian and northern Midland zones.

The 1950s were generally wet, although several years in that decade show widespread droughts. For example, moderate to severe drought conditions affected over 26 per cent of the country in 1950, while about 16 per cent was affected in 1952. This was followed by a rather wet period in 1953 to 1955. Drought returned in 1956, when about 68 per cent of the country experienced drought conditions. The year 1957 was exceptionally wet and no part of the country was affected by drought. This changed in 1958, when drought

returned to over 34 per cent of the country, but in the following year the drought's areal extent contracted to about 18 per cent. These droughts affected mostly the southern Midland, Guinea-Savanna and northern Coastal areas.

The 1960s featured generally normal to wet conditions, except in 1961, 1965 and 1967 when mild to severe drought affected 18 per cent, 16 per cent and 24 per cent of the country respectively. Interestingly the last two years of that decade were rather normal, with only isolated areas (6 per cent in 1968 and 3 per cent in 1969) of mild and severe drought. These years are often referred to as the beginning of the Sahelian catastrophic droughts. This would indicate that while drought started in earnest in 1968 in the northern part of the West African Sahel, the southern sector, which includes northern Nigeria, still enjoyed normal rainfall conditions.

The 1970s witnessed the beginning of a new decade of increased dryness. In 1970, moderate to severe drought affected over 50 per cent of the country, mainly the Sahelian, Midland and western Guinea-Savanna zones. In 1971, moderate drought affected 32 per cent of the country especially the Sahelian zone and western Coastal area. Areas of moderate drought decreased slightly in 1972 to around 23 per cent, but 1973 features a return to a countrywide (approximately 84 per cent) occurrence of drought with the Sahelian zone being affected by severe drought conditions. Only the western part of the Guinea-Savanna and Coastal zones

experienced near normal moisture conditions. Severe drought affected the extreme eastern Coastal area. The drought of 1972-1973 formed part of the catastrophic West African Sahel drought during which between 100,000 and 250,000 people reportedly died and nearly 3,5 million cattle

In 1974-1975 there was a slight respite in the intensity and areal coverage of the drought, with only about 13 per cent of the country affected in 1974 and 16 per cent in 1975. Moderate to severe drought returned again in 1976 and 1977 to cover 23 per cent and 74 per cent of the country respectively. In 1976 moderate to severe drought affected the western Guinea-Savanna and Coastal areas, the extreme eastern Coastal area and central parts of the Sahelian and Midland zones, while in 1977 drought was countrywide with large areas of severe drought conditions in the Sahelian, Midland and Guinea-Savanna zones.

The period 1978-1981 was dominated by near normal conditions. Drought returned again in the early 1980s and features on all maps until 1987. In 1982, moderate to severe drought covered 67 per cent of the country from the coast to the northern border, with the exception of the western Midland and Guinea-Savanna and extreme eastern Guinea-Savanna zones. In 1983, over 70 per cent of the country was covered by drought, with nearly 20 per cent affected by severe drought condi-

tions in the Sahelian and Midland zones. This makes the year 1983 the worst of the 1980s. In 1984 again over 60 per cent of the country was experiencing drought. The entire Sahelian and northern half of the Midland zones (areas north of latitude 10°N) and the entire Coastal area were affected by moderate to severe drought conditions. In 1985, 1986 and 1987 fifty eight per cent, 52 per cent and 47 per cent of the country respectively was covered by drought. This drought consistently affected areas north of latitude 10°N and in 1986 also the Coastal area.

Figure 5.9 shows the percentages of the subareas affected by extreme and moderate drought, normal, wet and very wet conditions for each of the years represented in Appendix B (from 1913-1987). It is evident from Figure 5.9 that near normal conditions predominate most of the time, except in some selected years, although the area of near normal conditions significantly decreased in the 1980s. Areas affected by drought, on the other hand, have increased in extent since the beginning of the 1970s with a consequent decrease in near normal and wet areas. The fact that for most of the time near normal moisture conditions generally dominated the region of study could be viewed as an indication of the validity of moisture anomaly Types 5 and 6 produced by the map pattern correlation method.

Therefore, although drought episodes always alternated with normal and wet years in the past (i.e. before the 1950s), the last two decades with their severe and widespread droughts on a yearly basis (particularly in the

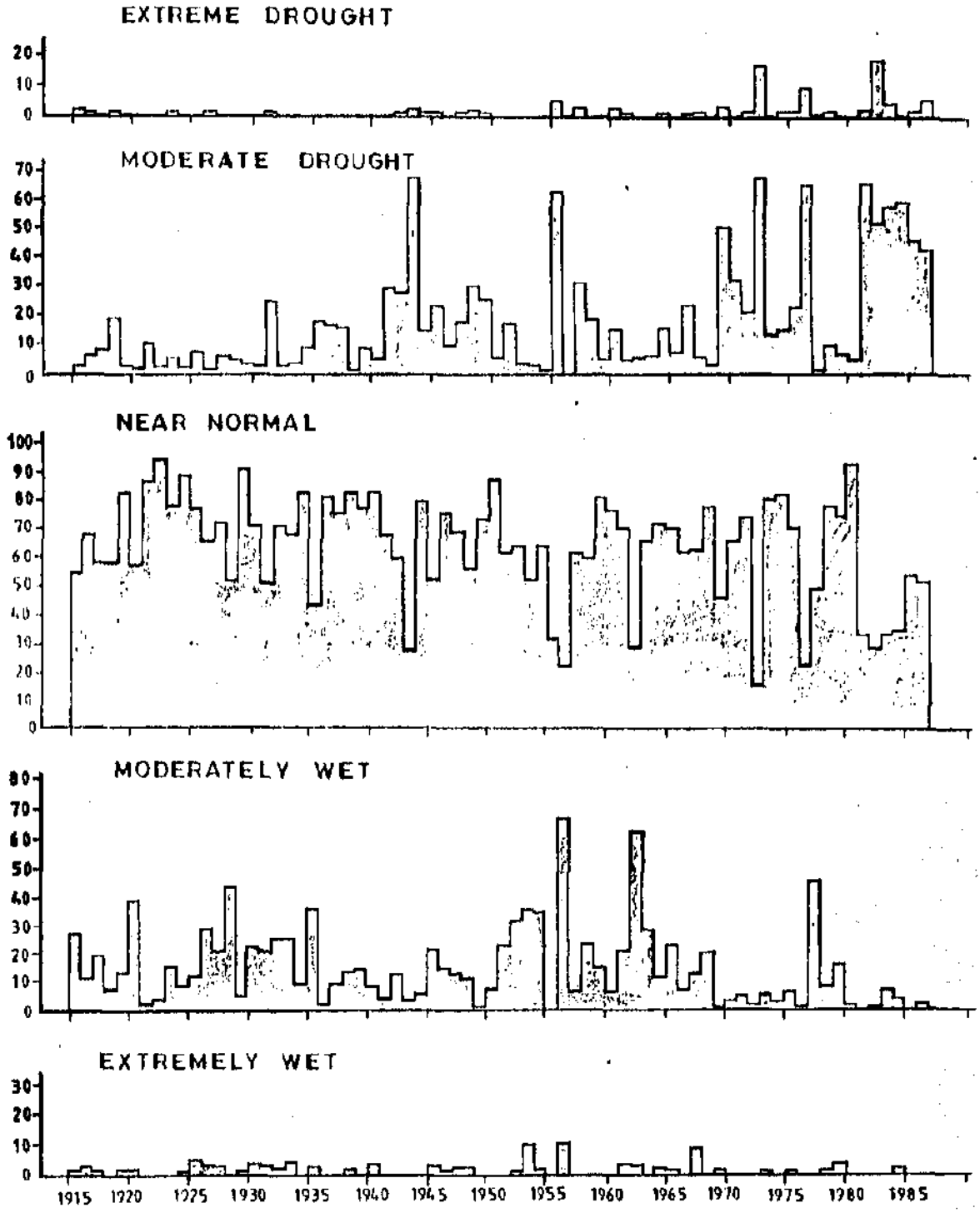


Fig. 5.9 Percentages of the subareas affected by the extreme and moderate drought, normal, wet and very wet conditions.

Sahelian and northern Midland zones) suggest that a dangerous trend towards increasing aridity is taking place.

Table 5.5. shows large-scale drought and wet years, taken as years when over 40 per cent of the region is affected by moderate to severe drought, or wet to extremely wet conditions. Only about 16 per cent and 8 per cent of the years could be identified as large-scale dry, or wet years respectively. An interesting observation from Figure 5.9 and Table 5.5 is the presence of consecutive wet years, although the 1953-1955 period shows extensive (over 32 per cent) areas of wet conditions. This also indicates that unlike periods before the 1960s and even the 1970s, when runs of large-scale dry years were interrupted by wet and normal years, the 1980s persistently showed large scale droughts.

5.4 Changing Frequency of Drought

The empirical probabilities of occurrence of moderate and severe droughts were calculated for each station. On the whole, the values show big variations among individual stations, especially in Guinea-Savanna and Coastal zones. In Guinea-Savanna, for example, Ibadan, Oyo, Offa and Ilorin in the western part of the zone show distinctly higher probabilities of drought occurrence than the rest of the zone. This observation of high probability of drought in the southwestern part of the Guinea-Savanna zone can be attributed to reduced rainfall resulting from inversions, or large scale subsidence, that was discussed in 5.3.2. This sub-area, therefore, appears to be susceptible to frequent drought

TABLE 5.5

YEARS WITH LARGE-SCALE DROUGHT AND WET CONDITIONS

Years	Area of drought		Years	Area of wetness	
	%	sq km		%	sq.km
1913	45.0	415,800	1921	40.9	377,916
1914	66.0	609,840	1929	43.9	405,636
1944	69.6	643,104	1936	38.8	358,512
1956	68.1	629,244	1957	78.5	725,340
1970	52.7	486,948	1963	65.7	607,068
1973	83.5	771,540	1978	47.0	434,280
1977	74.0	683,760			
1882	66.0	609,840			
1983	70.2	648,648			
1984	60.2	556,248			
1985	58.3	538,692			
1986	46.5	429,660			
1987	46.8	432,432			

occurrencies even when other areas experience normal moiture conditions.

To bring out some regional patterns in the frequency of drought in Nigeria, the mean probabilities of drought occurrence for each of the zones were determined. The areally averaged seasonal values of BMDI for each zone and year was calculated. Only years in which drought indices were available for a minimum of 50 per cent of the stations were utilized. From these area mean values of BMDI, the frequencies of occurrence of moderate and severe droughts were used to determine their empirical probabilities and recurrence intervals. The computed probabilities are taken as the mean probabilities of drought occurrence in different climatic zones.

The mean probabilities for moderate drought for the Sahelian, the Midland, the Guinea-Savanna and the Coastal zones are 0.27, 0.24, 0.26 and 0.25, respectively, with a recurrence interval of between 3 and 4 years. Severe drought, on the other hand, are not frequent. For example, Sahel has a mean probability of severe drought 0.03 (recurrence interval of 33 years), the Midland and Guinea-Savanna 0.02 (recurrence interval of 50 years) and the Coastal zone 0.03 (recurrence interval of 33 years). Surprisingly, the rainy Coastal zone shows rather high frequency of drought occurrence. Apart from already mentioned inversions and large scale subsidence within zone E, leading to reduced rainfall in July-August, another explanation for the high frequency of

drought occurrence in the Coastal zone could be found in the inherent property of the BMDI. Because the BMDI is strongly influenced by antecedent moisture conditions, severe August breaks in any year would produce negative seasonal BMDI values to indicate drought conditions. However, the impact of drought occurrencies in Coastal zone must not be compared to those in the more arid zones like Sahel, where rainfall total is small and highly concentrated within few rainy months. Thus, any slight reduction in the rainfall of any year in the dry zones could have a dramatic effect on the already fragile food and agricultural situation.

In order to indicate the periods when droughts were more, or less, frequent, the probabilities of droughts on decadal intervals were calculated for all the stations and mean probabilities for the four climatic subdivisions were obtained. The results are presented in Table 5.6.

It is clearly seen from the Table 5.6 that the frequency of droughts increases toward the recent decade in all zones, although the picture is more pronounced in the Sahelian and the Midland zones. The Guinea-Savanna and the Coastal zones show high frequency of drought occurrencies in the 1930s and 1940s. Yet, there is an evident increase in the frequency of droughts in these two zones in the last three decades. The 1980s were characterised by an exceptionally high frequency of drought conditions with nearly all the years being affected, particularly in the Sahelian and the Midland zones. Other decades of relatively high frequency of drought are the 1940s and 1970s in all zones.

TABLE 5.6 CHANGING PROBABILITIES OF MODERATE TO SEVERE DROUGHT IN NIGERIA. RECURRENCE INTERVALS IN YEARS ARE GIVEN IN BRACKETS.

Subdivision	1911-20	1921-30	1931-40	1941-50	1951-60	1961-70	1971-80	1981-87
Sahel	0.28 (3.6)	0.18 (5.6)	0.12 (8.3)	0.31 (3.2)	0.06 (16.7)	0.23 (4.3)	0.40 (2.5)	0.84 (1.2)
Midland	-	0.12 (8.3)	0.20 (5.0)	0.32 (3.1)	0.22 (4.5)	0.24 (4.2)	0.29 (3.4)	0.51 (2.0)
Guinea-Savanna	0.28 (3.6)	0.10 (10.0)	0.39 (2.6)	0.45 (2.2)	0.28 (3.6)	0.25 (4.0)	0.26 (3.8)	0.30 (3.3)
Coastal	0.30 (3.3)	0.20 (5.0)	0.24 (4.2)	0.36 (2.8)	0.21 (4.8)	0.13 (7.7)	0.35 (2.9)	0.38 (2.6)

These are in a good agreement with the drought distribution depicted earlier on the spatial maps. Also, there is a clear indication of drought persistence in Nigeria since the 1970s.

5.5 Spatial Coherence of Drought

The nature of the drought's spatial coherence in Nigeria was further examined by evaluating the probability of simultaneous incidence of drought in different stations. Nine stations from different geographical zones were selected for this experiment. Each of these stations was chosen in turn as a base and years when it was affected by moderate to severe drought were determined. Then, the probabilities of drought occurrence in other stations, when the base station experienced moderate to severe drought, were computed. The procedure was repeated for nine representative stations in the country.

Because the mean probabilities of moderate/severe drought for the Sahel, the Midland, the Guinea-Savanna and the Coastal zones were 0.30, 0.26. 0.28 and 0.28, respectively (see Section 5.3.4), it was assumed that stations with at least 40 per cent probability of being afflicted by moderate to severe drought when the base station experienced drought of similar magnitude are taken to indicate a significant tendency to more dry and less wet conditions. Conversely, stations with 30 per cent, or less, probability of moderate to severe drought when the base station experienced similar drought magnitude are assumed

to indicate a strong tendency towards wet moisture conditions. The results of the analysis for the selected station are illustrated in Figure 5.10 a-i and are discussed below.

When Sokoto, taken as a base, experiences drought, the whole of the Sahelian zone has a significant probability of the occurrence of moderate to severe drought (Figure 5.10a). There are two other areas of probable drought conditions in the central Midland and the extreme eastern Coastal area. This pattern resembles the anomaly Type 4 produced by the map pattern correlation method and is characteristic of the droughts of the 1970s and 1980s. During these periods drought affected the Sahelian and the northern Midland zones, but normal to wet conditions (except for some isolated areas) prevailed elsewhere in the country.

The percentage probability of drought when Kano experienced drought is shown in Figure 5.10b. It shows the probability of drought extending south of latitude $10^{o}N$ into the Midland and Coastal areas. The extreme western Sahelian and Midland zones are not affected by drought.

The coherence of moderate to severe drought when Maiduguri is taken as the key station show that significant probability of drought exist north of latitude $10^{o}N$ (the Sahelian and northern Midland zones), but not in the western Sahelian zone (Figure 5.10c). There is a significant probability that the western Guinea-Savanna and the extreme eastern coast of the country will simultaneously

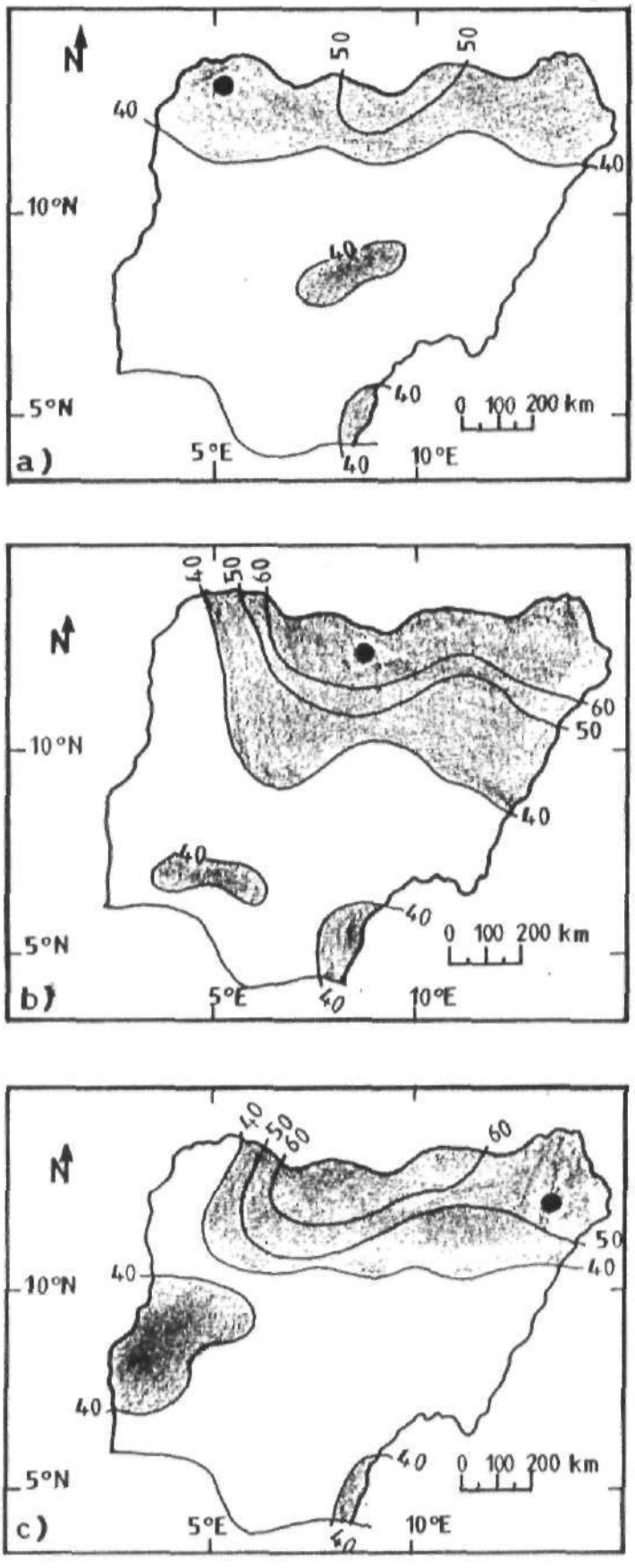


Fig. 5.10. Percent probability of drought in Nigeria when a) Sokoto, b) Kano, and c) Maiduguri EXPERIENCED DROUGHT

experience drought.

Figure 5.10d shows the percentage probability of drought when Bauchi experienced drought. This suggests a significant probability of moderate to severe drought affecting the Sahelian, Midland and northern Guinea-Savanna zones east of longitude 5°E. However, when Minna experiences drought (Figure 5.10e) the probability of moderate to severe drought is significant in a limited area of the central Sahelian, Midland and Guinea-Savanna zones, while there is a strong tendency to wetter conditions in the rest of the country. Moderate to severe drought affecting Yola suggests a significant probability of drought between longitudes 5°E to 10°E in the Sahelian zone and in the eastern Midland area. These are coherent with droughts affecting south-western Nigeria (Figure 5.10f).

When Zaria, taken as a base, experiences drought (Figure 5.10g), the pattern of significant probability of other areas simultaneously having drought resembles Figure 5.10c. In this case the Sahelian and northern Midland zones east of longitude 5°E and the western Guinea-Savanna and Coastal zones will experience drought, while wetter conditions exist elsewhere.

A significant probability exists for the southern part of the Midland zone and the Guinea-Savanna zone to experience drought when there is drought in Ibadan (Figure 5.10h). Also, drought will affect the Coastal and

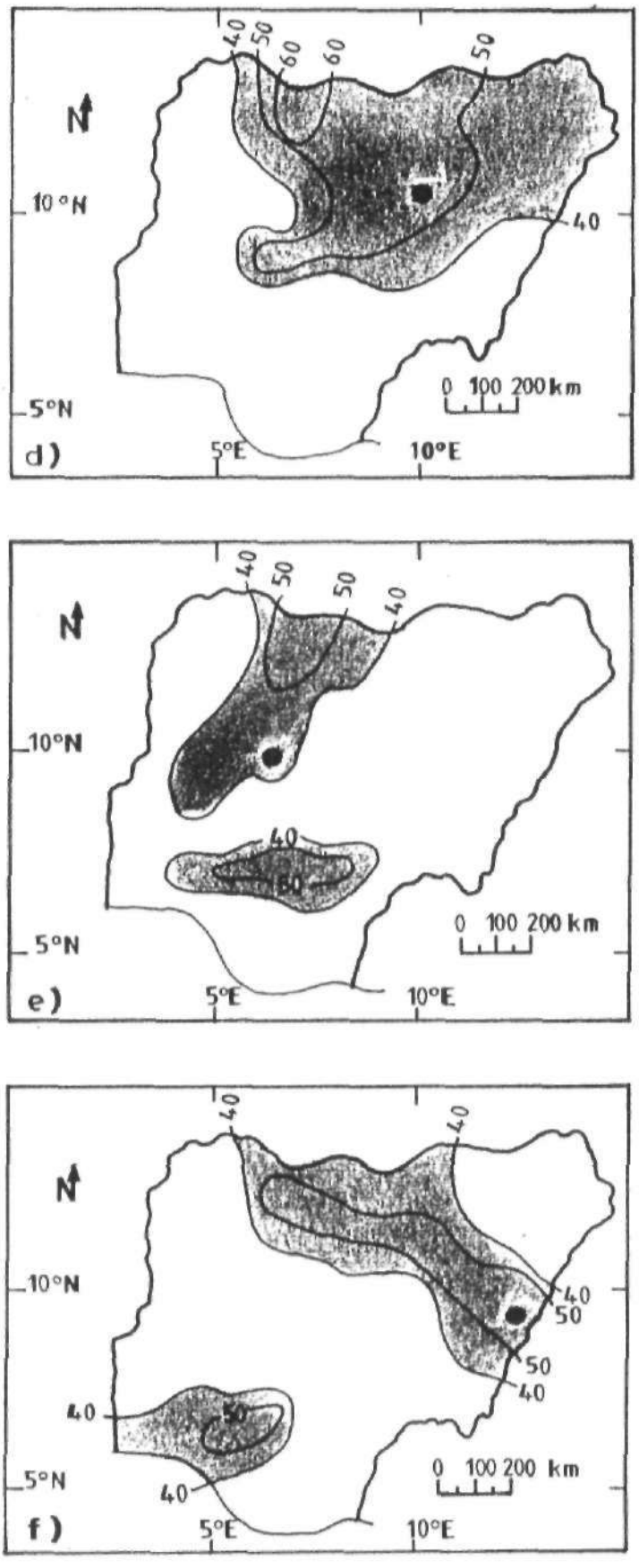


Fig. 5.10. (cont.) The same except for d) Bauchi, e) Minna, and f) Yola.

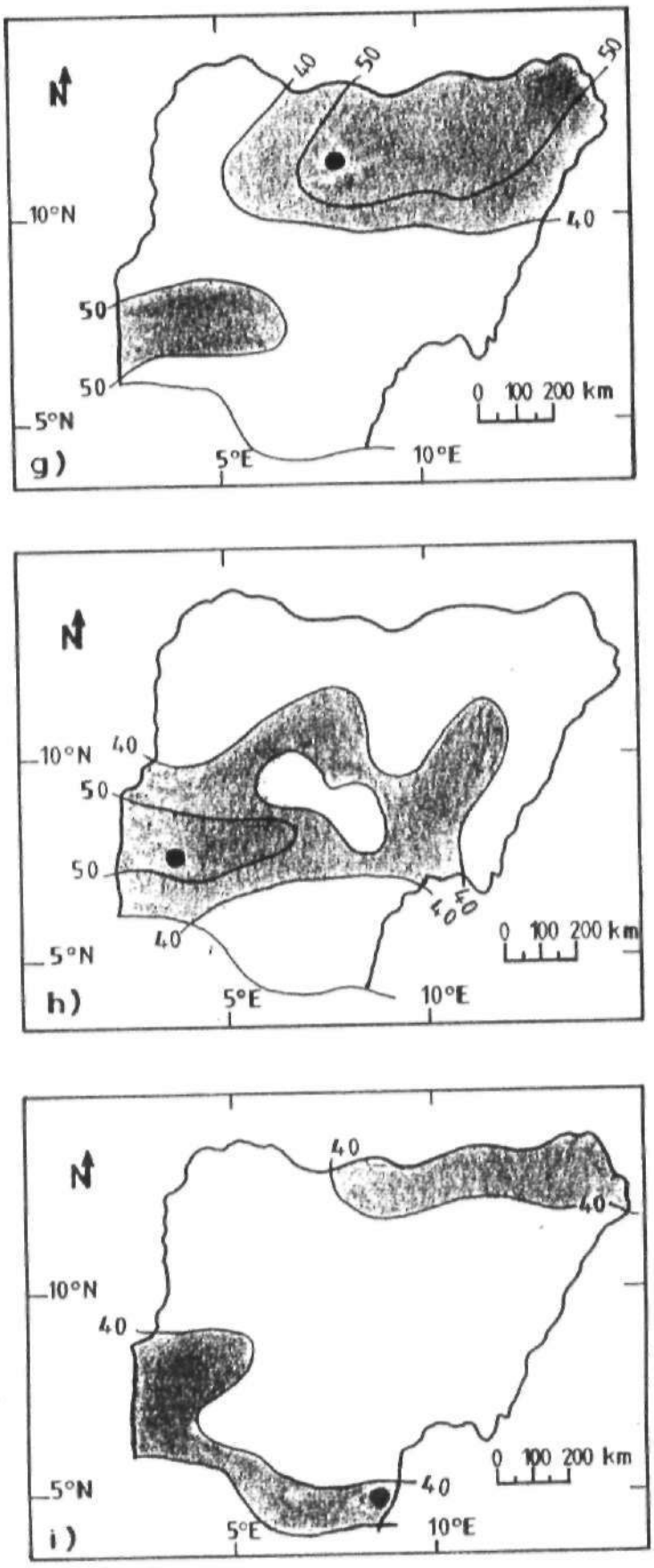


Fig. 5.10. (cont.) The same except for g) Zaria, h) Ibadan, and i) Calabar.

western Guinea-Savanna zones simultaneously with the north-eastern part of the Sahelian zone when drought conditions exist in Calabar (Figure 5.10i).

Overall, when any of the Sahelian stations (e.g. Sokoto, Kano, Maiduguri) and northern Midland stations (Zaria, Bauchi) experience drought a significant probability (over 50 to 60 per cent) exists for extensive areas of the Sahelian and Midland zones to be affected by drought. However, when drought affects the extreme north western part of Nigeria, areas with significant probability of drought appear to be less spatially extensive than when droughts affect the central and eastern Sahelian and Midland parts. Secondly, the coherence of moderate to severe drought when northern and north eastern parts of Nigeria experience drought shows strong tendency to wetter conditions in the western subareas of northern Nigeria (Figure 5.10b,c,d,g). There is, therefore, an opposing east-west moisture gradient particularly pronounced in the Sahelian and Midland zones. This difference between the western and eastern Sahel and Midland zones lies in the WNW-ESE surface orientation of the ITD, that can produce a situation whereby the extreme western subareas will be within the rainfall zone behind the ITD, while the eastern Sahel and Midland zone remains dry.

Droughts affecting the south western part of Nigeria could also be extensive (Figure 5.10h), covering most of the Guinea-Savanna and western Coastal zones.

In conclusion, the extent and spatial patterns of

drought in Nigeria indicates that the probability of the whole region suffering from drought and complete crop failure is insignificant. The noisy spatial patterns also indicate that different areas may be affected by drought at different times, although the extent and persistence of droughts in the recent decades suggest different forcing factors (i.e. global sea surface temperature anomalie and land cover changes).

Chapter 6

TEMPORAL CHARACTERISTICS OF DROUGHT IN NIGERIA

6.1 Introduction

Drought in Nigeria is a recurrent phenomenon and knowledge about its periodic occurrence is very important. This knowledge will add to our capability for drought prediction which could be beneficial for better agricultural planning.

Searches for rainfall periodicities and trends in West African rainfall have been the subject of several studies (Ayoade, 1973; Bunting et al., 1976; Ogallo, 1979; Nicholson, 1983, 1985; Farmer and Wigley, 1985; Dennet et al., 1985 and Hutchinson, 1985 among others). It was established that the total rainfall is characterized by fluctuations with low periodicities and that there has been a trend towards increasing aridity.

Recently considerable attention has been focused on the investigation of the nature of non-random rainfall fluctuations in Nigeria. Olaniran and Sumner (1990) using spectral analysis, investigated the long-term variability of Nigerian rainfall in terms of periodicities, fluctuations and trends in the onset, retreat and duration of the rainfall season. They established among other things that the mean dates and duration of rainfall in Nigeria (Coastal, Guinea-Savanna, Midland and Sahel zones) exhibit evidence for quasi-2-year, quasi-3-year, quasi-4-year and quasi-6-year oscillations.

Adejuwon et al. (1990) examined trends and fluctuations of annual rainfall and four critical seasonal components of rainfall: wet season, onset, peak and retreat data for six-

teen stations in Nigeria for the period 1922-1985. They confirmed the negative trends for the series and general tendency towards increasing aridity. Spectral analysis indicated prominent periods of between 2- and 8-year cycles, associated with the cyclical behavior of Quasi-Biennial Oscillation (QBO) (interaction between sea-surface and the air above it) and El Nino Southern Oscillation (ENSO). Other periodicities observed include 16-, 18-, and 21-year cycles.

Anyadike (1993), using power spectrum analysis, examined non-random changes of Nigerian rainfall such as fluctuations and trends. Power spectrum analysis revealed the existence of significant oscillations in the rainfall totals with time periods of 2.00-2.82 years in the middle belt and 8.00-9.60 years in all the regions.

The works of Olaniran and Sumner (1990), Adejuwon et al. (1990) and Anyadike (1993) used the traditional spectral method of Blackman and Tukey (1959). This investigation is different from the above mentioned ones in the sense that it is applying the low-pass filter technique and a newer spectral analysis method, known as the Non-Integer method (NI), to a long series of drought index (BMDI), calculated in Chapter 4. The objective is to examine the apparently stochastic characteristics of BMDI series in order to detect trends and cyclic variations and to examine the hypothesis that drought, as a stochastic process, contains periodic, or near periodic, components.

In this chapter the low-pass filter technique for detecting marked fluctuations in drought index series is presented and discussed first in section 6.2. Next, the characteristics

of three spectral methods are discussed in section 6.3, with particular emphasis on the Non-Integer technique. The results of the low-pass filter application to the BMDI series and the Non-Integer method of spectral analysis are discussed in section 6.4.

6.2 Low-Pass Filter Method

Smoothing, or filtering, of temperature and rainfall data has been a popular technique in time-series analysis. Simple moving averages (running means) in particular have been widely used to smooth, or filter, time series. Although appropriate in many instances, several significant problems are associated with the use of simple moving averages. These were discussed by Mitchell et al. (1966), Pittock (1983), Karl (1988), Kite (1989), Fuenzalida and Rosenbluth (1990) and Howarth and Rogers (1992) among others.

Howarth and Rogers (1992) summarized the performance of 5-, 10-, and 20-year moving averages and compared them with the results of the binomial filters. They concluded that although moving averages possess certain advantages (i.e. they are simple to calculate and explain, produce more precisely defined cutoff in the main lobe of the window, more sensitive to recent information at the end of the time series), their disadvantages must be taken into account when selecting a technique for smoothing the series. Major problems associated with equally weighted moving averages can be summarized as follows:

(a) the presence of high amplitude fluctuations, resulting in a "rougher" looking curve;

(b) the introduction of periodicities at low frequencies that are not present in the original series;

(c) the introduction of phase shifts at certain frequencies;

(d) polarity reversals that produce "peaks" in the filtered series where "valleys" exist in the original series (essentially a 180^{o} phase shift).

Some of these problems are a direct result of the oscillations of the response, or transfer, function above and below zero at periods greater than the length of the moving average. Some, as a result of little robust behavior, being very sensitive to outliers in the series.

Many of these problems can be avoided by constructing filters with weights proportional to the binomial coefficients or, alternatively, to the ordinates of the normal probability distribution. Ideally a filter should be designed to pass all information at the desired frequencies, no information at other frequencies and should have a short span in order to minimize loss of data at the ends of the series. Furthermore, no phase shift should occur at any frequency (Howarth and Rogers, 1922). Based on the above the binomial/Gaussian low-pass filter approach for smoothing the series is considered preferable for this study.

Some of the well known applications of low-pass filter method to rainfall and temperature data are due to Partharathy and Dhar (1974), Giles and Flocas (1984), Barring and Mattsson (1992) and Howarth and Rogers (1992) among others. Low-pass filter method was recently applied to Nigerian rainfall

by Olaniran (1989), Anyadike (1993) and Oladipo (1993).

In this study the drought series (BMDI) for all the stations were subjected to low-pass filter method in order to understand the nature of the trend. The weights used were nine ordinates of the Gaussian probability curve: 0.01,0.05, 0.12, 0.20, 0.24, 0.20, 0.12, 0.05 and 0.01. The response curve of the Gaussian low-pass filter has a response function that is equal to unit at infinite wavelength and it tails off asymptotically to zero with decreasing wavelength. The response is approximately given by the equation (WMO,1966):

$$R(f) = \exp(-2\pi^2\sigma_g^2 f^2) \qquad (6.1)$$

where σ_g is the appropriate standard deviation, i.e.$6\sigma_g$=10 years, or σ_g = 1.67 years.

Smoothed curves of the BMDI series for some selected stations are presented in Figures 6.1 to 6.4 and discussed in Section 6.4.

6.3 Spectral Methods

Spectral analysis, essentially a modification of Fourier analysis, is the name given to methods of estimating the spectral density function by a sum of sine and cosine terms (Fourier series representation). The computation of power spectra was systematized and standardized by the development of the Blackman-Tukey methods during the late 1950s. The stimulus for this development and wide acceptance of these methods was the increase of the use of Fourier methods in many fields, particularly in electrical engineering and meteorology and

computational efficiency, since fewer computational operations are required with a truncated autocorrelation function (Otnes and Enochson, 1972; Shickedanz and Bowen, 1977). Later Cooley and Tukey (1965) described a high-speed algorithm for computation of power spectra (Fast Fourier Transform) that has made feasible the finite Discrete Fourier Transform (DFT) analysis of extensive sets of data (see Brighlam, 1974).The efficiency of the FFT made it practical to compute the spectrum directly without the use of lag correlation function via the Fourier coefficients.

Recently much attention has been given to overcoming certain inherent problems in the conventional methods of spectral analysis such as the Fourier transform of an autocovariance function. New algorithms were proposed as a result, namely the Maximum Entropy Method (MEM) (e.g. Burg, 1967, 1968; Ulrich and Bishop, 1975), Maximum Likelihood Method (MLM) (e.g. Lacoss, 1971, 1976), the Variable Record Length (VRL) technique (e.g. Schickedanz and Bowen, 1975), the Non-Integer (NI) method (e.g. Schickedanz and Bowen, 1977) and the G-Spectral (GS) method (e.g. Gray et al., 1978). Extensive review of these methods is presented in Oladipo (1987) and only a brief summary of some of these methods is presented below.

6.3.1 Fast Fourier Transform and Periodogram Analysis

Traditionally two conventional methods for estimating variance spectral density function were widely used. One method first estimates the autocorrelation function of the data,

multiplies the estimate by the taper and obtains an estimate of the power spectral density using Fourier transforms (Blackman and Tukey, 1959). The second power spectral density estimate, based on the direct approach via the fast Fourier transform (FFT) operation on the data, is the one typically referred to as the periodogram (Jones, 1965).

The Blackman and Tukey technique first estimates a finite number of discrete autocorrelation function values, or lags, for a finite data sequence. The DFT of the estimated autocorrelation lag estimates is then computed to obtain a smoothed power-spectrum estimate as follows: given $q = 2^n$ data points $(x_o, x_i \ldots x_{q-1})$, the discrete Fourier transform (DFT) of the signal x_k is given by

$$F_j = \sum_{k=0}^{q-1} x_k \exp(-2\pi ijk/q), \quad [j = 0, 1, \ldots, q-1] \qquad (6.2)$$

where x_k is the set of equispaced time series observations for $k = 0 \ldots q-1$, F_j is the complex value Fourier coefficient, $i = \sqrt{-1}$, and j is the frequency index (i.e. $w = 2\pi j/q$). FFT is an algorithm that reduces the number of computations required for the DFT from q^2 to $q \log_2 q$.

The periodogram method uses the square magnitude of the discrete Fourier transform (DFT) of the "window" data sequence as a power spectral estimate. For example, for an equispaced time series observations $(x_n; n = 0, 1, \ldots, N-1)$, its power density as a function of frequency (P(f)) defined at the interval $-1/(2\Delta t) \leq f \leq 1/(2\Delta t)$ is given by Gottmann (1981) as:

$$P(f) = \frac{1}{N\Delta t}\left|\Delta t \sum_{n=0}^{N-1} x_n \exp\left(-i2\pi fn\Delta t\right)\right|^2$$

$$= \frac{1}{N\Delta t}\left|X(f)\right|^2 \qquad (6.3)$$

$$\text{for } n = 0, 1, \ldots, (N-1)/2$$

where Δt is the sampling interval, N is the number of observations, i is $\sqrt{-1}$, and X(f) is the DFT which is computed by the fast Fourier transform (FFT) algorithm of Robinson (1967).

The two methods produce similar results and differ only in their computational approach. There are, however some limitations in these conventional methods of spectral analysis. Both methods have poor frequency resolution for short records and for wave length of 5 to 25 years (Schickedanz and Bowen, 1977). Also, these techniques make rather unrealistic assumptions about the extension of the data beyond points in the available data sample. In the Blackman and Tukey method a zero extension is assumed leading to leakage and reduced resolution, and the technique may sometimes produce negative power at some frequencies if the data samples are inadequate (Chen and Stegen, 1974; Oladipo, 1987). The periodogram assumes a periodic extension of the data, thus inserting periodicities in the spectral estimate that do not exist in the data.

Furthermore, both the periodogram and Blackman and Tukey methods use spectral window functions that are generally independent of the data and the choice of which is highly subjective and sometimes arbitrary. The need to increase spectral resolution often lead to reduced statistical stability

of spectral estimates and these have to be traded against each other whenever the available data are limited (Lacoss, 1971).

6.3.2 Maximum Entropy Spectral Analysis

A radically different method for the estimation of the power spectrum, designed to remove the drawbacks in the use of autocorrelation and periodogram, was independently proposed by Burg (1968) and Parzen (1969). The maximum entropy spectral analysis (MESA) represents a successful attempt for estimating a high resolution spectrum with supressed sidelobes through minimal modification of the original data series making greater use of its information content. The idea is to choose the spectrum that corresponds to the most random or the most unpredictable time series whose covariance functions are consistent with the measured data.

MESA is based on the principle of entropy maximization. The spectral distribution is determined by extrapolating the known lag correlations to the finite lag in such a way that the time series characterized by the extrapolated autocorrelation function has its entropy maximized (Oladipo, 1988). The only assumption in MESA is that of a Gaussian zero mean process.

Since its development independently by Burg (1967) and Parzen (1969), detailed arguments about MESA from the standpoint of theory and practice are given in Ulrich (1972), Ulrich and Bishop (1975), Kaveh and Cooper (1976), Kay and Marple (1981), Padmanabhan and Rao (1986) and Oladipo (1988) among others.

A summary of MESA spectral estimate is given below. Let

$$x_i = x(t_i): \quad 1 = 1,2,...N, \quad t_{i-1} = \Delta t \tag{6.4}$$

be a discrete equal spaced time series of zero mean which is a segment of an infinite stationary series. For such a process, the spectral density with maximum entropy, corresponding to the autocorrelation function values for the M + 1 lags ($M < N$), is given (Kirk et al., 1979) by

$$S(f) = \frac{P_{M+1} / W}{\left| 1 + \sum_{k=1}^{M} a_{M,k} \exp(-i2\pi f k \Delta t) \right|^2} \tag{6.5}$$

where $S(f)$ is the spectral estimate at frequency f, $a_{M,k}$ are a set of prediction error filter coefficients (a_k) of order M, P_{M+1} is the mean output power of the filtered series, Δt is the sampling interval which is equal to unity here and thus W, the Nyquist interval $\left[-1/(2\Delta t) \leq W \leq 1/(2\Delta t)\right]$ is also equal to unity.

MESA has been found to have extremely high resolution and was successfully employed by many (e.g., Currie, 1981; 1984a,b; 1987a,b; Padmanabhan and Rao, 1986; Oladipo, 1988 to mention a few) to search for periodicities in hydrologic and climatological time series. However, the Maximum Entropy Spectral Analysis posess certain problems. For example, the variance of MESA estimates of power density is not properly understood and there is still no known way to obtain variances on the spectral estimates. In addition, MESA does not yet have a consistent statistical test for the

evaluation of significant peaks, and the few confidence limit procedures proposed are complicated and rather computationally expensive (e.g., Baggeror, 1976a,b; Reid, 1979; Oladipo, 1988).

6.3.3 The Non-Integer (Non-Harmonic) Method of Spectral Analysis

In an attempt to solve the problems associated with the Blackman and Tukey and periodogram methods of computation of climatological power spectra, Schickedanz and Bowen (1975) first proposed a technique using the concept of variable record length (VRL). This method increases the resolution of spectra peaks by computing multiple sets of orthogonal Fourier coefficients with terms lying between the original harmonics. The major advantage of the VRL technique is the greatly improved resolution of wavelengths in the 10-25 year region, which is often a critical region of interest for climatologists and is not detectable by the usual techniques of spectral estimation.

However, there are some problems associated with the VRL technique. One such problem is the stopping criteria for the truncation and augmentation processes, making it unsuitable for a short data series. Another problem is that of obtaining an exact test of significance for the spectral estimates. Lastly, the spectral estimates computed from this method can not sum to the variance of the original series.

Later, however, Schickedanz and Bowen (1977) attempted to eliminate the problems associated with the VRL method.

This led to the development of a non-integer (NI), or non-harmonic spectral technique for estimating spectra at frequencies which are not integer multiples of 1/n using multiple regression approach. The NI method increases the resolution by modifying the usual Fourier equations so as to permit the computation of non-orthogonal (non-independent) coefficients using non-integer values.

Frequencies at which the spectra is usually estimated are directly connected to the number of harmonics used. Periods of harmonics vary as P/1, P/2, ... P/(P/2), with corresponding frequencies of 1/P, 2/P, ...(P/2)/P, where P is the total length of the period spanned by the data. In this respect considerable information may be missed in the frequencies lying between the above harmonic frequencies. If non-integer values are used for the harmonic number, it is possible to obtain as many sine and cosine functions as are needed to adequately sample the spectra. The large amount of resolution, therefore, leads to the detection of more hidden peaks in the spectra of time series than the conventional method of DFT (Schickedanz and Bowen, 1977).

Furthermore, in NI method the sine and cosine cycles will no longer terminate at the end points of the record (as it is in DFT), so the analysis will no longer be "forced" i.e., the sine and cosine components will not be limited to frequencies that are whole number multiples of the fundamental frequency.

Unlike the periodigram, the NI method does not require any smoothing of raw estimates of the spectrum since the

regression used in estimating spectra smooths them considerably. Moreover, the statistical significance of the spectral peaks can be tested from readily available statistical tables. The NI method of spectral analysis is aimed at resolving frequencies in the range of 5 to 25 year periods, common in many climatological time series.

This study, therefore, uses the NI method of spectral analysis in order to detect periodic or quasi-periodic components in the Bhalme-Mooley drought index series for Nigeria. Detailed treatment of the subject is given in Schickedanz and Bowen (1977) and only a basic description of the method is presented here.

In the DFT method of power spectral estimation, the transformation from the time to the frequency domain is made by first obtaining the Fourier sine and cosine coefficients. Let the variance (x_t) of the BMDI series be given by the following sine and cosine function series.

$$x_t = \bar{x} + \sum_{i=1}^{N/2}\left[a_i \sin(2\pi it/P) + b_i \cos(2\pi it/P)\right] \qquad (6.6)$$

When the sines and cosines corresponding to a given harmonics is combined, Eq. 6.6 is

$$x_t = \bar{x} + \sum_{i=1}^{N/2} C_i \cos\left[2\pi i(t - t_i)/P\right] \qquad (6.7)$$

where P is the length of the observation period, x_t are a sequential series taken at times t = 1,2,...N, $\bar{x}$ is the average of the series, i is an integer from 1 to N/2,

C_i's are the amplitude of N/2 cosine waves, a_i and b_i are the Fourier sine and cosine coefficients given as:

$$a_i = (2/N)\sum_{t=1}^{N} x_t \sin(2\pi it/P),\ i = 1,2,\ldots,\ N/2 \qquad (6.8)$$

and

$$b_i = (2/N)\sum_{t=1}^{N} x_t \cos(2\pi it/P),\ i = 1,2,\ldots,\ N/2 - 1 \qquad (6.9)$$

t_i is the time at which the i^{th} harmonic has maximum and is given as:

$$t_i = \phi P/(2\pi i) \qquad (6.9)$$

where ϕ, the phase angle is arctan (a_i/b_i). The value of a_i for the last harmonic is zero and that of b is

$$b\left[N/2\right] = \frac{1}{N}\sum_{t=1}^{N} x_t \cos(2\pi it/P) \qquad (6.10)$$

In Eq. (6.10), $\left[N/2\right]$ is the integer part of N/2. The constant terms $\bar{x}$, C_i and t_i are to be determined by the method of least squares.

Non-integer method then consists of evaluating the Fourier sine and cosine coefficients for non-integer values by performing a multiple regression analysis of the time series under study on the sine and cosine waveforms for any given i such that:

$$\hat{x}_t = a + (bs_i)(XS_{it}) + (bc_i)(XC_{it}),\quad t=1,2,\ldots,N \qquad (6.11)$$

where a is the intercept, bs_i and bc_i are the partial regression coefficients, and XS_{it} and XC_{it} are the sine and

cosine functions given respectively as:

$$XS_{it} = \sin (2\pi it/P) \tag{6.12}$$

$$XC_{it} = \cos (2\pi it/P) \tag{6.13}$$

and are not limited to frequencies that are whole-number multiples of the fundamental frequency. The multiple correlation coefficient R_i and correlation coefficient r_i between XS_{it} and XC_{it} are then determined for any given i, with R_i^2 (coefficient of determination) being the fractional amount of variance explained by sine and cosine waveforms. The normalized spectral density is then given as:

$$R_i = \sum_{t=1}^{N} (x_t - \hat{x}_t)^2 / \sum_{t=1}^{N} x_t^2 \tag{6.14}$$

If i is an integer value, R_i^2 is identical to the usual normalized spectral density.

R_i and R_i^2 were tested for significance by the F test which can be expressed as:

$$F_i = \frac{R_i^2 / k}{(1 - R_i^2) / \nu} \tag{6.15}$$

where k is equal to the number of independent variables and ν (= N-k-1) is the number of degrees of freedom of the error mean square. Rearranging Eq. 6.15 gives:

$$R_i^2 = kF_i/(\nu + kF_i)$$

with the values of F_i at the 0.05 and 0.10 levels of signifi-

cance determined from the tables of Rohlf and Sokal (1969). Distribution of R_i^2 with frequency F_i associated with i^{th} waveform is the variance spectrum. NI spectral estimates, however, were determined at equal wavelengths rather than frequency and wavelengths resolvable to 0.1 year were searched for. Moreover, the estimates of the regression coefficients are of secondary importance to the estimates of R_i^2.

6.4 Results

6.4.1 Trends and Fluctuations by Low-Pass Filter

In the Sahel, as depicted in Fig. 6.1, there is a good correspondence between different stations, with peaks (above average rainfall) and troughs (droughts) coinciding fairly well, but varying slightly in magnitude and time. A clear downward trend in the series is present in all curves starting in early 1920s in the western Sahel (Sokoto and Birnin-Kebbi) and eastern Sahel (Maiduguri) and early 1930s in central Sahel (Katsina, Hadejia and Kano).

This downward trend culminated in widespread droughts of the 1940s described in Chapter 5. Figure 6.1 also shows positive index values for the 1950s with the peaks of that period centered on mid-1950s in all stations except Katsina, Hadejia and Potiskum. In the Katsina and Hadejia series, the period between early 1950s and mid-1960s is oscillatory and the peak index values for Potiskum is centered on the late 1950s and early 1960s.

The second pronounced downward trend started from the

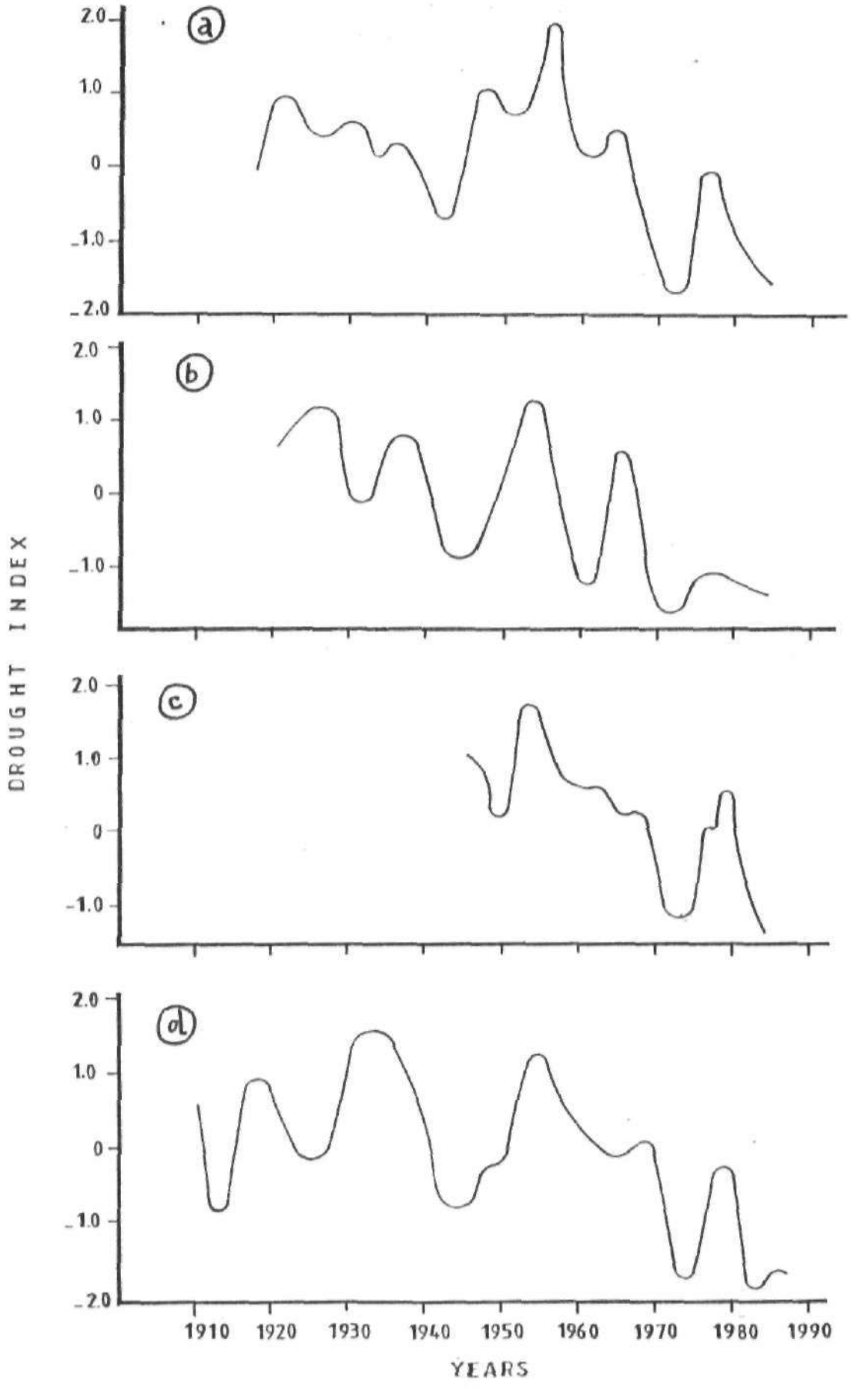


Fig. 6.1. Seasonal values of BMDI series for Sahelian station smoothed by low-pass filter. (a) Sokoto (b) Birnin-Kebbi (c) Gusau (d) Kano

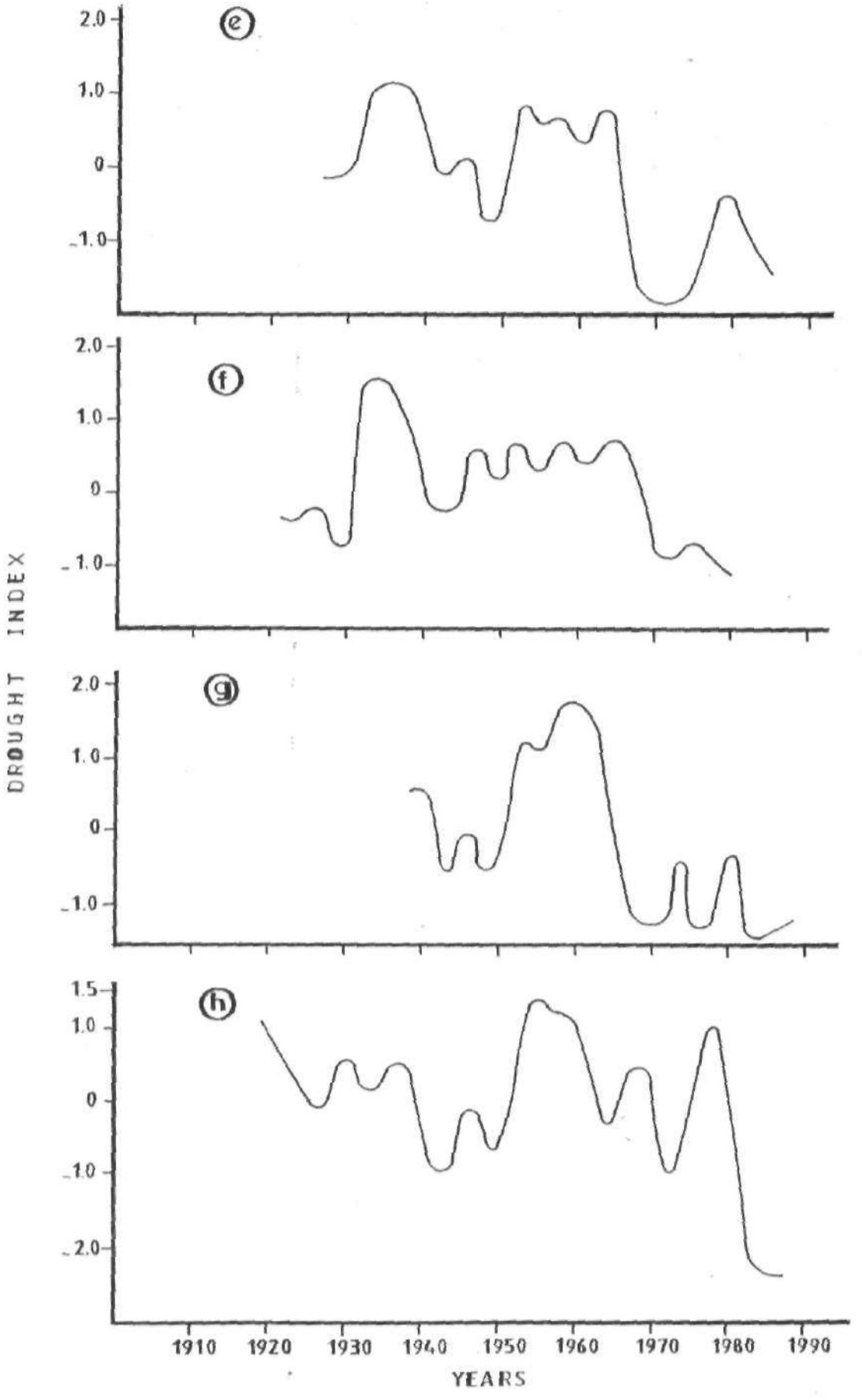


Fig. 6.1. (cont.) (e) Katsina (f) Hadejia
(g) Potiskum (h) Maiduguri

mid-1950s, except for Katsina, Hadejia and Potiskum, where it started in mid-1960s. This downward trend, as depicted by the troughs, extended to the end of the series, to indicate drought persistence especially since the mid-1970s.

Figure 6.2 indicates that, on the whole, the series for the Midland zone are largely oscillatory, although Bauchi, Lafia and Jos (central part of the Midland zone) show weak downward trends starting from the mid-1950s, mid-1960s and late 1960s respectively. The western part of the Midland zone (e.g. Kaduna, Minna and Bida series) experienced a downward trend from about the mid to late 1920s through the 1940s, although the signals are weaker than those of the Sahelian zone. The peaks of the 1960s and the troughs of the 1970s and 1980s coincided fairly well for all the stations in the zone. This suggests that the wet and dry conditions of the 1960s, 1970s and 1980s respectively affected large areas of the Midland zone.

Figure 6.3 shows seasonal BMDI series for some Guinea-Savanna stations. All curves, except Ilorin, are fairly well in phase with each other. Droughts of the late 1930s (described in Chapter 5) are represented by troughs in the long series of Oyo and Ibadan. Unlike the Sahelian and Midland zones that experienced non-drought conditions in the 1950s the Guinea-Savanna zone was drier than normal during this period. This was followed by wet conditions in the 1960s with peaks for nearly all stations centered on early to mid-1960s. The curve for Ilorin is not perfectly in phase with the other stations in the zone and its wet peak is centered

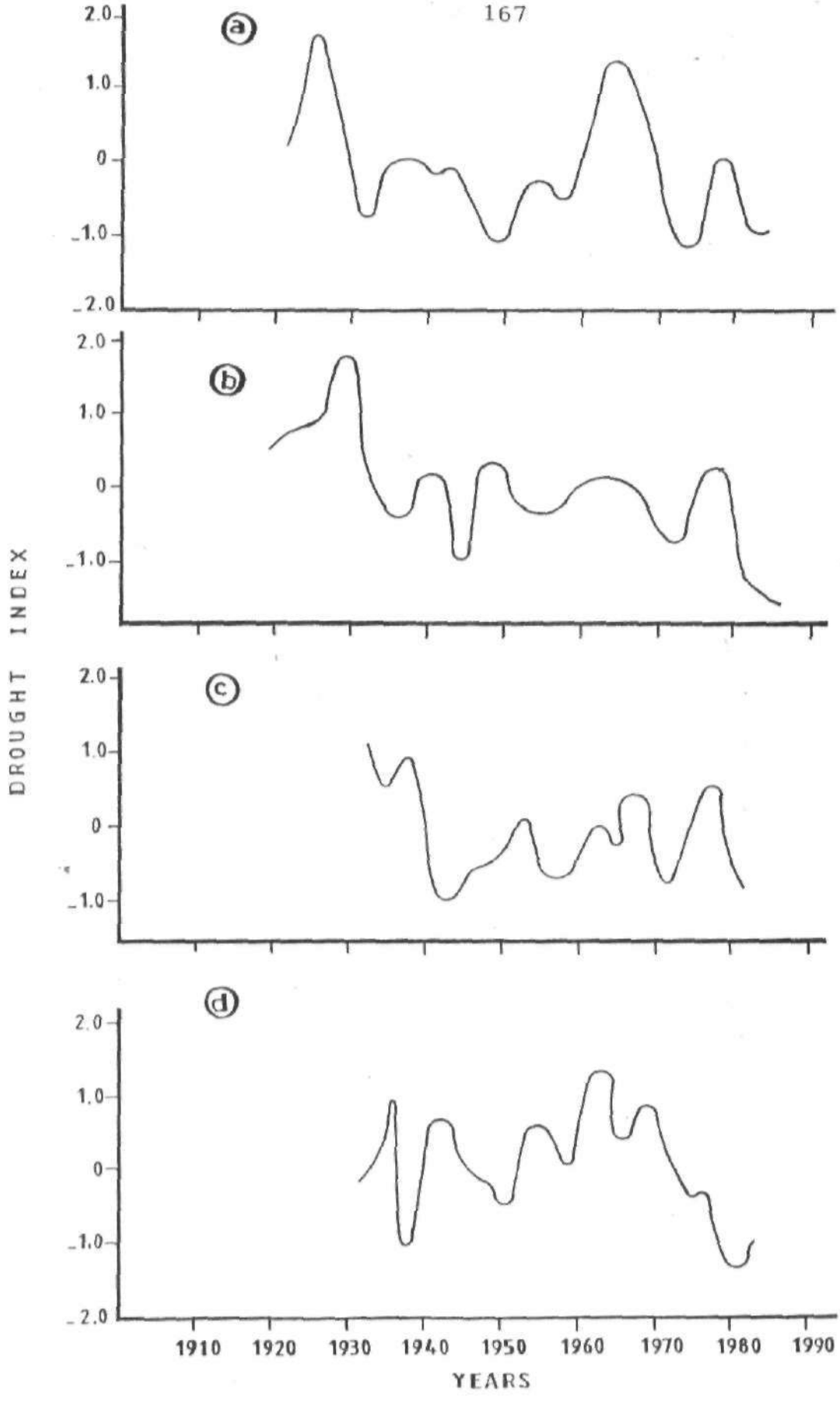


Fig. 6.2. Seasonal values of BMDI series for Midland stations smoothed by low-pass filter.
ⓐ Kaduna ⓑ Minna ⓒ Bida ⓓ Lafia

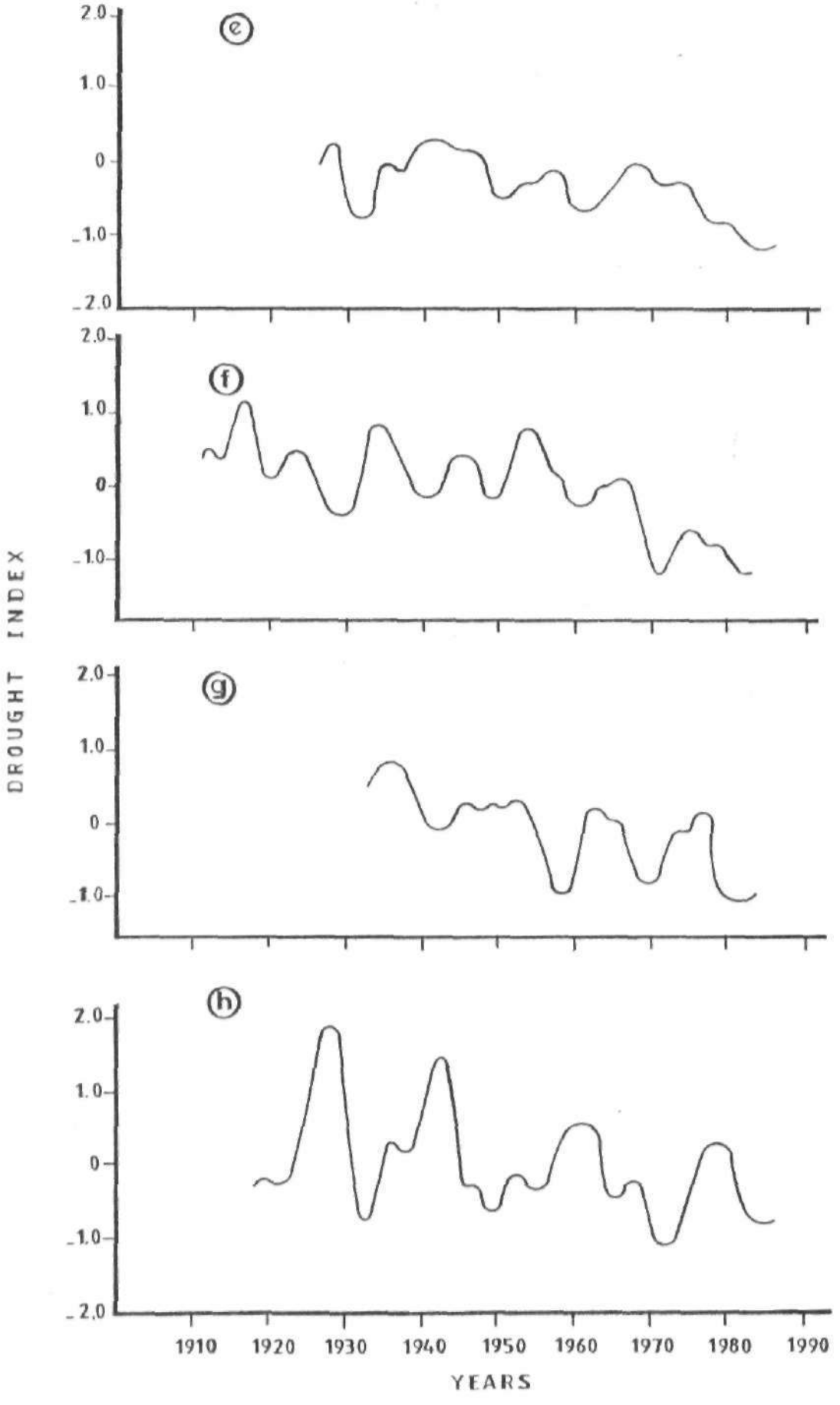


Fig. 6.2. (cont.) (e) Jos (f) Bauchi (g) Pankshin (h) Yola

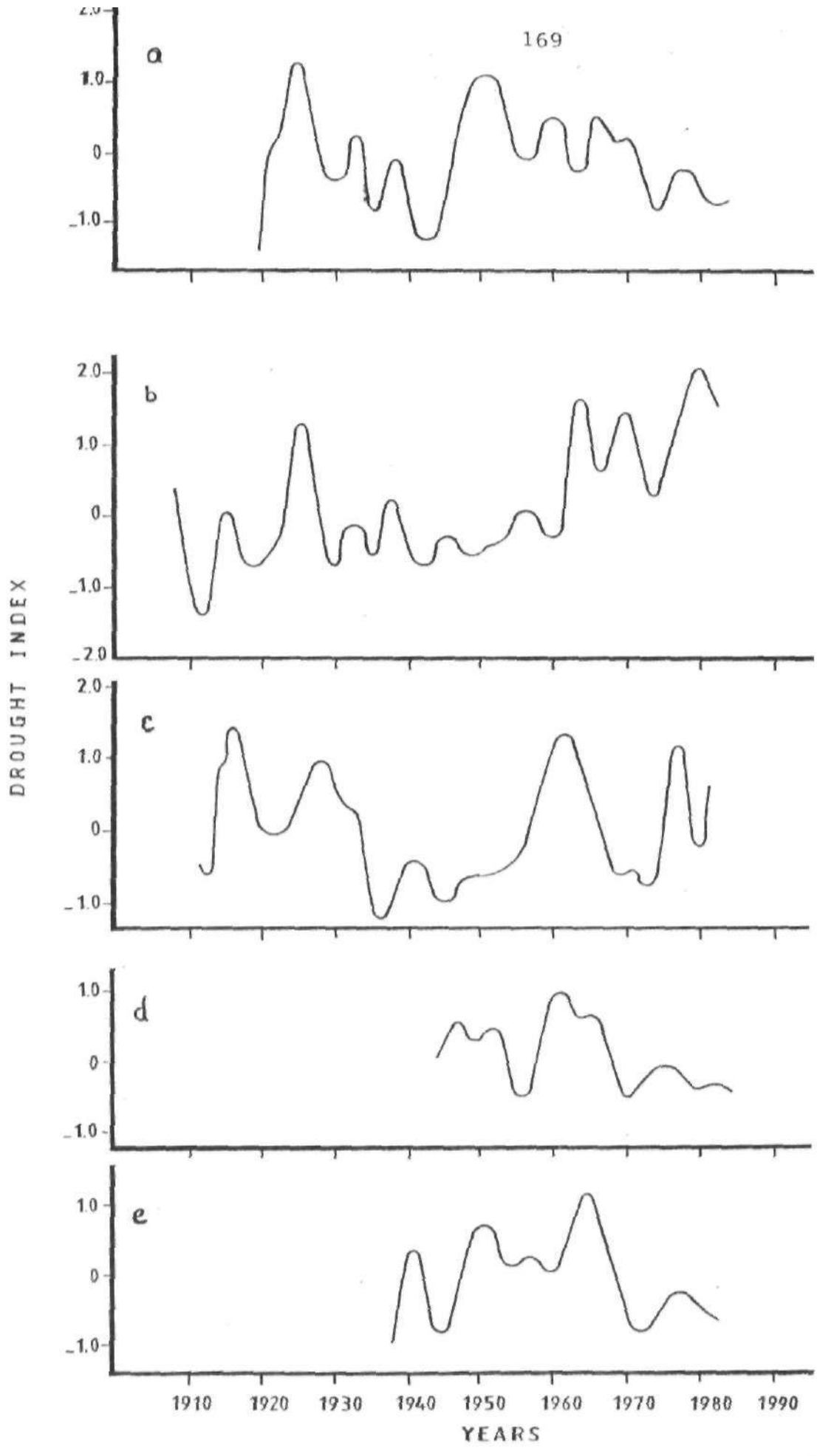


Fig. 6.3. Seasonal values of BMDI series for Guinea-Savanna stations smoothed by low-pass filter. (a) Ilorin (b) Oyo (c) Ibadan (d) Ado-Ekiti (e) Makurdi

on the early 1950s. Some evidence of downward trend is present in Ilorin, Makurdi and Ado-Ekiti series beginning in the mid-1950s for Ilorin and mid-1960s for others.

The Coastal zone series are largely oscillatory (Fig. 6.4) although there are periods of definite upward or downward trends. Deep troughs in nearly all the stations are centered on early to mid 1920s, late 1940s and early 1970s indicating drier than normal conditions during those periods. On the whole, a weak upward trend is noticeable in many stations that are fairly inland (e.g. Ondo, Agbor, Kwale and Enugu) starting between the mid 1950s and late 1960s to the mid 1980s.

In summary, all the Sahelian and the eastern Midland series show downward trends to indicate drought conditions in the past two decades. This trend started earlier in the western Sahel than in the eastern Sahel, pointing to differences in the drought-inducing atmospheric mechanisms between the two sub-regions.

In order to examine the zonal patterns of the drought series, BMDI values for all the stations for each zone were averaged and the results subjected to the same filtering process. The zonal means were based on a minimum of four series for the Sahelian and Guinea-Savanna zones and six series for the Midland and Coastal zones. Figure 6.5 shows the fluctuations in the zonal series.

Generally, a clear downward trend is present in the Sahelian and Midland zones, starting from the mid-1950s in the Sahel and late 1960s in the Midland zone with troughs (drought

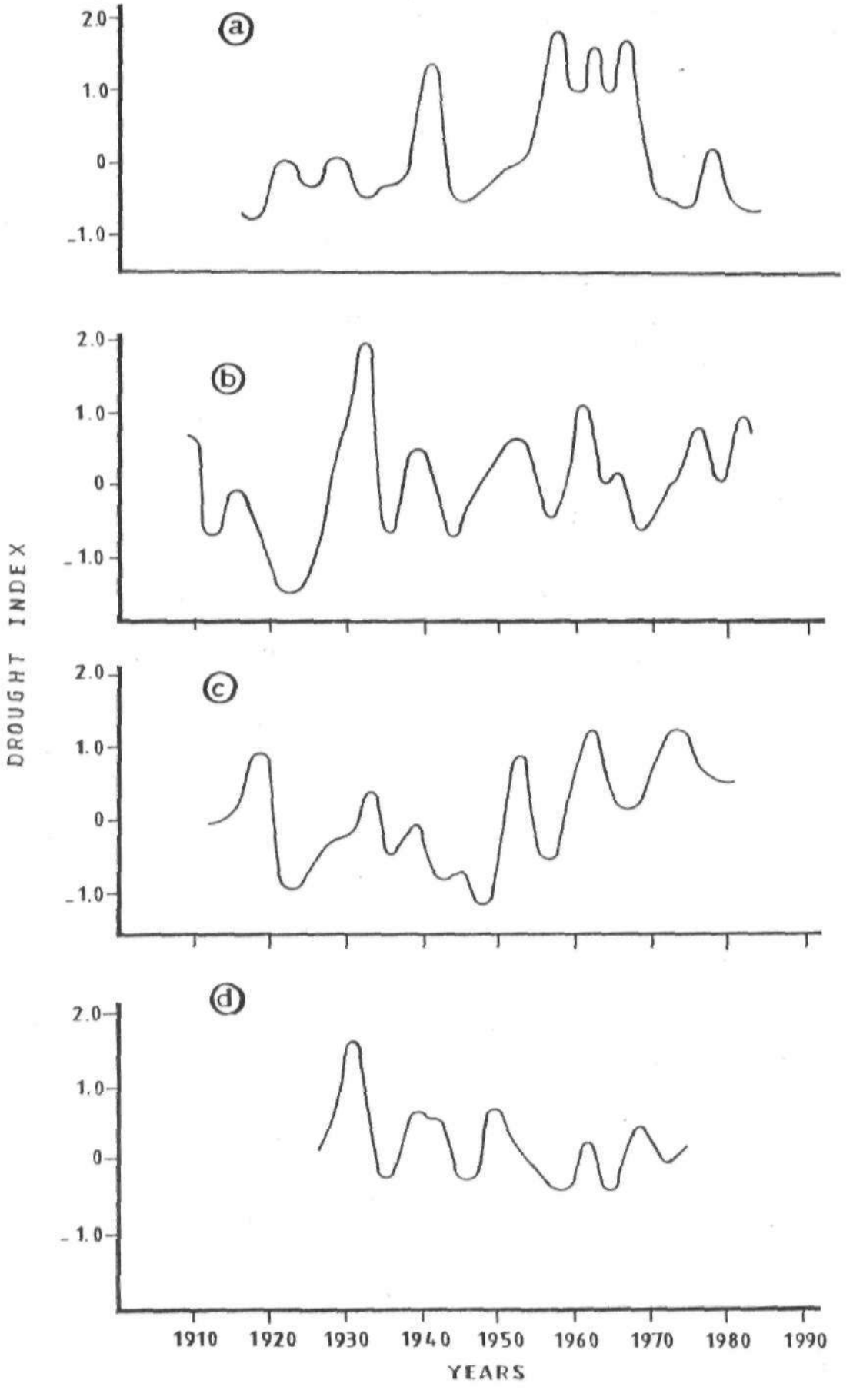


Fig 6.4. Seasonal values of BMDI series for Coastal stations smoothed by low-pass filter.
(a) Lagos (b) Ondo (c) Agbor (d) Kwale

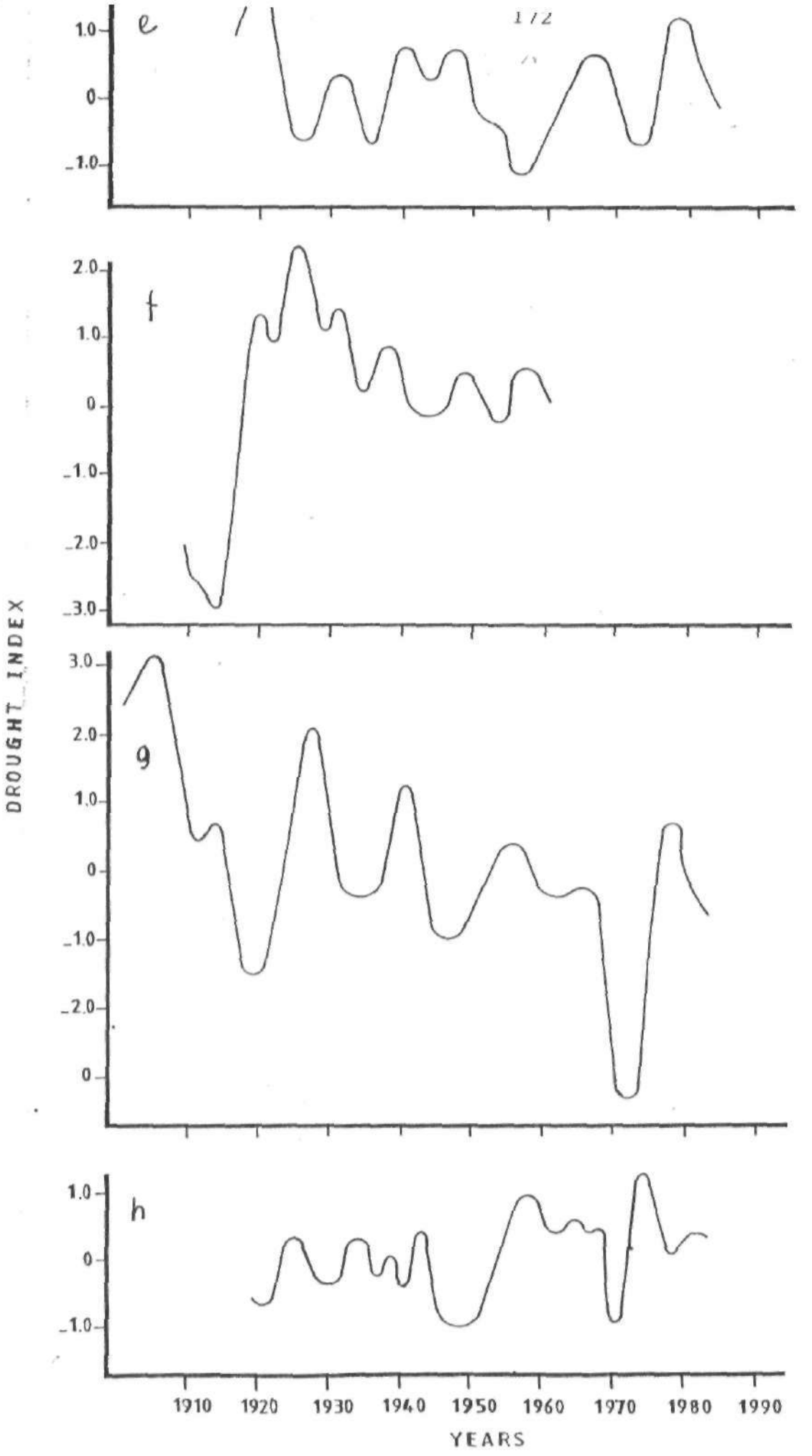


Fig. 6.4. (cont.) ⓔ Enugu ⓕ Afikpo
ⓖ Calabar ⓗ Port-Harcourt

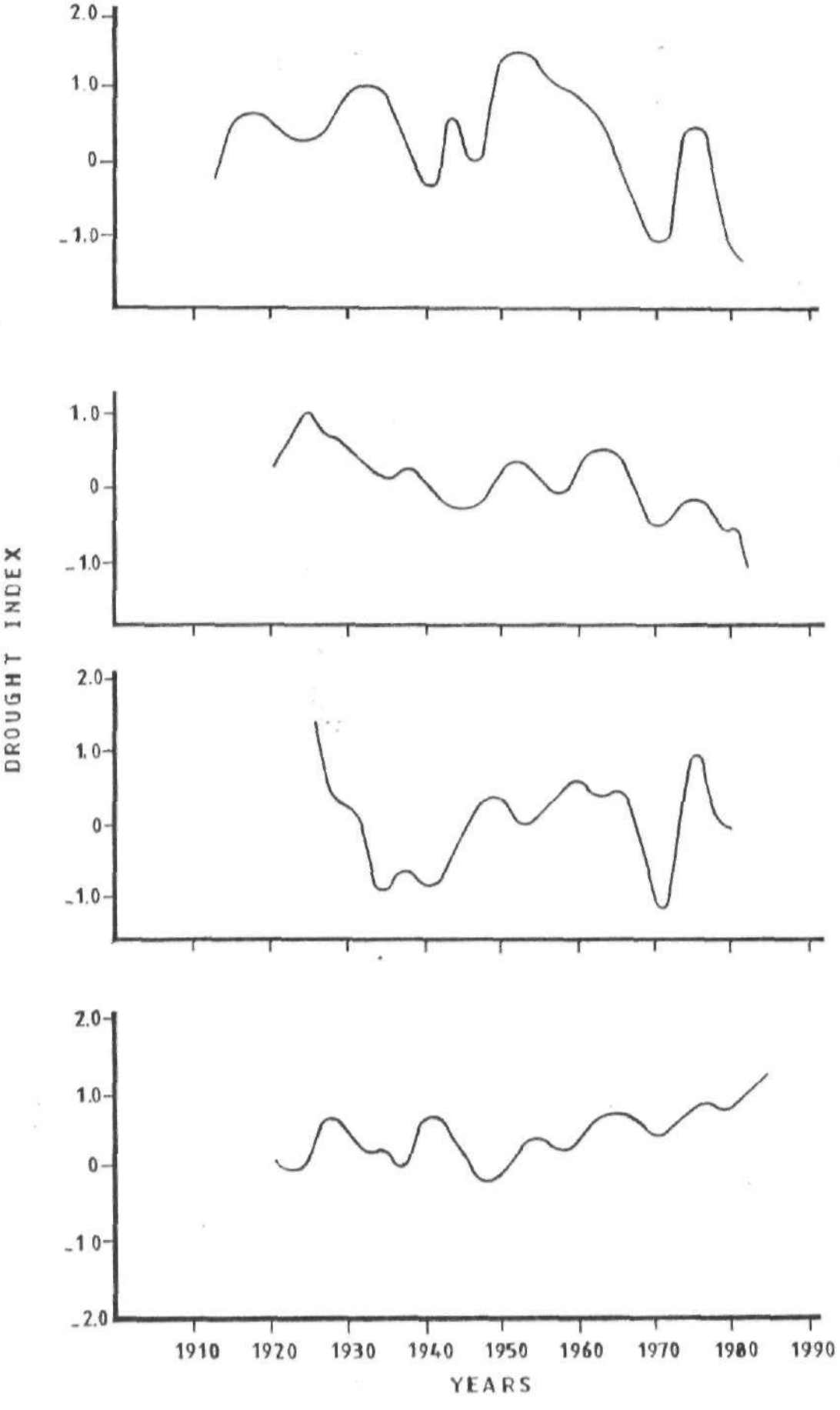


Fig. 6.5. Seasonal values of BMDI series for four geographical regions smoothed by low-pass filter . ⓐ Sahelian zone ⓑ Midland area ⓒ Guinea Savanna zone ⓓ Coastal area

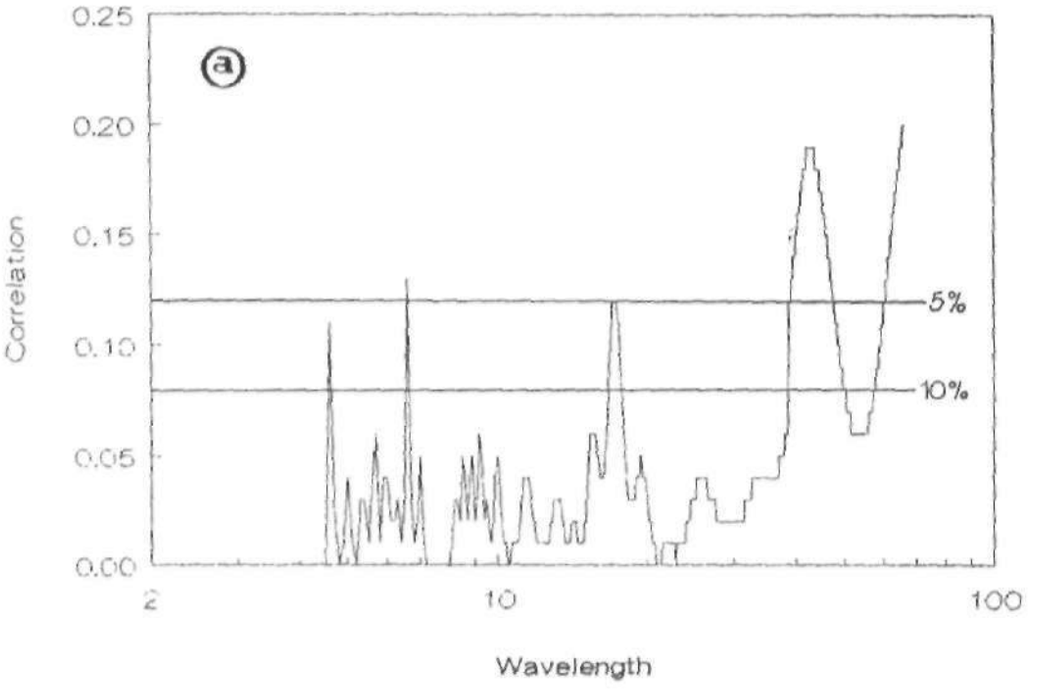
periods) centered on late 1960s-early 1970s. The Guinea-Savanna curve is generally stationary, but with troughs centered on the 1930s and early 1940s to indicate drought conditions. This is similar to what was noted for the Sahelian and Mid-land zones. On the other hand, the Coastal zone shows a slight upward trend, starting in early 1950s to the end of the series to indicate a trend towards a wetter than normal moisture condition in recent years.

6.4.2 Periodicities Due to Non-Integer Method of Spectral Analysis

All the fourty-one series used in this study were subjected to the Non-Integer method. The maximum resolvable period was 66 years, with some stations record lengths being less than that. Figures 6.6 and Appendix C show the normalized spectra of the seasonal drought index series for some selected stations, while Appendix D presents significant periodicities (years) at 5 per cent and 10 per cent significance levels for all the stations.

All graphs were examined for periodic variations. The strong peaks at 2.0-year frequency values were rejected because they represent the Nyquist frequency, with energy being folded back into the spectrum. As the series contain as much as 641 data points, correlation values as low as .08 are statistically significant at 10 per cent level and useful for the description of recurrent spatial spectral patterns of drought. Periods shorter than two years are not resolvable with seasonal data, while periods greater than one fourth , or fifth of the record length might not be very

PLOT OF SOKOTO STATION



PLOT OF JOS STATION

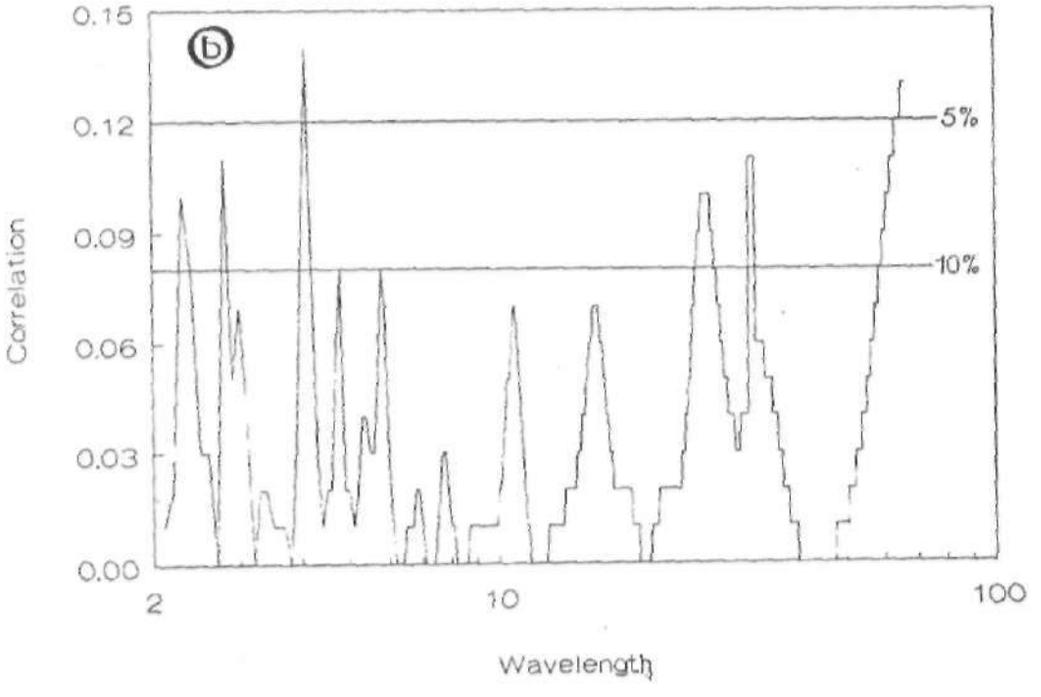
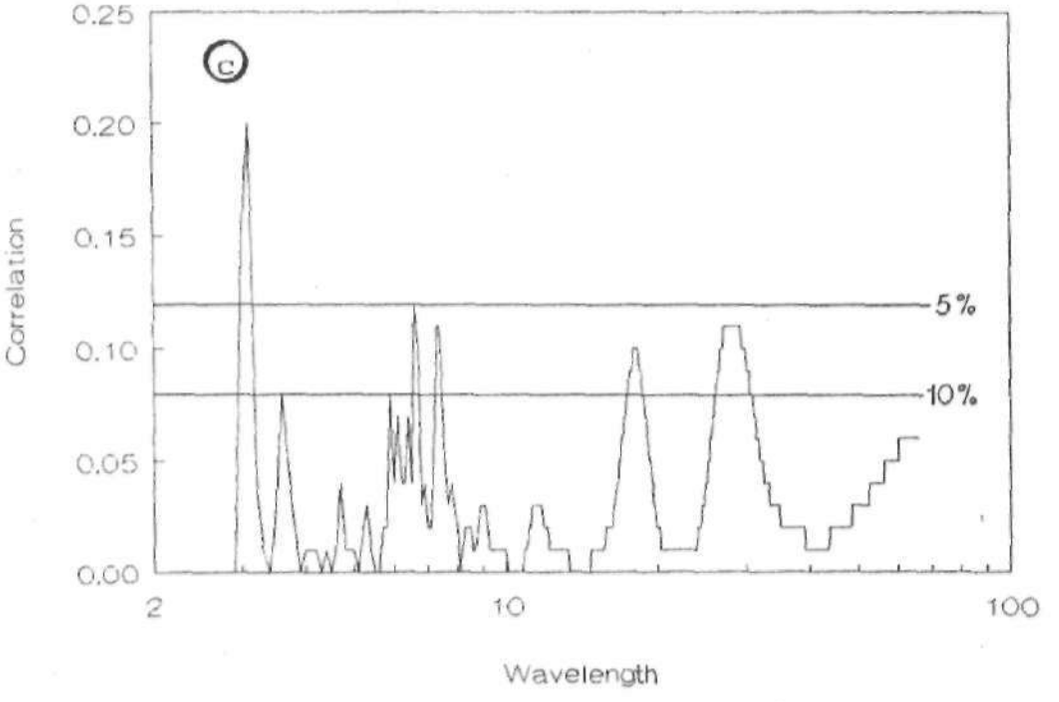


FIGURE 6.6 NON-INTEGER SPECTRA OF DROUGHT FOR a)SOKOTO (SAHELIAN ZONE), and b)JOS (MIDLAND AREA).

PLOT OF IBADAN STATION



PLOT OF LAGOS STATION

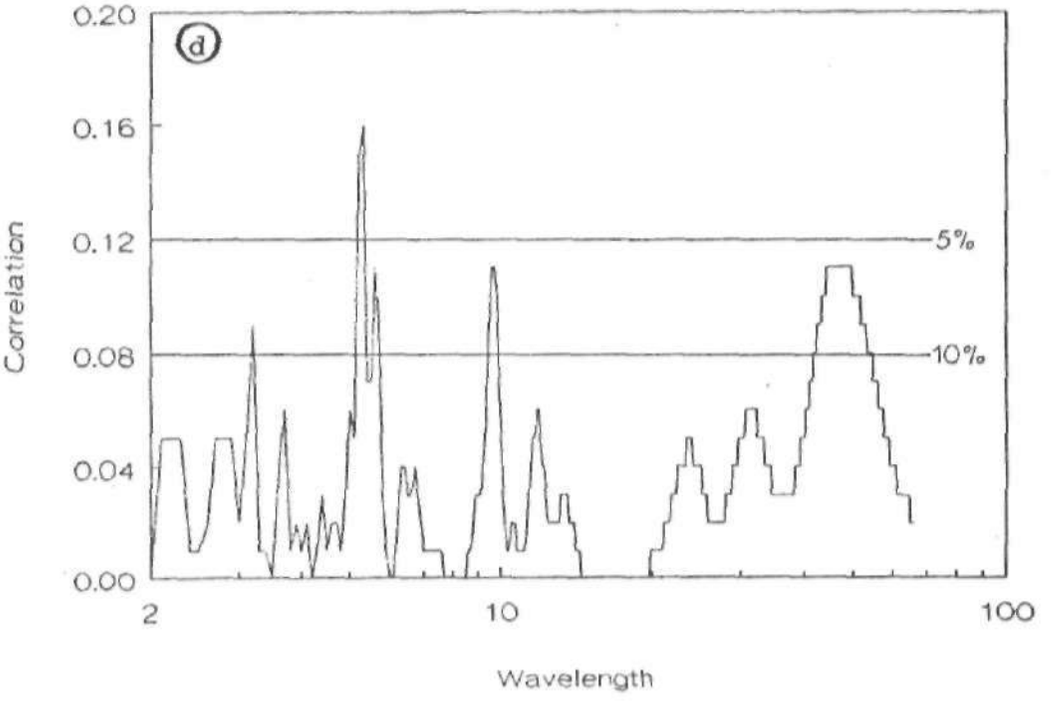


FIGURE 6.6 (cont.) for c) IBADAN (GUINEA-SAVANNA)
d) LAGOS (COASTAL AREA)

reliable, although theoretically the NI technique has power to detect long-term trends or cycles (Schickedanz and Bowen, 1977). Thus, only periods between 2.1 and 13.9 years are examined and discussed.

Figure 6.7 (a,b,c,d) indicates the number of stations with significant periodicities in various frequency bands, while Figure 6.8 shows the spatial distribution of stations with significant periodicities. An examination of Figure 6.6 reveals some significant periodicities, that came mainly from broad frequency bands of 2.1 to 13.9 years and are now discussed.

The 2.1-2.9 year frequency bands. Significant peaks at this approximately quasi-biennial oscillation are particularly characteristic for stations in the Midland zone, less in the Guinea-Savanna and Coastal zones and virtually absent from the Sahel (Figs. 6.7 and 6.8a). They are distributed over a broad geographical region. The quasi-biennial behavior is well known to exist in many climatological phenomena. For example, Bunting et al. (1976), Kraus (1977), Nicholson (1983) and Anyadike (1993) have verified quasi-biennial oscillation for West African and Nigerian rainfall series. The fundamental physical causes of quasi-2-year oscillation is still not very clear, but it is thought to be generally related to ENSO-like quasi-periodicities in Atlantic sea-surface temperatures, and viewed as a pulsation of the general atmospheric circulation (e.g. Farmer and Wigley, 1985; Oladipo, 1987).

The 3.0-3.9 year frequency bands. Most of the stations with quasi-triennial periodicity in their drought index series

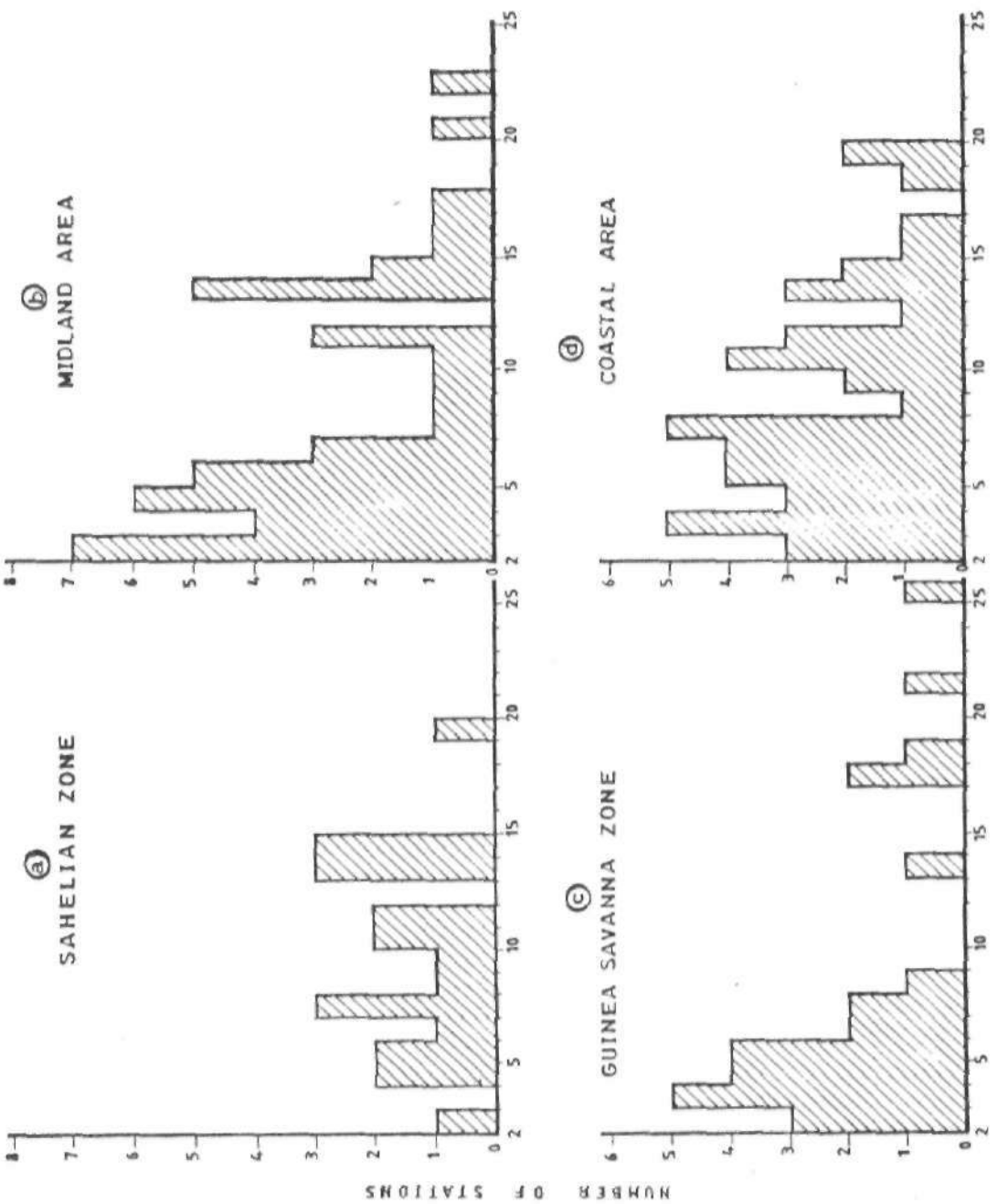


Fig. 6.7 : The number of stations with significant periodicities in various frequency bands for four geographical regions in Nigeria.

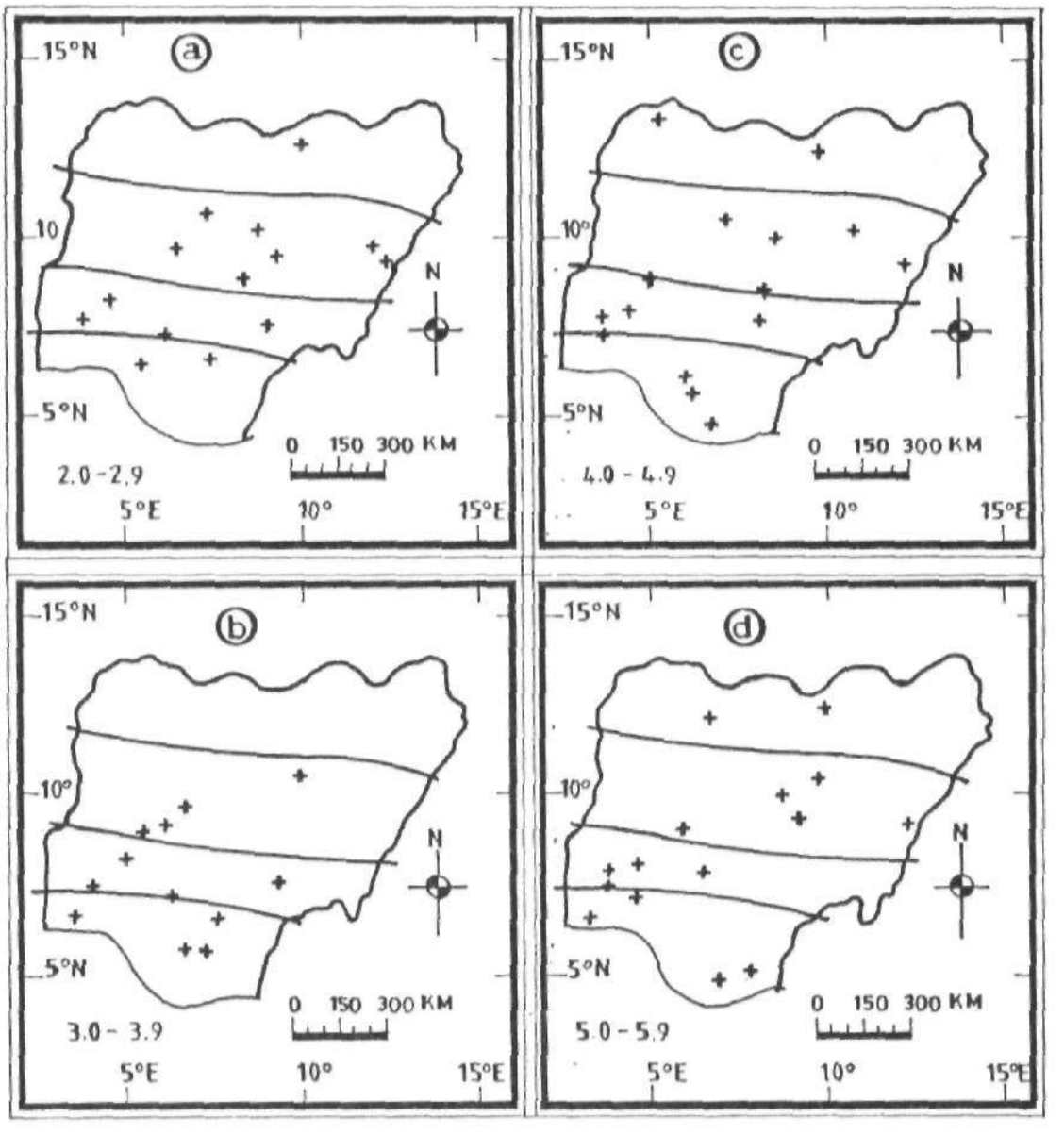


Fig. 6.8 Spatial distribution of stations with significant 2.0-2.9, 3.0-3.9, 4.0-4.9 and 5.0-5.9 year periodicities.

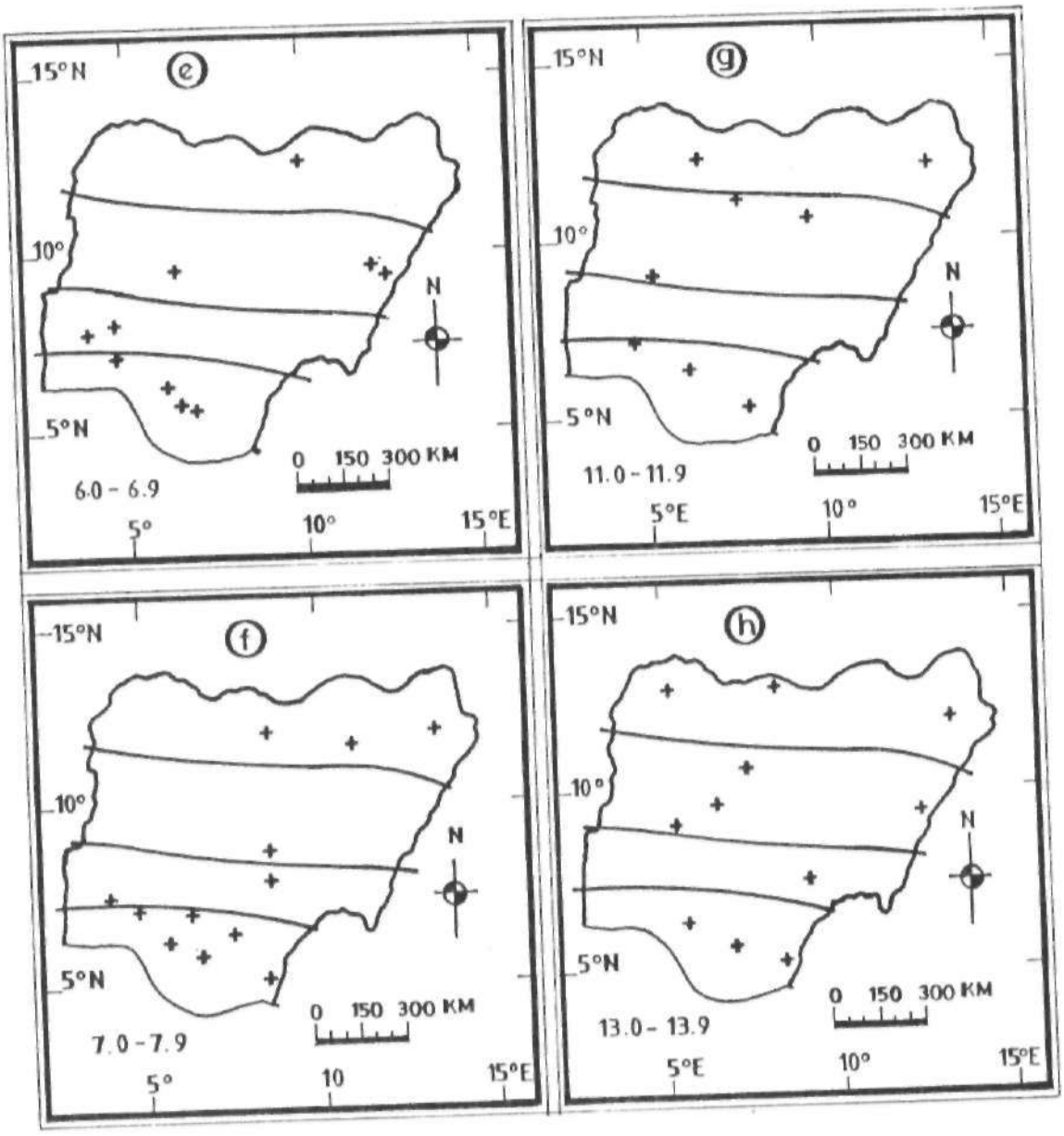


Fig. 6.8 (cont.) for 6.0-6.9 , 7.0-7.9, 11.0-11.9 and 13.0-13.9 year periodicities .

are concentrated in the Coastal zone. They are less prominent in the Guinea-Savanna and Midland zones and are absolutely absent from the Sahel (Figs. 6.7 and 6.8b). The geographical distribution of these significant peaks are shown as clustered in the Coastal and south-western part of the country. The exact significance of the quasi-3-year oscillation is not known, although similar peaks were reported by Bunting et al. (1976) and Nicholson (1983) for the Sahelian and West African rainfall. The quasi-triennial periodicity has also been verified for the Indian drought indices (Bhalme and Mooley, 1983) and it is suggested to be close to first of the tree modal peaks of such quasi-periodic climatic phenomenon as Southern Oscillation (SO).

The quasi-4-year (4.0-4.9) frequency bands. Stations with significant quasi-4-year oscillation are present in all zones and are distributed in broad gegraphical regions(Figs. 6.7 and 6.8c). These frequency bands are still very close to the quasi-periodic climatic phenomenon such as ENSO and they have been noted for the West African rainfall series by Bunting et al. (1976), the North African series by Kraus (1977) and South African rainfall series by Tyson and Dyer (1978).

The 5.0-5.9 year frequency band. Quasi-periodic variations in this range are most common in the central part of the Midland zone, the western part of the Guinea-Savanna and the Coastal zones. As can be seen in the Fig. 6.8d there is a cluster of stations with significant periodicities in this frequency band in the south western part of the region. Again, this oscillation is not prominent in the Sahel. The approxi-

mately 5-year periodicity in climatological time series has received little attention, although their occurrence has been reported by many researchers (e.g. Rhode and Virgi, 1976; Fleer, 1981; Nicholson, 1983) and have been investigated as possible harmonic subperiod of the 11-year sunspot cycle by Fleer (1982). This oscillation is also close to the second modal peak (5.5 year) of the Southern Oscillation (SO) phenomenon. Because Fleer's (1982) analysis did not indicate any linear relationship between rainfall fluctuations and the 11-year sunspot cycle the occurrence of significant 5.0-5.9 year oscillation may indicate some non-linear effect in the drought-inducing atmospheric mechanism.

The 6.0-6.9 year frequency band. Ten stations indicate significant oscillations in the 6.0-6.9 year range. Most of these are in the Coastal zone and only two stations in Guinea-Savanna zone fall into this category (Figs. 6.7 and 6.8e). Frequencies in this range are not common and have not been verified for the Sahelian and West African rainfall. However, their existence in the South African and North African rainfall series have been noted by Tyson and Dyer (1978) and Vines (1980) and Kraus (1977) respectively.

The 7.0-7.9 year frequency bands. Most stations with significant quasi-7-year oscillations are found in the eastern Sahel and the Coastal zones. This agrees with the findings of Bunting et al. (1976) and Nicholson (1983) who reported similar periodicities with peaks at 7.94 and 7.85 for the Sahelian rainfall series. The exact significance of this oscillation is not very clear as it appears not to be directly rela-

ted to any known major atmospheric mechanism. This oscillation is in the longer frequency band and its persistence may be useful for drought prediction.

The 11.0-11.9 year frequency bands. Significant periodicities in the range are found in the Sahelian, Midland and Coastal zones, but not the Guinea-Savanna zone (Fig.6.8g). Together with significant peaks in quasi-10-year oscillations for about five stations they represent about 29 per cent of all the stations in Nigeria. This oscillation is compatible to the 11-year sunspot cycle and to the third modal peak of the Southern Oscillation (SO). The quest for evidence of solar influence on precipitation series proved very difficult and produced conflicting results. On the one hand, some authors (Bunting et al., 1976; Kraus, 1978; Nicholson, 1983) verified 11.10 and 11.75 year peaks in the Sahelian and West African rainfall series. Other authors (Bhalme and Jadhav, 1983; Currie, 1984a, b; Currie and Hamid, 1986) found evidence of the solar cyclic 11-year induced drought in the Indian and North American drought index series. On the other hand, Oladipo (1987) did not find 11-year solar signals in North American drought index series. Also, spectral analysis of Nigerian rainfall by Olaniran and Sumner (1990), Adejuwon et al. (1990) and Anyadike (1993) failed to detect this oscillation in the series. In our own case, although the quasi-11-year peaks have been identified in the region of study, the insignificant number of stations (29 per cent) and their geographical distribution make it difficult to confirm the existing effect of the popular 11-year sunspot cycle in modulating

drought over the region.

The 13.0-13.9 year frequency bands. Significant peaks in this frequency bands are found in the Sahelian, Midland and Coastal zones. These oscillations are virtually absent from the Guinea-Savanna zone (Fig. 6.7 and 6.8h). Only about 27 per cent of the stations show evidence for the occurrence of these frequency bands. The number is not significant enough to justify the existence of a spatially coherent response of drought to these cycle bands. However, Nicholson (1983) has identified 13.34 year cycle for the Sahelian rainfall.

The number of stations with significant periodicities in various frequency bands is summarized graphically in Figure 6.9. The figure and the Appendix D clearly indicate that periodicities greater than quasi-13-year frequency bands are not significant either in terms of the number of stations, or their areal distribution. Moreover, periodicities greater than one fourth, or fifth of the data lengths are not reliable (Neill, 1980). Hence, they are not discussed.

In summary, the results of the spectral analysis suggest that the drought spectra are dominated by peaks in the 2.1-7.9, 10.0-11.9 and 13.0-13.9 bands. The examination of the geographical distribution of stations with significant periodicities in these bands indicates near total absence of 2.0 to 3.9 from the Sahelian and western Midland zones. Instead the Sahelian zone shows quasi-7 and the quasi-13-year periodicities. Within the Sahel, the western and central part is dominated by peaks in longer range (13.0-14.9-years),

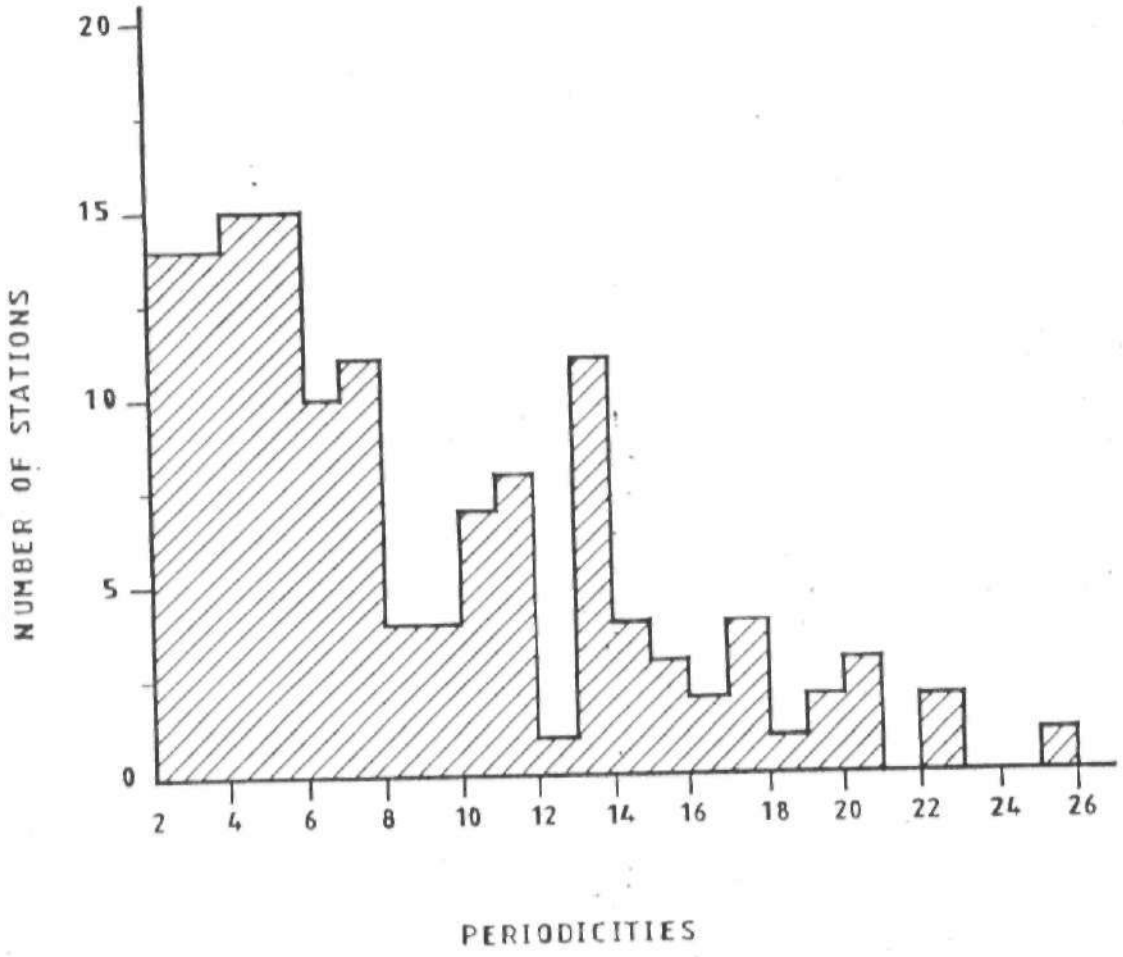


Fig. 6.9 The number of stations with significant periodicities in various frequency bands in Nigeria.

while the eastern part is dominated by shorter (7.0-7.9 and 10.0-13.9) cycles.

The results of the NI method also suggest that the number of cases with significant spectral peaks is not great, varying between 20 per cent to 37 per cent of the stations in the discussed frequency bands. We may then conclude that there is a general incoherence in the temporal behavior of drought occurrence in Nigeria, although certain significant peaks within the 3.0 to 7.9 range appear more frequently in the Coastal and south western parts of the country, than in the Sahel. Because these periods can be short-lived time variabilities, they may be associated with some sort of quasi-periodic oscillations in some atmospheric variables such as the Atlantic sea-surface temperature. In this respect, they may not be useful for prediction purposes. Longer periodicities in the 11.0-13.9 frequency bands may prove useful for prediction. Unfortunately the number of stations with these peaks that are statistically significant is small and they are restricted to areas north of $10^{o}N$ and Coastal zone.

On the whole, the spectral behavior indicates that drought occurrence in the region has a large random component making its prediction difficult.

CHAPTER 7

SUMMARY AND CONCLUSIONS

7.1 Summary

The purpose of this study is to describe the spatial and temporal characteristics of April-October drought occurrence in Nigeria by means of statistical analysis. Fourty-one stations seasonal rainfall data for a long period of years was used for this study. In order to give comprehensive description of April-October precipitation climatology, precipitation series were analysed for homogeneity, randomness and secular variations.

Three meteorological drought indices, depicting the intensity and severity of drought were calculated. The procedure of Bhalme and Mooley (1980) was followed closely to derive a drought index for Nigeria (BMDI). This index was used in subsequent investigations.

Map-pattern correlation procedure which correlates two variables in a manner similar to simple correlation of time series of two or more variables, was used to stratify the time period into moisture anomaly patterns. In order to deliniate spatial coherence of drought, maps of mean drought patterns for each year were generated. The frequency, persistence and spatial coherence of drought was investigated next.

In an attempt to detect trends and fluctuations, the long series of drought index (BMDI) were smoothed by low-pass filter. The Non-Integer method of spectral analysis that is effective in resolving 2-25 year periodicities was applied to the drought index series. The spectra obtained was exa-

mined for significant periodicities in drought occurrences and the spatial coherence of significant peaks in the country.

7.2 Conclusions

The following conclusions from the above investigations can be made:

(i) The April-October precipitation series show high variability over the Sahelian, eastern Midland and western subsectors of the Guinea-Savanna and Coastal zones. This can be attributed to the interannual variabilities in the northward excursion of the ITD, frequency of squall lines and thunderstorms and lack of widespread convergence for the Sahel and subsidence and stabilization of monsoon air for the south western part of the region. The rainfall series for the Sahelian and northern Midland zones have indicated a significant trend towards increasing aridity, especially in the last 30 years. Decadal analysis reveal the 1950s as generally significantly wetter and 1970s and 1980s as generally significantly drier than other periods in the Sahel and northern part of the Midland zone.

(ii) The Bhalme and Mooley Drought Index (BMDI) provides provides a realistic assessment of drought conditions in the country and is considered superior to the van Rooy and Tabony indices.

(iii) Six moisture anomaly patterns, corresponding to four basic modes of moisture anomaly and consistent with the ITD model, are identified by map-pattern correlation method. These patterns show distinct differences between

areas north of 10°N and the rest of the country. The correlation method stratifies about 46 per cent of the years and it is thought to be successful in grouping years with spatially coherent moisture conditions.

(iv) Areas of mild to severe drought covered at least 11 per cent of Nigeria in about 43 per cent of the years.

(v) On an annual basis, there are considerable changes in the patterns, size and position of dryness and wetness in Nigeria. Drought does not cover the entire country and no particular areas were persistently affected until the 1970s and 1980s. Droughts of 1971, 1973 and 1980s show areas north of 10°N to be persistently affected.

(vi) Areas affected by drought have increased in extent since the beginning of the 1970s with a consequent decrease in near normal and wet areas to indicate a trend toward increasing aridity. The frequency of drought increased towards the recent decades in all zones, although it is more pronounced in the Sahelian and the Midland zones. The 1980s were characterised by an exceptionally high frequency of drought conditions with a presence of consecutive years of large-scale droughts (over 40 per cent of the region affected) since 1982, confirming persistence of drought in Nigeria. The agricultural implications of the above is that this region might suffer from crop failure, ecological degradation, desertification and other adverse implications of drought if the recent trend persists.

(vii) A clear downward trend as indicated by low-pass filter analysis is present in the Sahelian and Midland zones starting from the mid 1950s in the Sahel and late 1960s in the Midland zone. On the other hand, low-pass filter results for the Coastal zone revealed a slight upward trend that started in the early 1950s indicating wetter than normal moisture conditions in recent years.
(viii) Significant cycles in the frequency bands between 2.1-13.9 years were identified in the drought series for some stations. The drought spectrum is dominated by peaks with periods of about 2.1-7.9, 11.0-11.9 and 13.0 to 13.9. These significant periodicities are generally spatially sporadic. However, peaks in the 3.1-7.9 frequency range are more evident in the Coastal and south western parts of Nigera, than in the Sahel. Longer frequencies with peaks between 11.0-11.9 and 13.0-13.9 years are evident in the Sahel and the Coastal zone, although the number of stations with significant periodicities is still small. Based on the number of stations with significant quasi-11 year oscillations and their spatial distribution it is suggested that a linear relationship between drought index fluctuations and 11-year sunspot cycle may not be realistic as it is often portrayed in literature. In general, the drought series display a random variation, making its accurate prediction difficult. Nonetheless, 11.0-11.9 and 13.0-13.9 oscillations, if spatially coherent, may prove useful for prediction of drought in the Sahel. This calls for a future exami-

nation of the cyclic pattern of drought in this range with the use of more sophisticated prediction methods.

(ix) From the above discussion we conclude that seasonal (April-October) drought in Nigera is a complex random event, generated by non-linear atmospheric occurrencies.

It is clear from this study that drought as an inevitable part of climate in Nigeria will continue to affect the country, but the extent and time of occurrence will be difficult to predict because of its remarkable changes in size and position from year to year.

A disturbing conclusion from this study is the observed trend towards increasing aridity in recent years, particularly in the Sahel and northern part of the Midland zone. This trend affecting the arid and semi-arid regions of Nigeria, has since the 1970s resulted in devastating droughts. These droughts have been widespread, persistent and capable of producing serious impacts on the environment. These impacts are usually most visible in the state of plant communities, depletion of soils and shallow ground-water resources and general deterioration of the environment.

In the midst of worsening climatic conditions (particularly in the arid and semi-arid zones of Nigeria) the population (human and animal) has continued to grow at a fast rate. The increasing population pressure in these areas has resulted in the overall misuse of the ecosystems of these areas. There are now evidences of ecological degradation in the region due to the grazing pressure, extention of cultivation into marginal areas, vegetation removal and wood cutting.

Human and animal pressure on the fragile ecosystems of the areas north of 10°N became particularly destructive at the time when the extreme climatic variability, manifested by frequent droughts, is affecting the region.

So far, the approach to drought in Nigeria has been inconsistent and uncoordinated, comprising mainly of remedial and ad hoc measures (e.g. provision of subsidized food, water and basic medicare). Even the establishment of afforestation programmes and the construction of boreholes and large-scale dams for irrigation as anti-drought mitigation measures, are not sufficient and satisfactory on many grounds (see Olofin, 1991 and Oladipo, 1992 for details).

The key to a better drought management is in the formulation of a comprehensive, coherent and systematic development strategy aimed at disaster preparedness, drought monitoring, assessment and management. To this end the institutional approach advocated by Oladipo (1993) needs to be examined and evaluated for its appropriateness in addressing the problem of drought management in Nigeria.

In conclusion, this study has shown that drought is a climatic nemesis and an enigmatic phenomenon that requires continuous study. This calls for further investigations in the areas of modelling of drought and the adaptation of appropriate methods for drought management in Nigeria.

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APPENDIX A

SEASONAL MEANS OF BHALME AND MOOLEY (BMDI), VAN ROY (RAI) AND TABONY DROUGHT INDICES FOR SOME SELECTED STATIONS IN NIGERIA

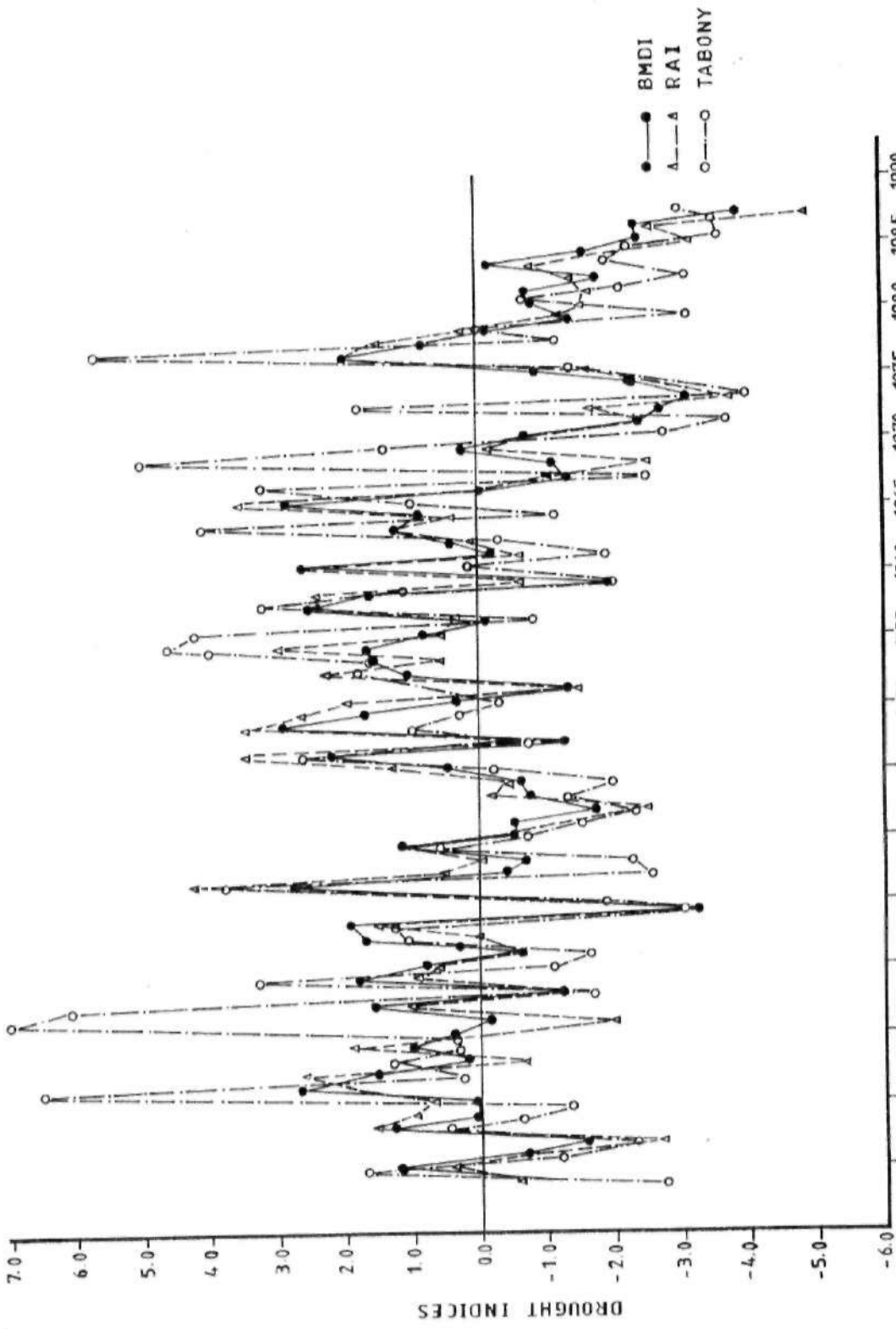


Fig. 1. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Sokoto (SAHELIAN ZONE).

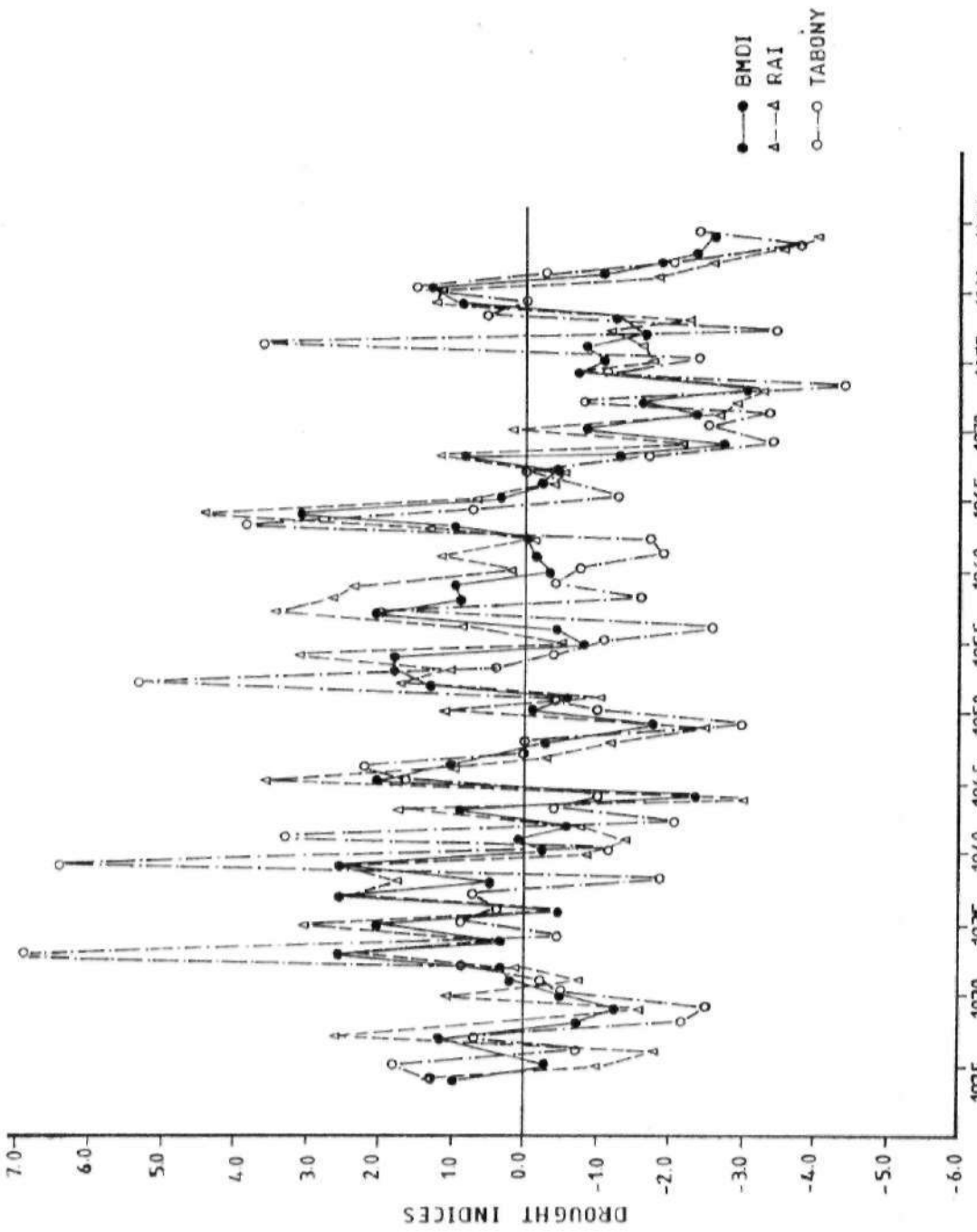


Fig. 2. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Katsina (SAHELIAN ZONE).

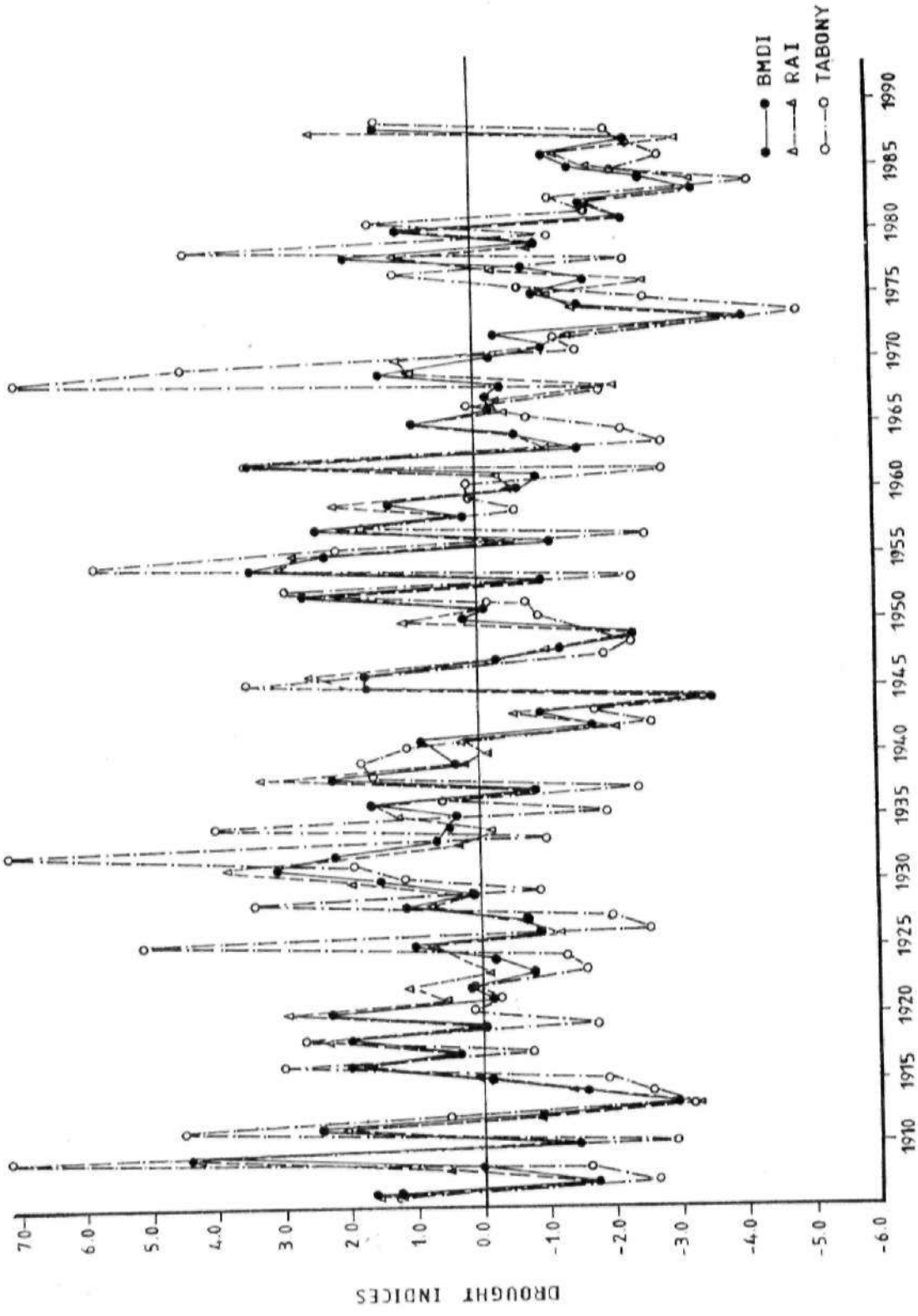


Fig. 3. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Kano (SAHELIAN ZONE).

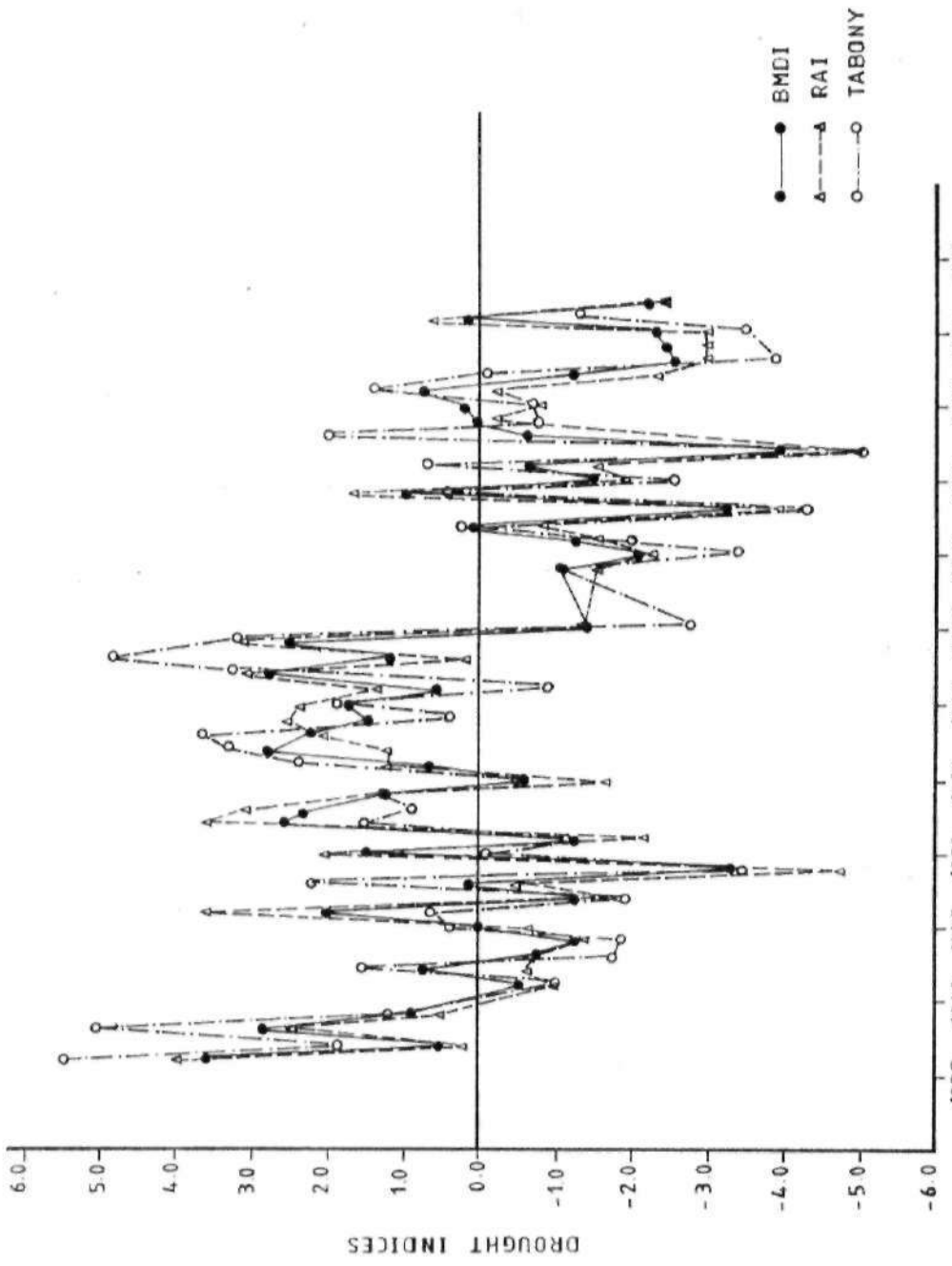


Fig. 4. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Potiskum (SAHELIAN ZONE).

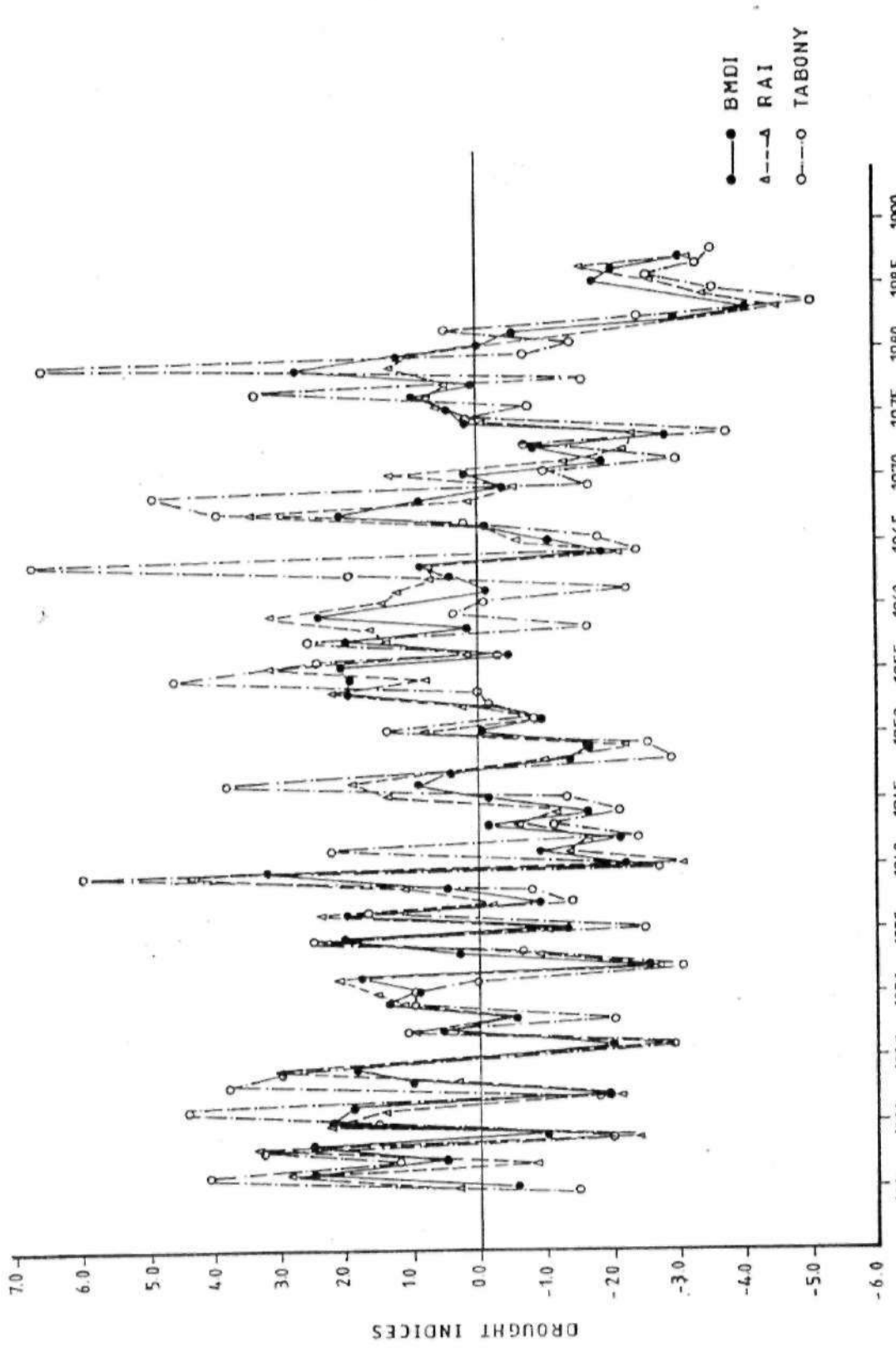


Fig. 5. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Maiduguri (SAHELIAN ZONE).

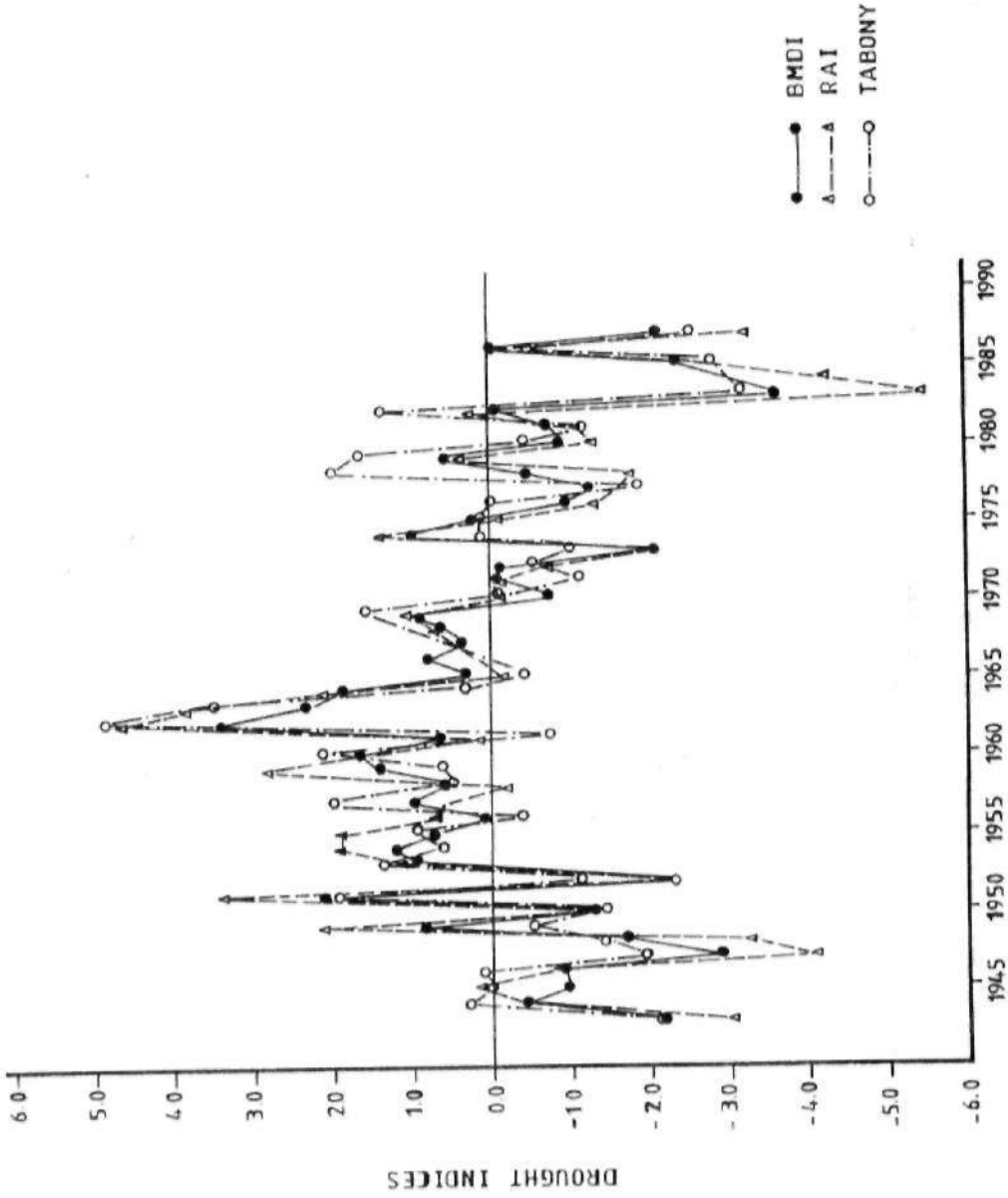


Fig. 6. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Yelwa (MIDLAND AREA).

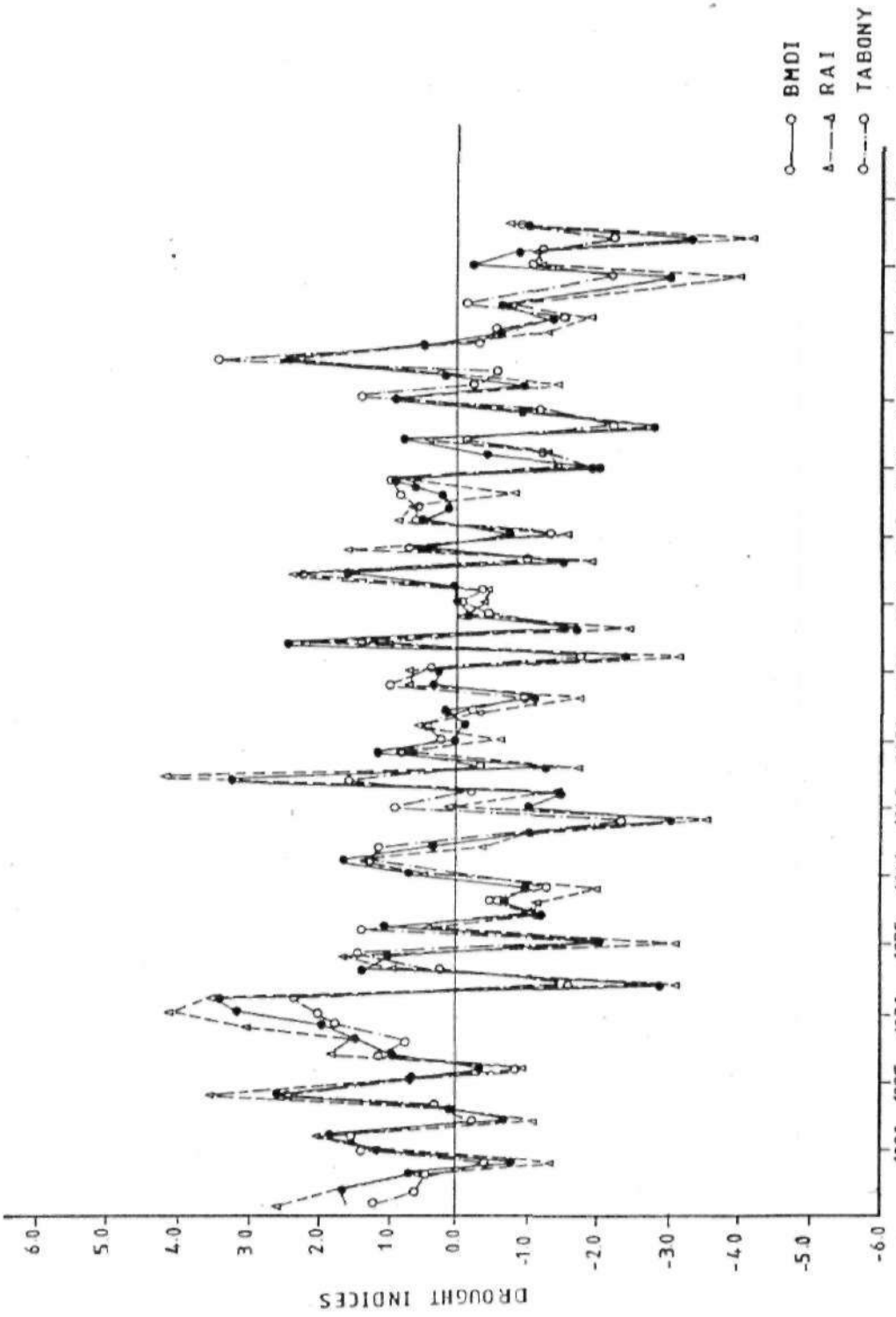


Fig. 7. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Minna (MIDLAND AREA).

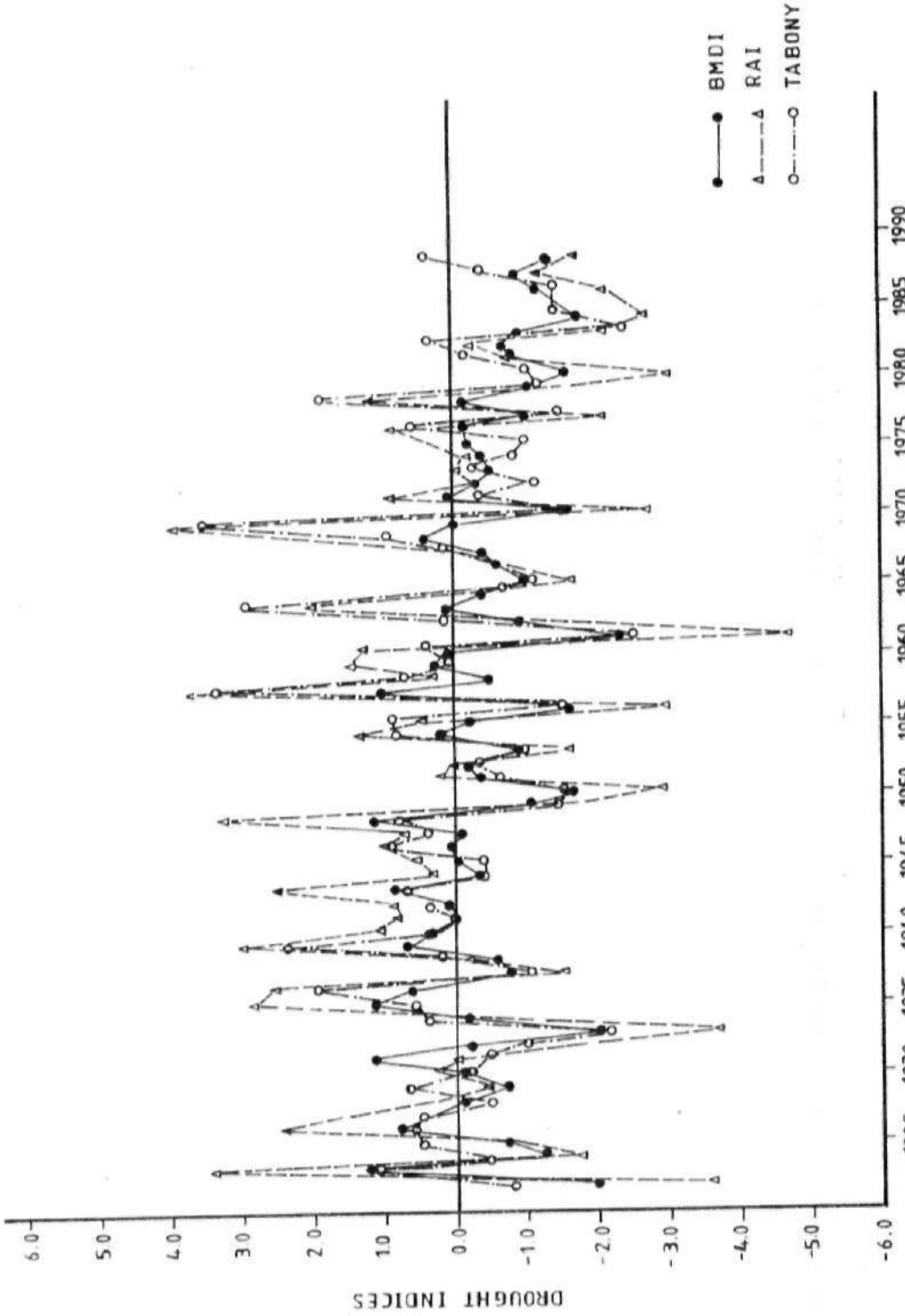


Fig. 8. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Jos (MIDLAND AREA).

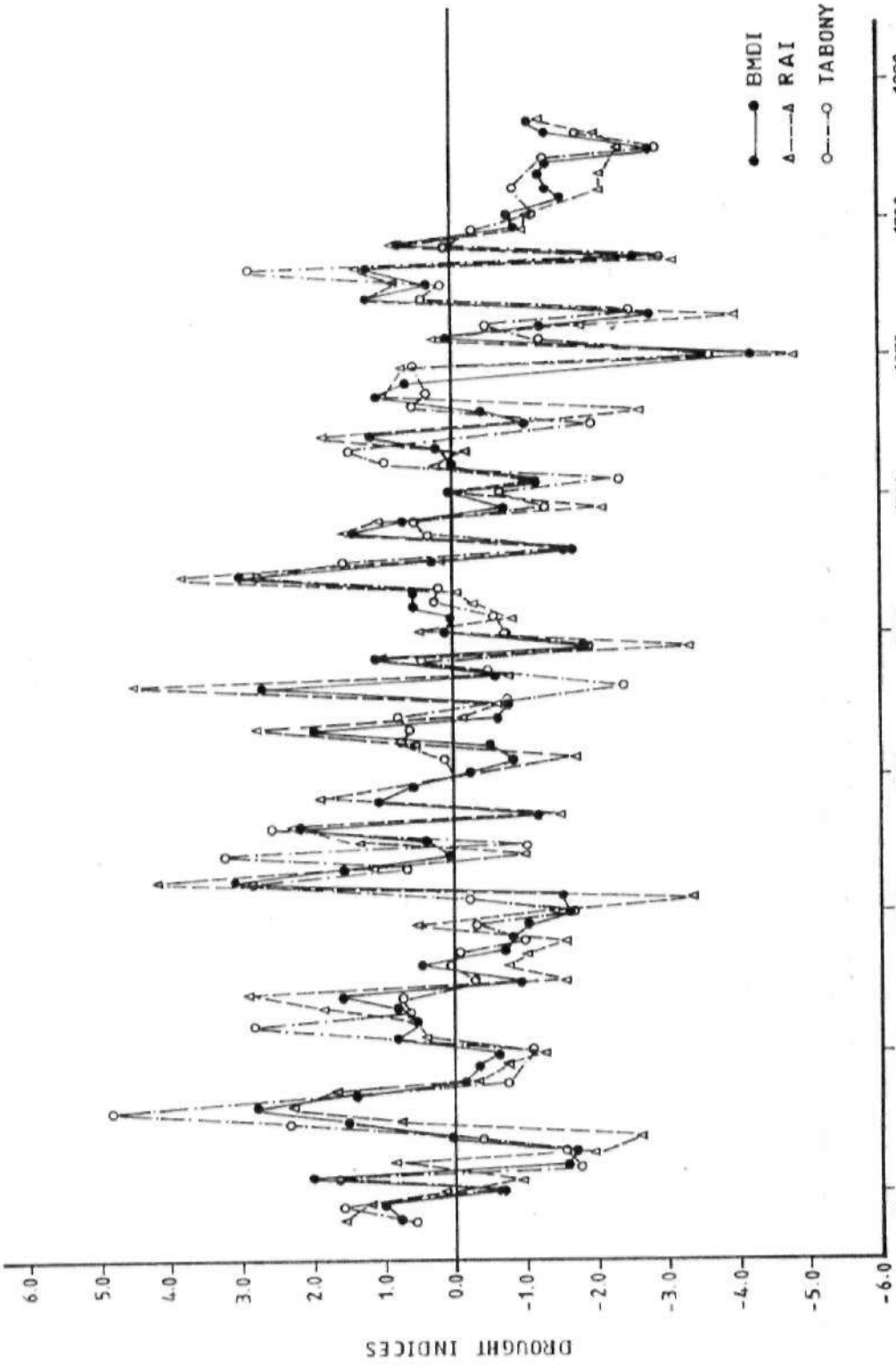


Fig. 9 Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Bauchi (MIDLAND AREA).

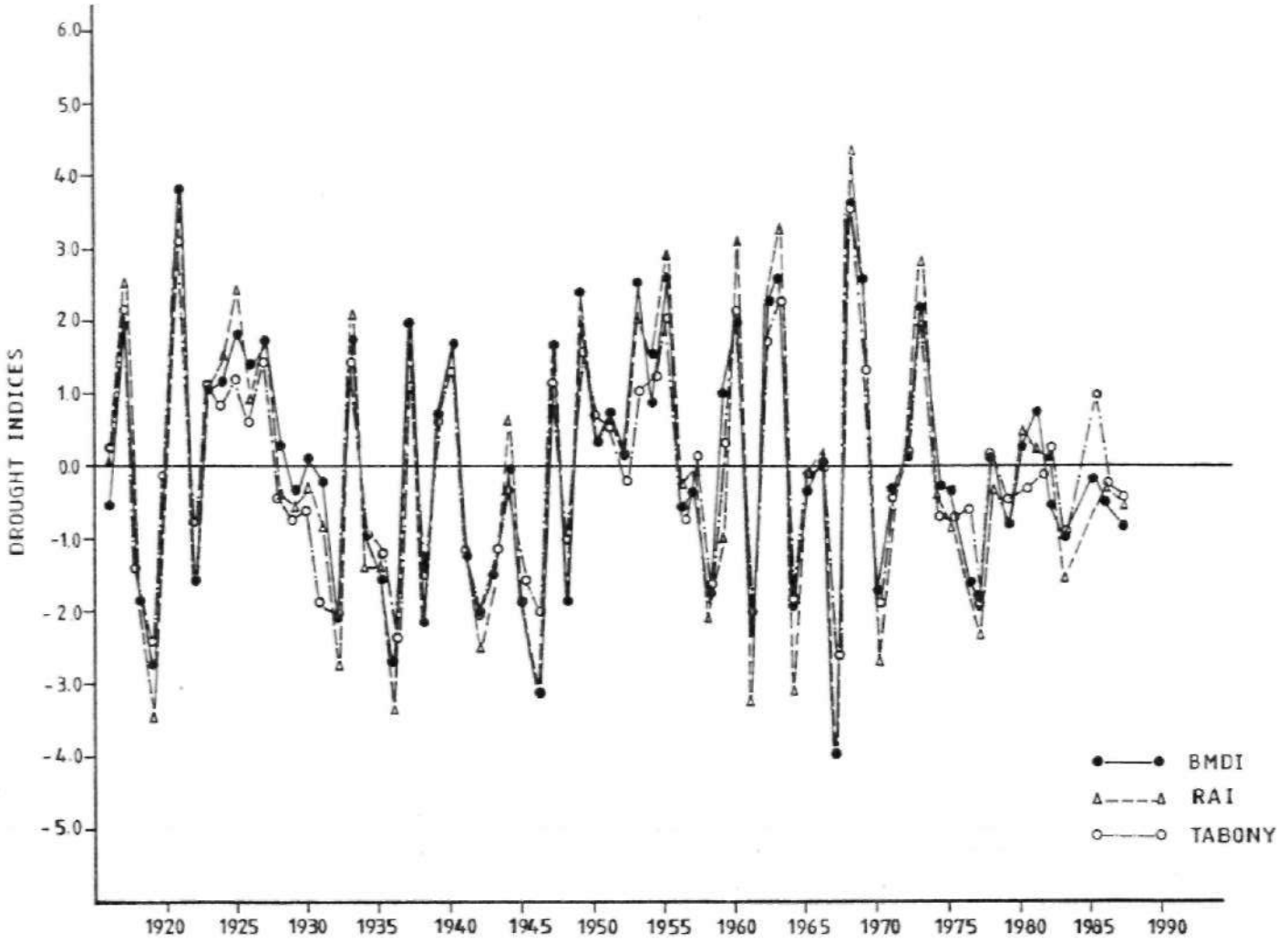


Fig. 10. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Ilorin (GUINEA SAVANNA ZONE).

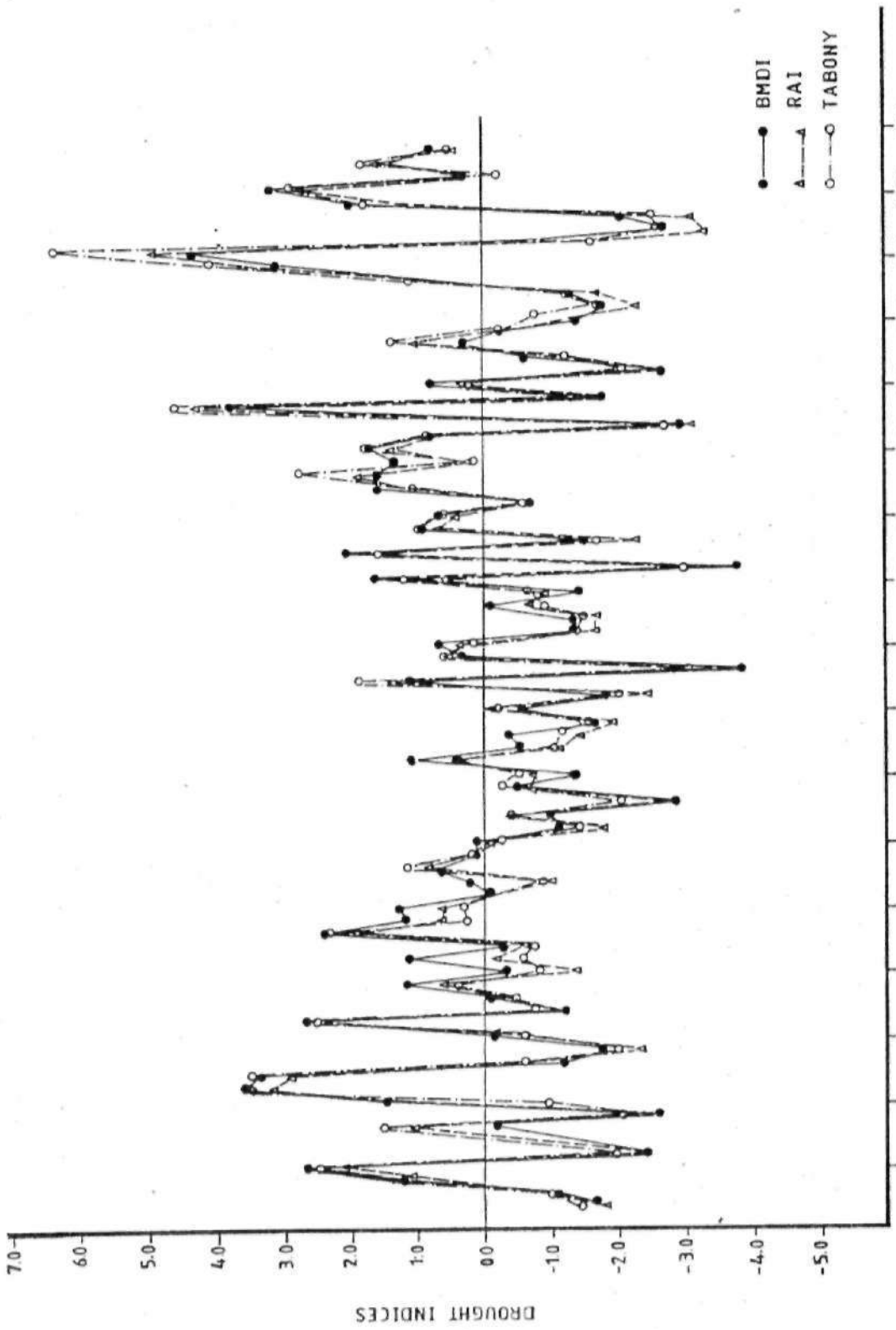


Fig. 11. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Ibadan (GUINEA SAVANNA ZONE).

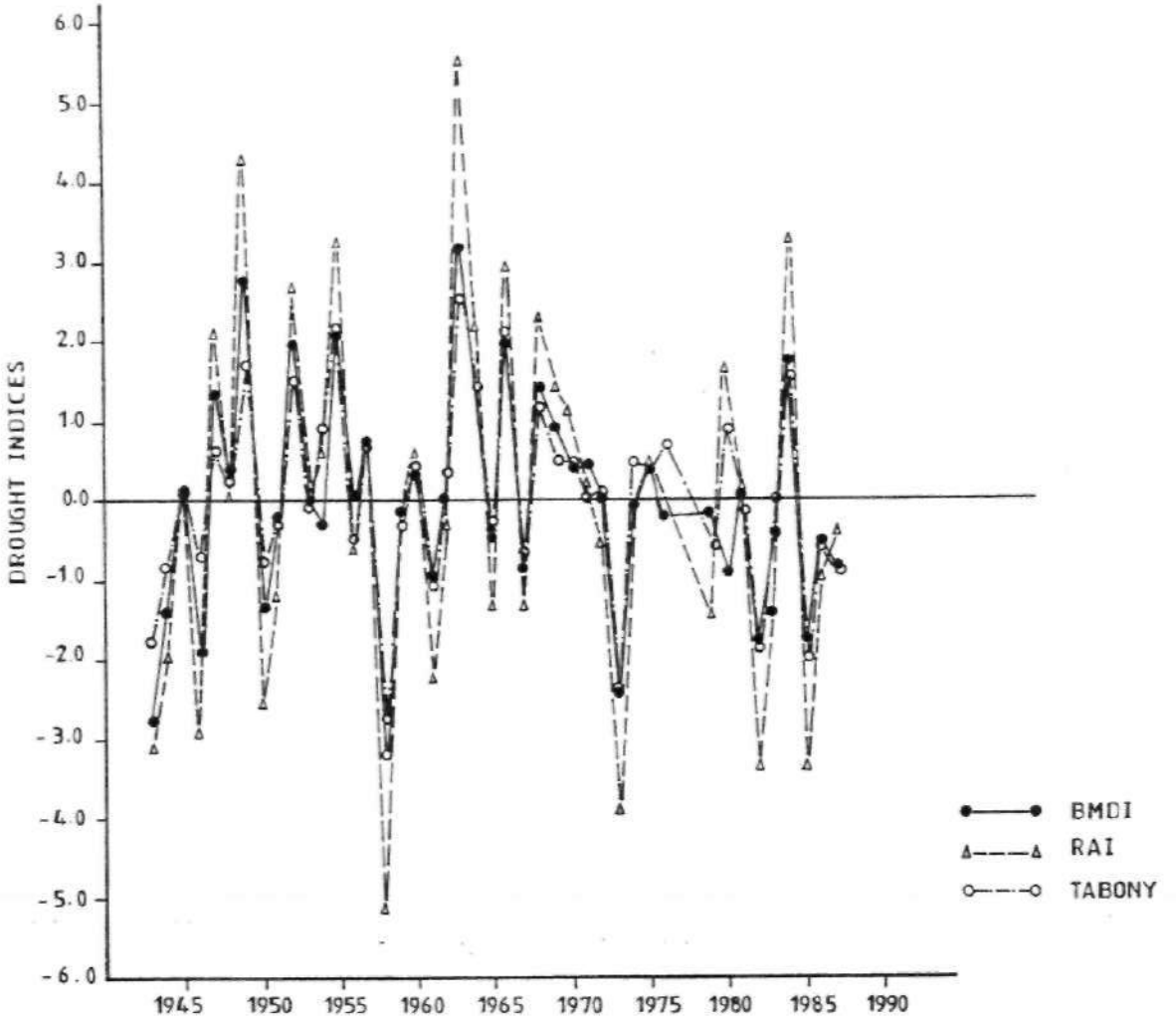


Fig. 12. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Makurdi (GUINEA SAVANNA ZONE).

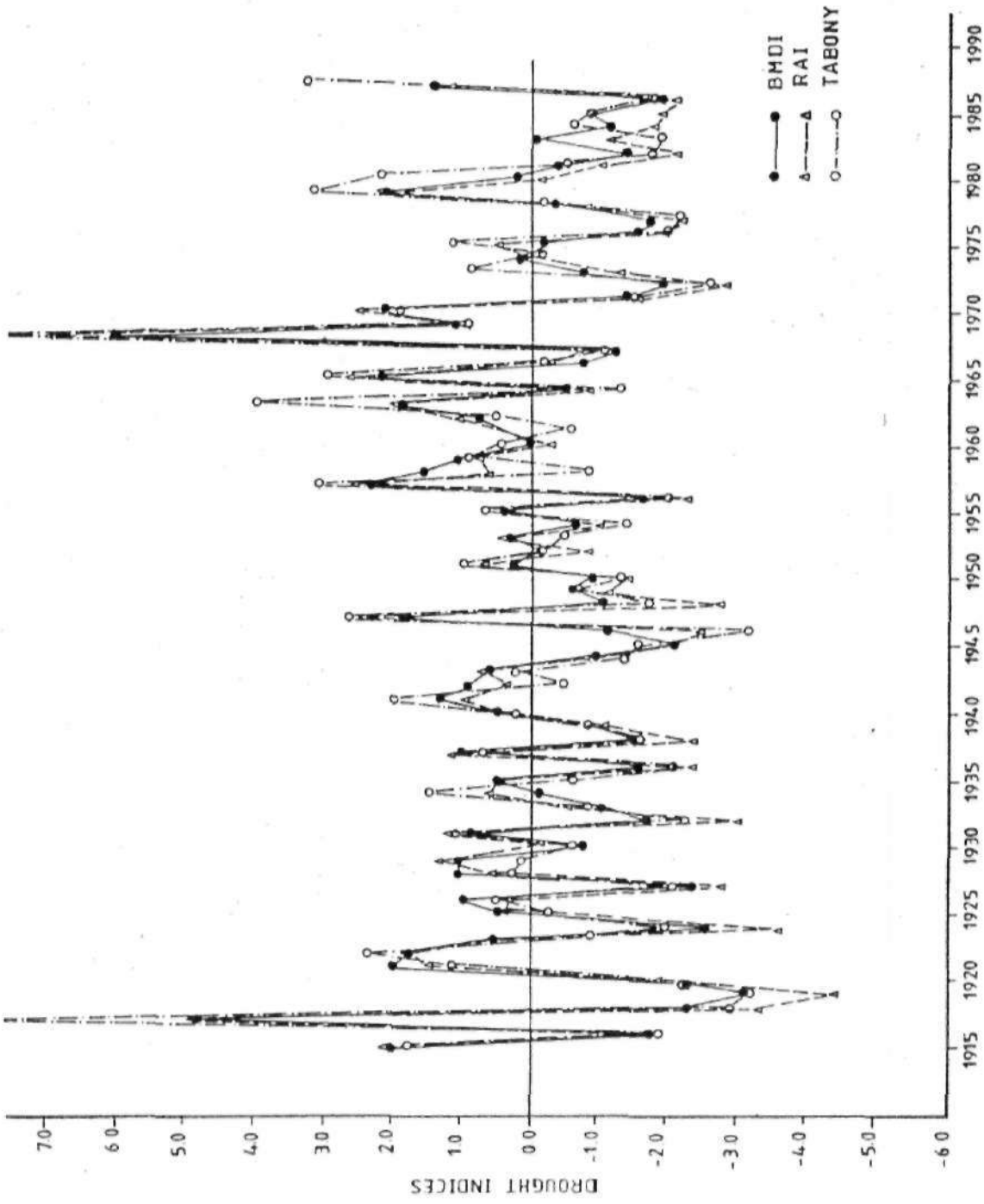


Fig. 13. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Lagos (COASTAL ZONE).

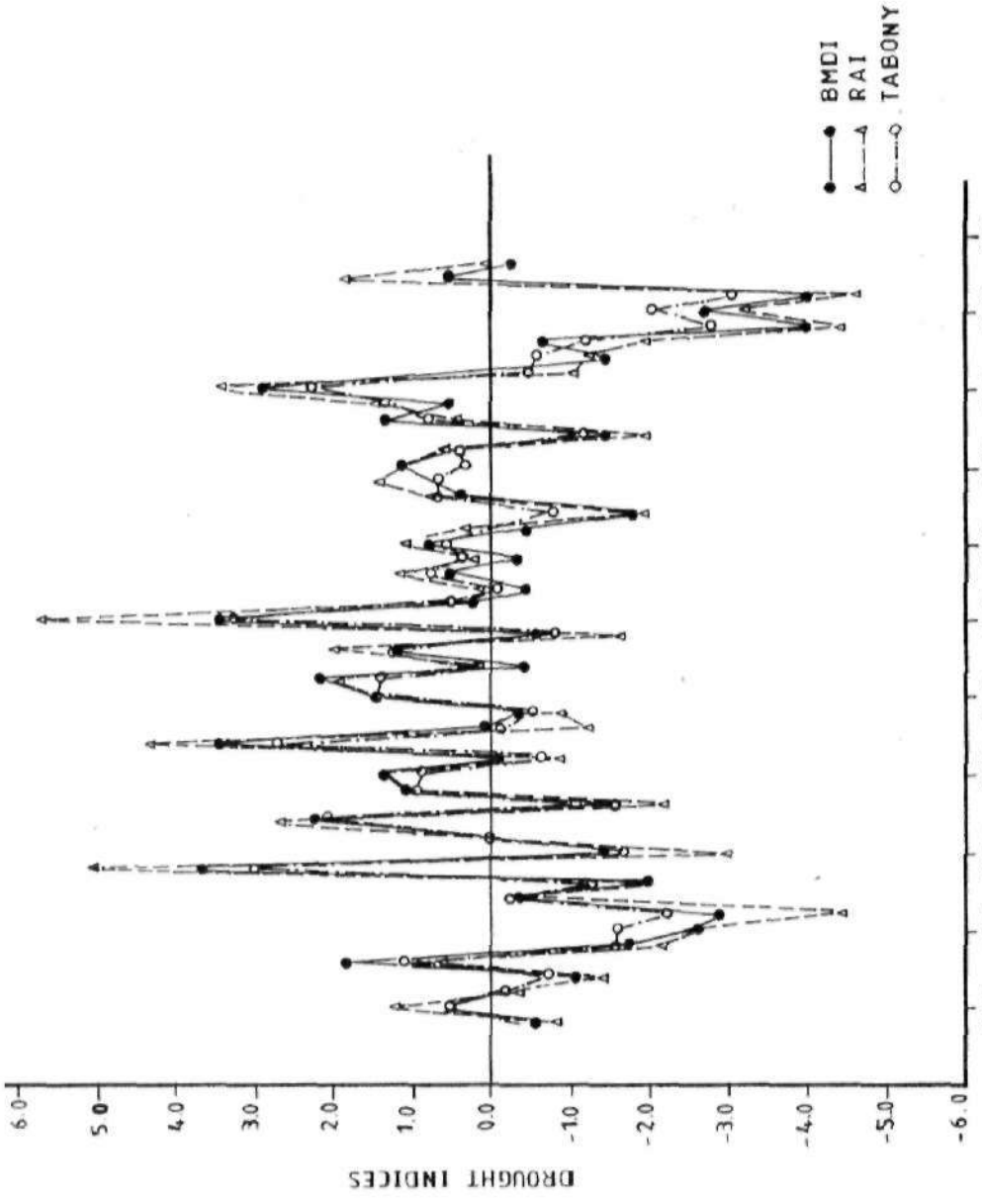


Fig.14. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Benin-City (COASTAL ZONE).

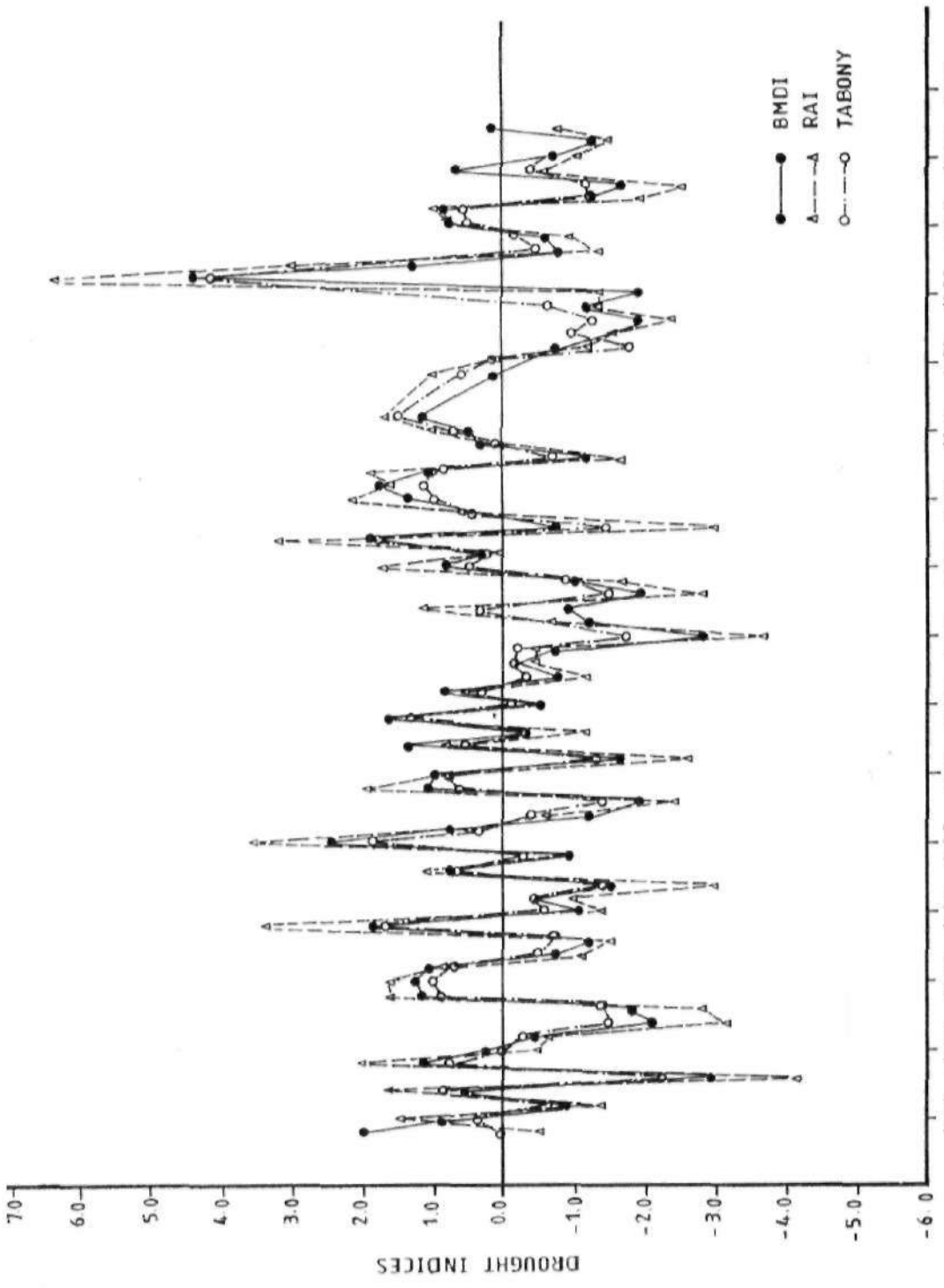


FIG. 15 : Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Port-Harcourt (COASTAL ZONE).

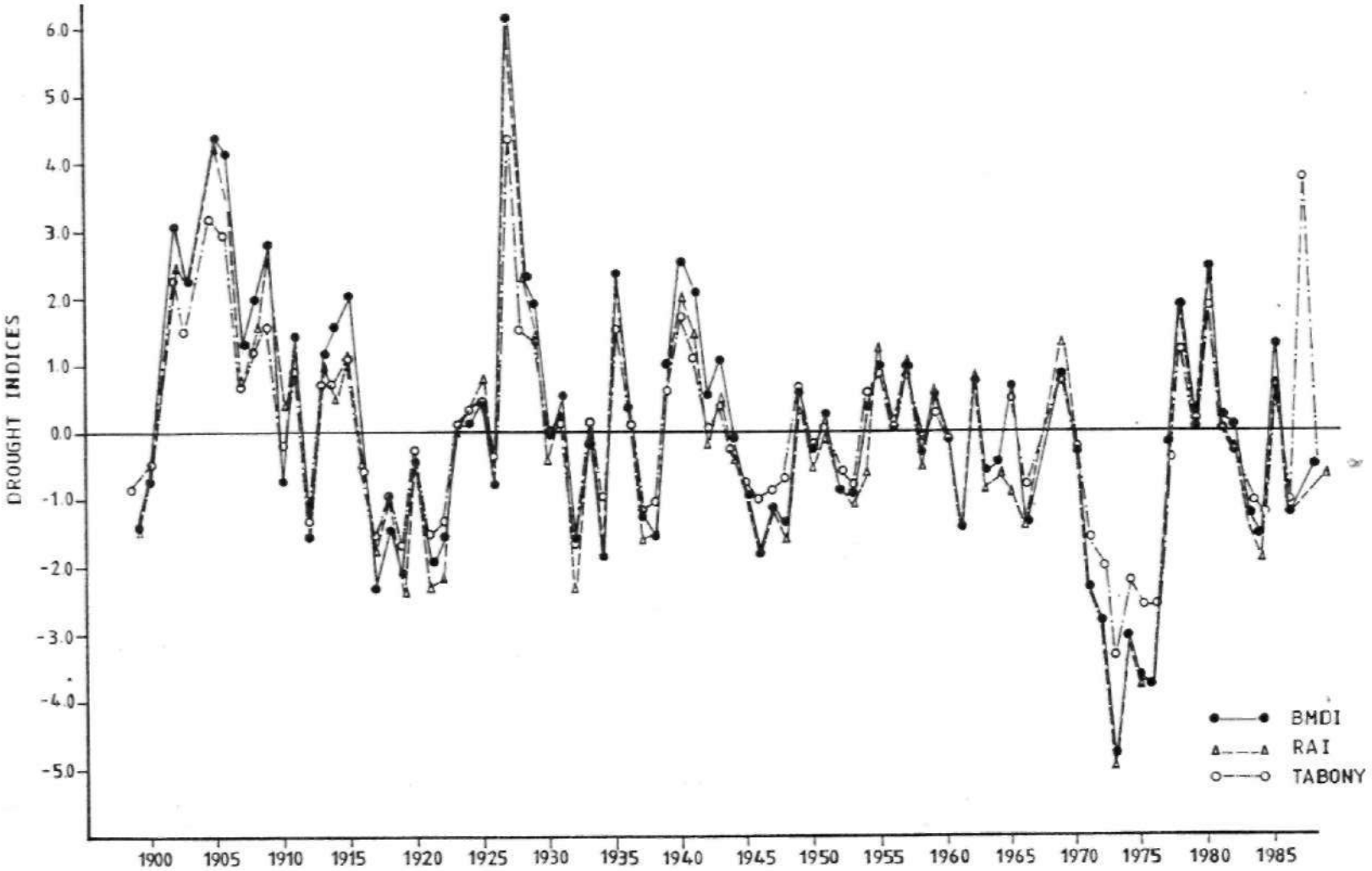
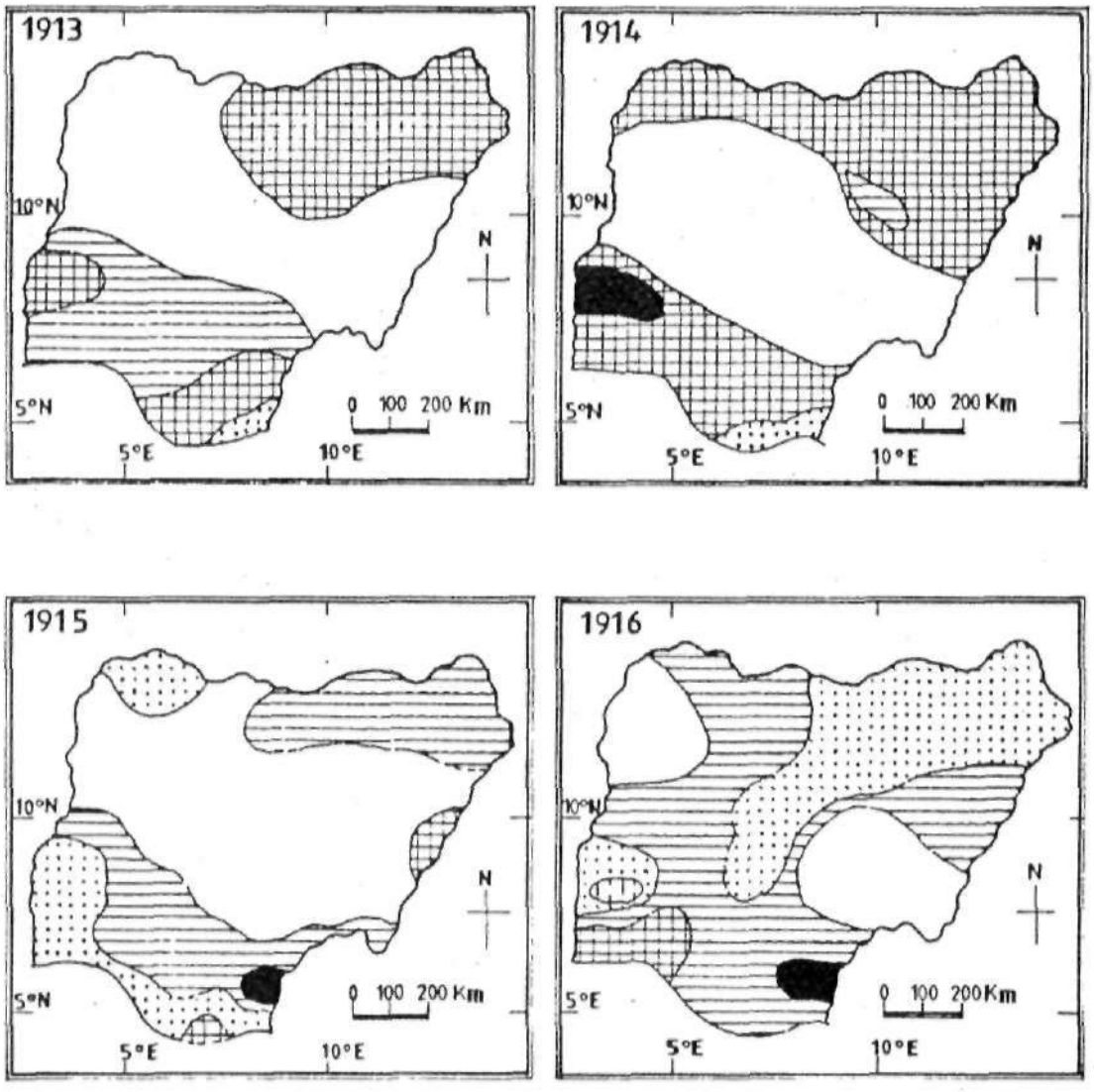


Fig. 16. Seasonal means of Bhalme and Mooley (BMDI), Van Rooy (RAI) and Tabony drought indices for Calabar (COASTAL ZONE).

APPENDIX B

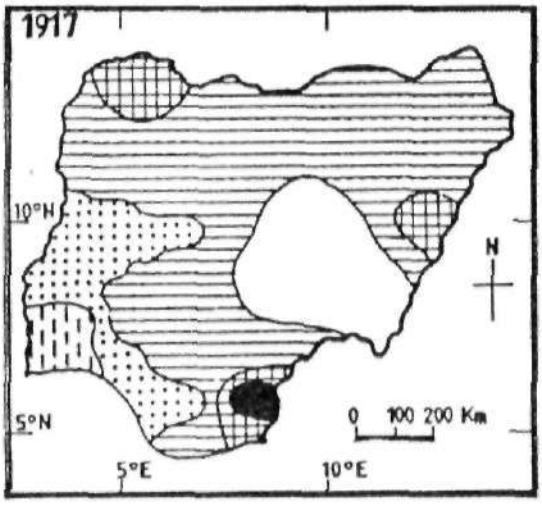
INTERANNUAL VARIATIONS OF MOISTURE ANOMALY PATTERNS FOR 1913-1987

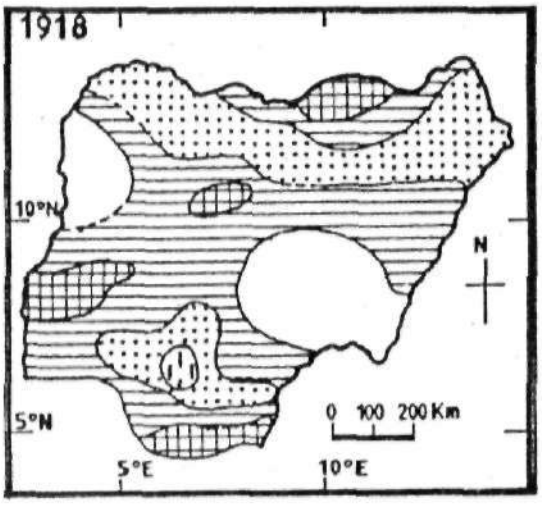


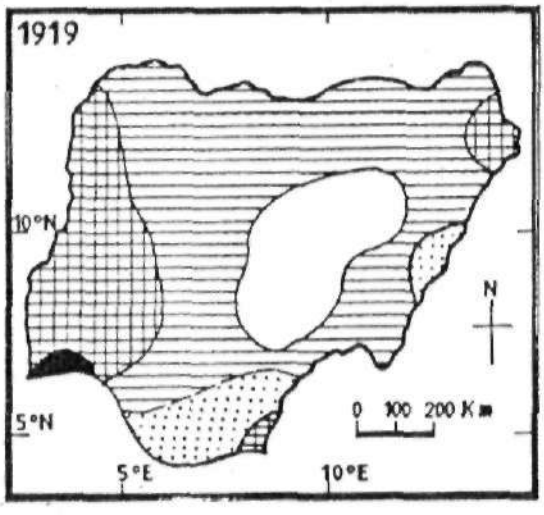
Appendix B: Interannual variations of moisture anomaly patterns for 1913-1916.

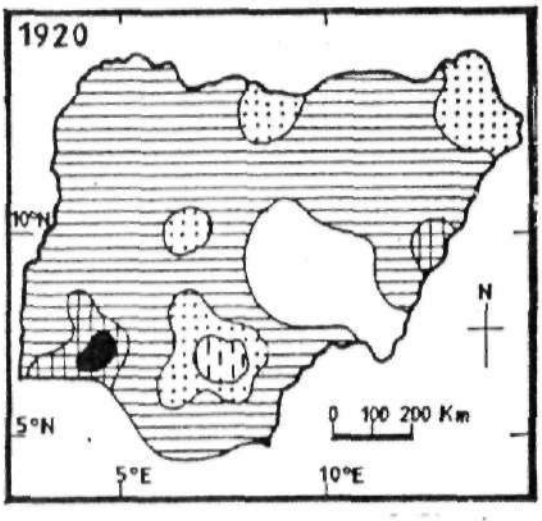
LEGEND

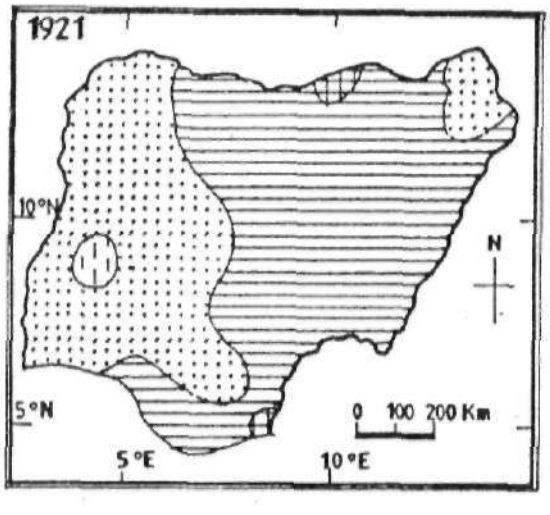
- Areas with no data
- Very wet (BMDI > 3.0)
- Moderately wet (3.0 ≥ BMDI > 1.0)
- Near normal (1.0 ≥ BMDI > -1.0)
- Moderate drought (-1.0 ≥ BMDI > -3.0)
- Severe drought (BMDI < -3.0)

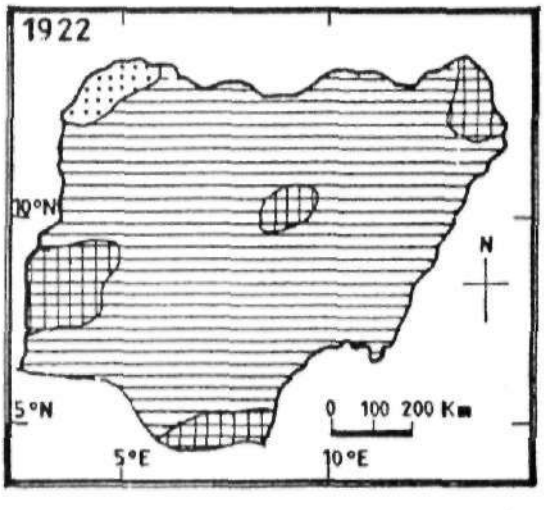




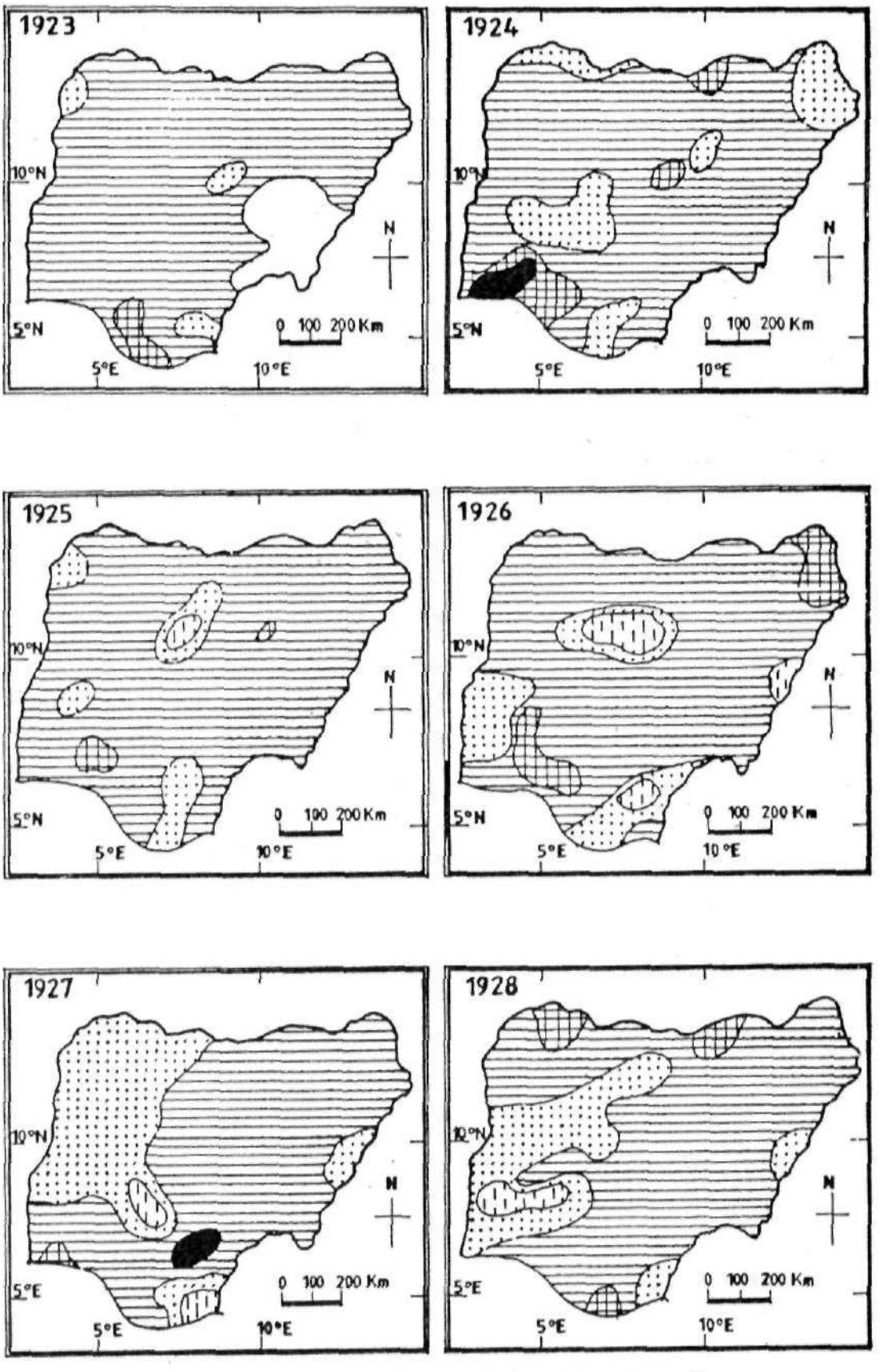




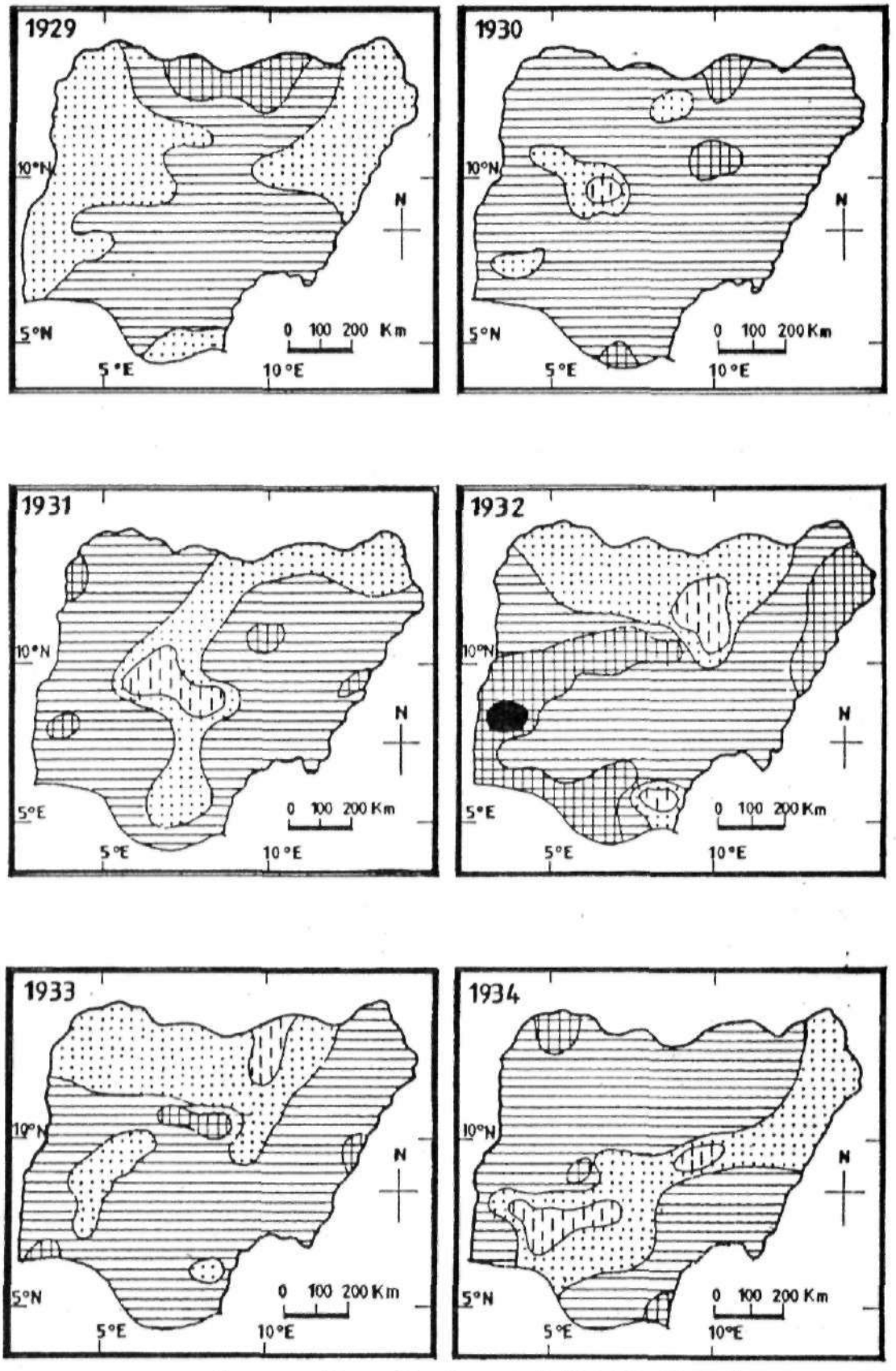




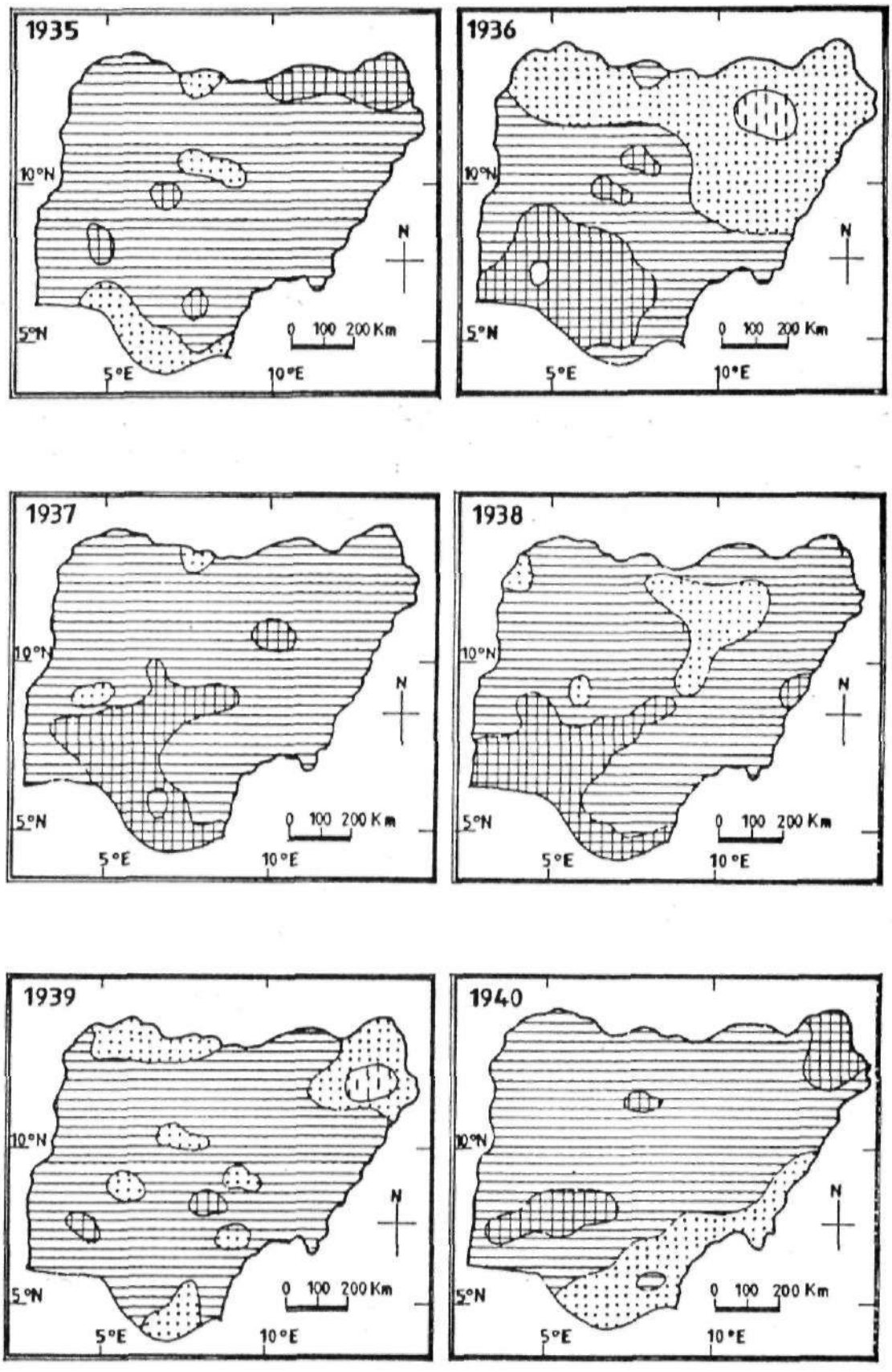
Appendix B (cont.) (1917-1922)



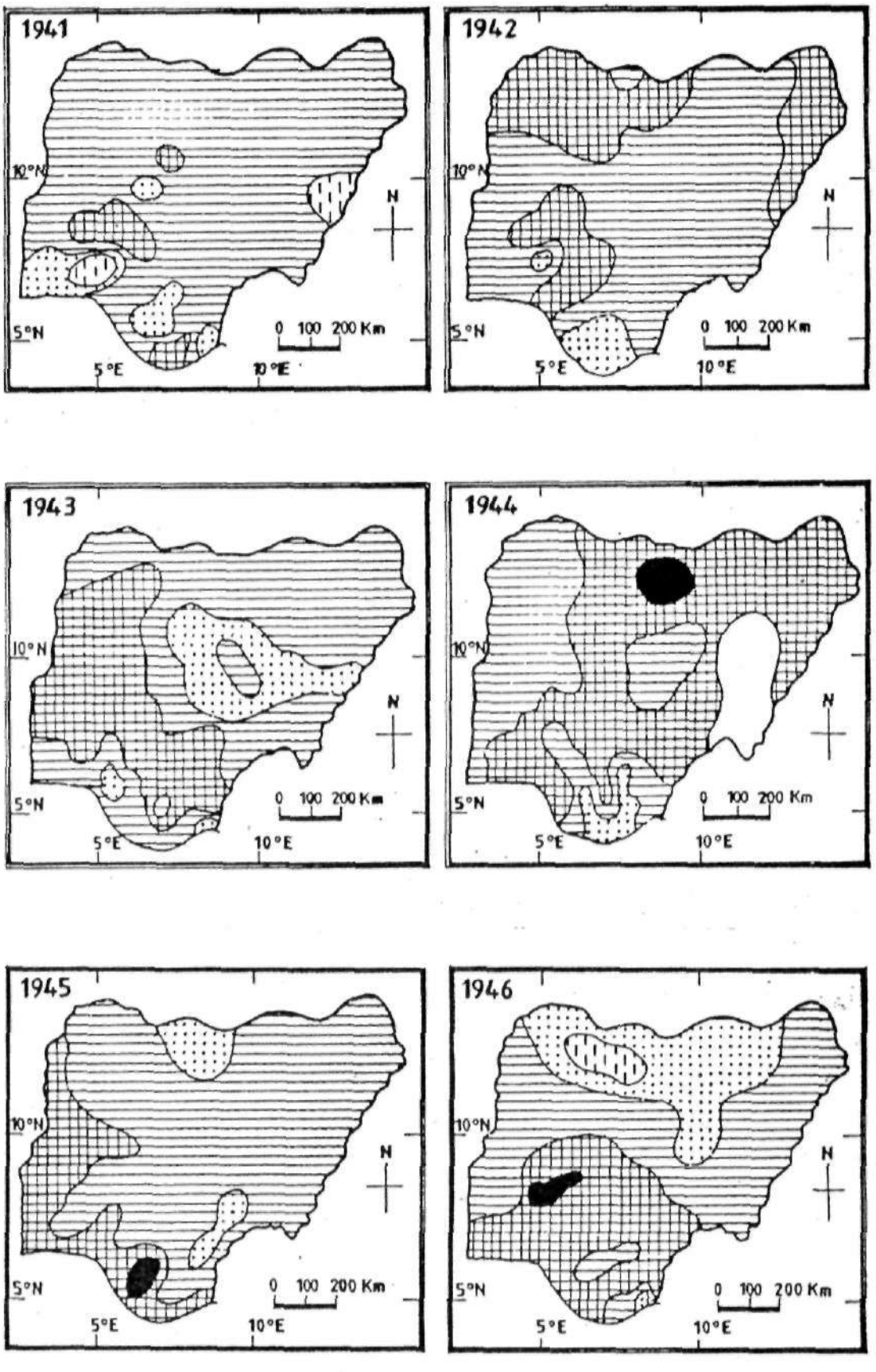
Appendix B (cont.) (1923-1928)



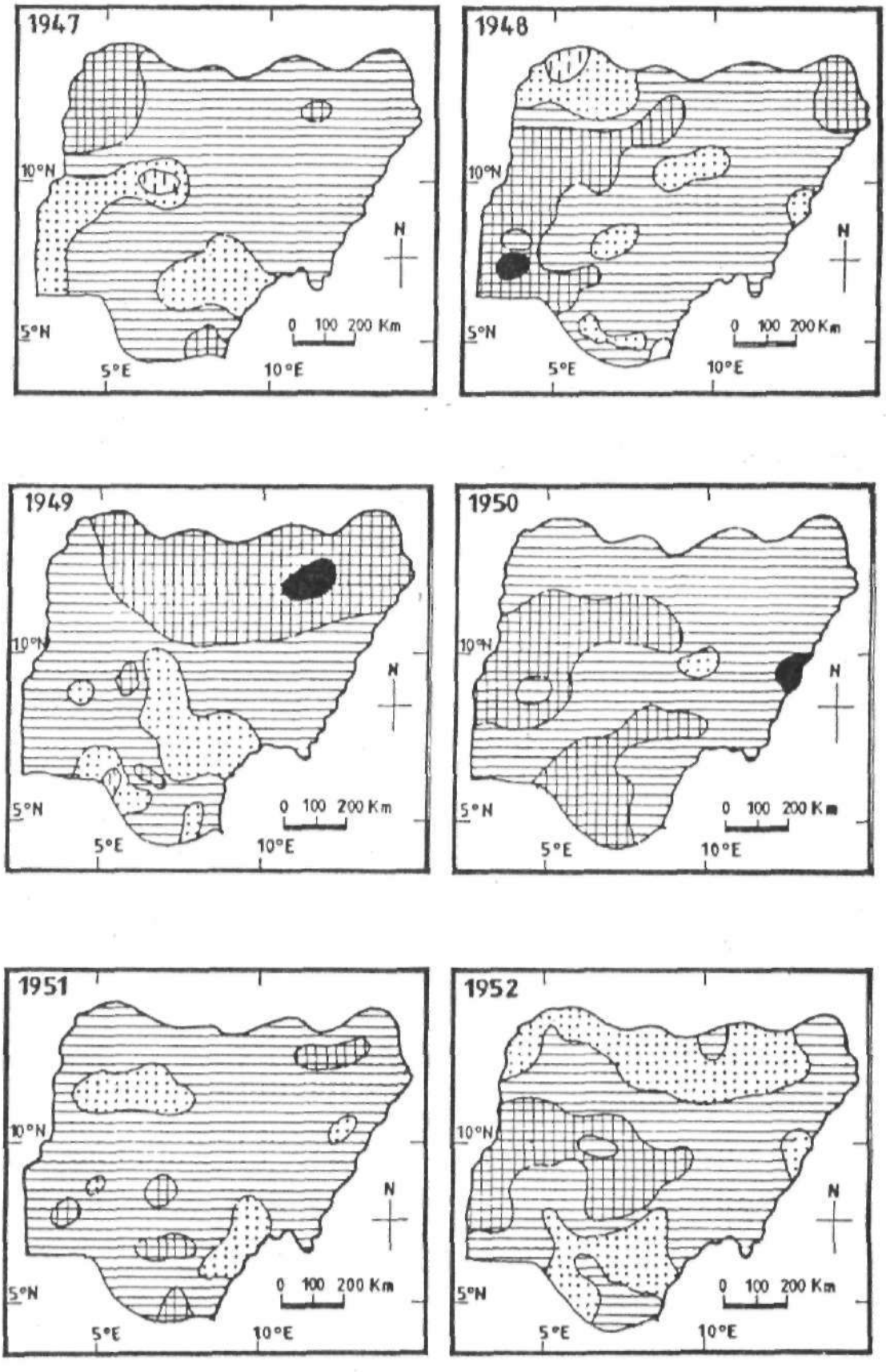
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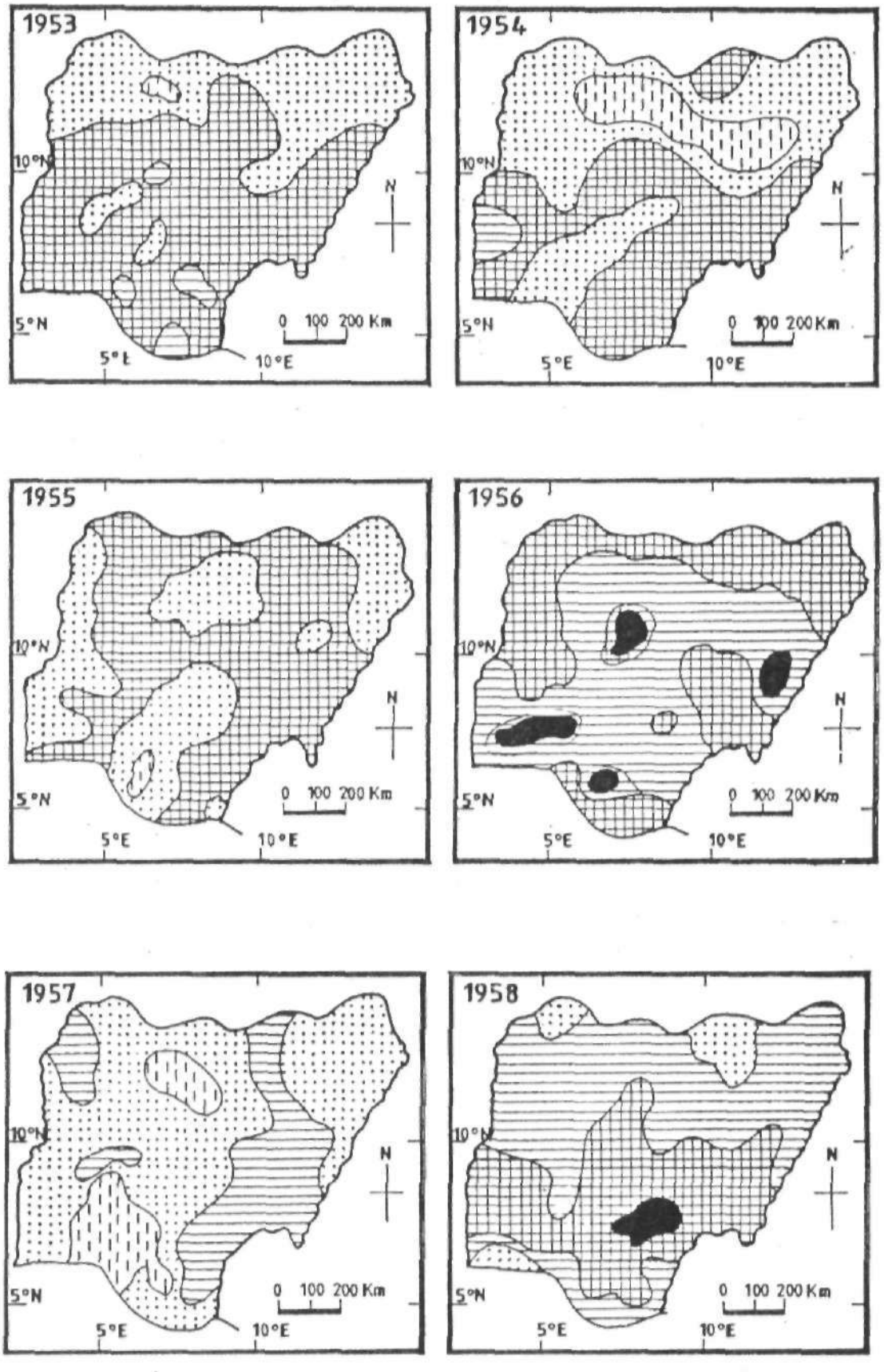
Appendix B (cont.) (1935 - 1940)



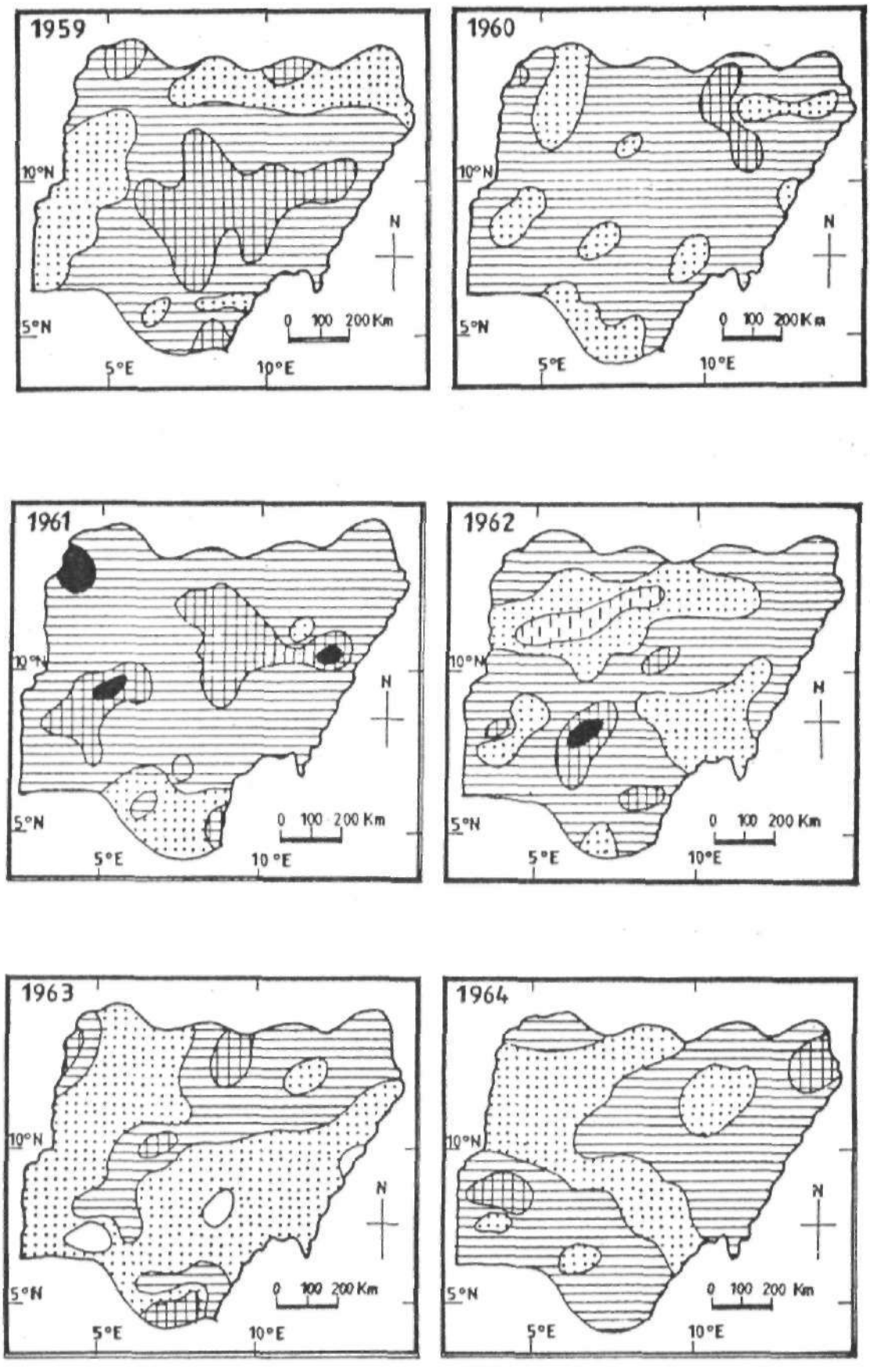
Appendix B (cont.) (1941 - 1946)



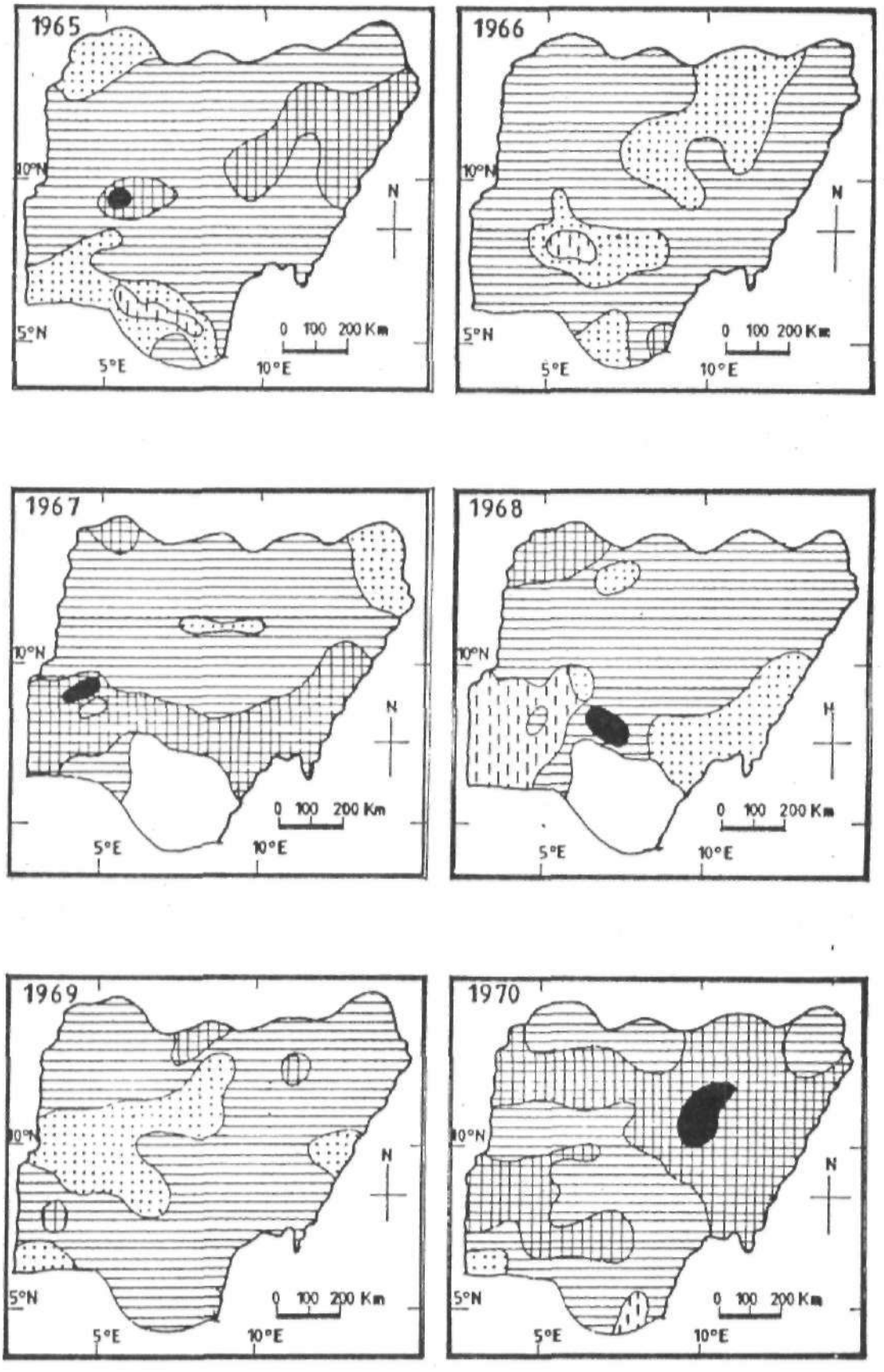
Appendix B (cont.) (1947 - 1952)



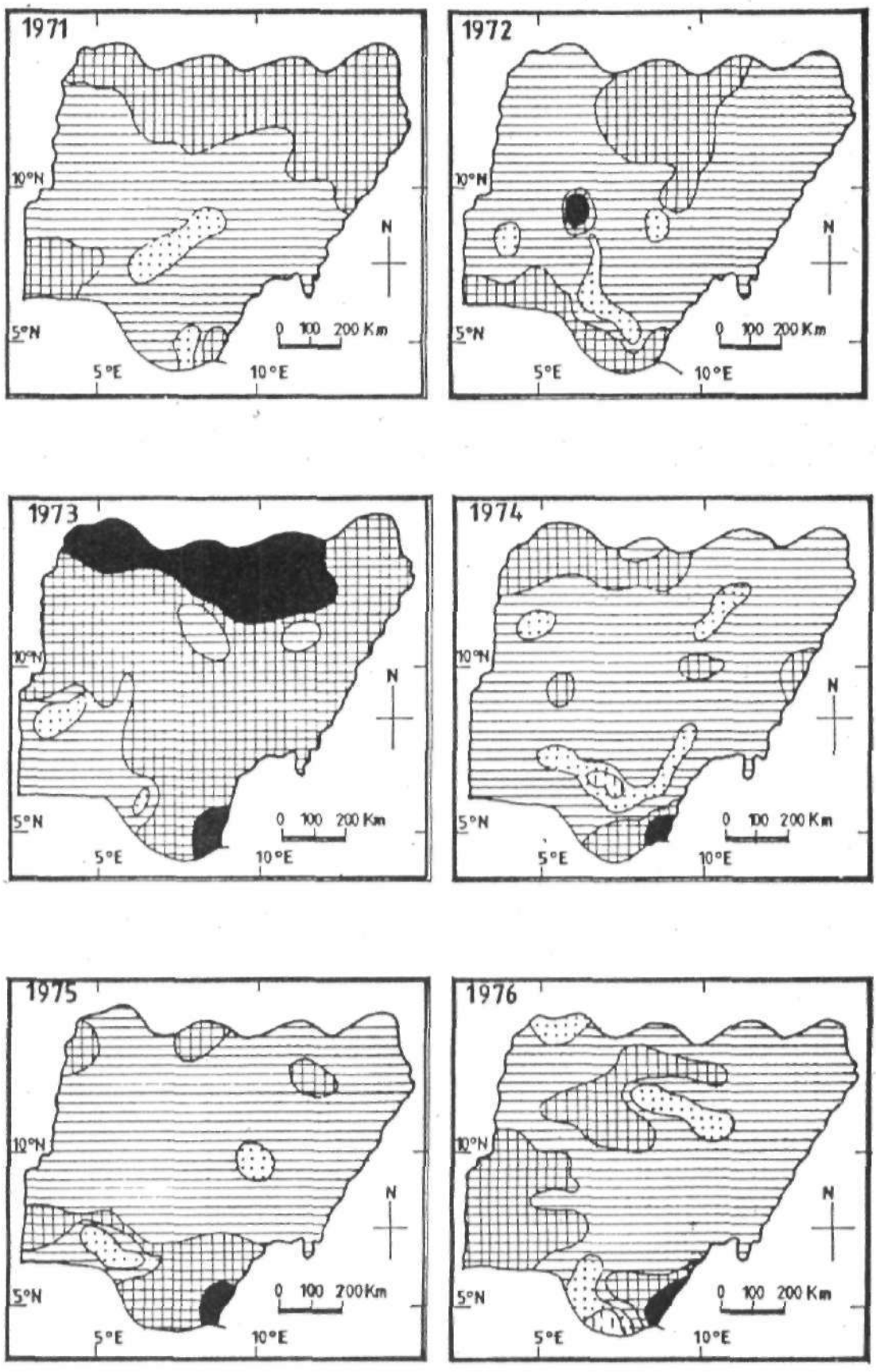
Appendix B (cont.) (1953 - 1958)



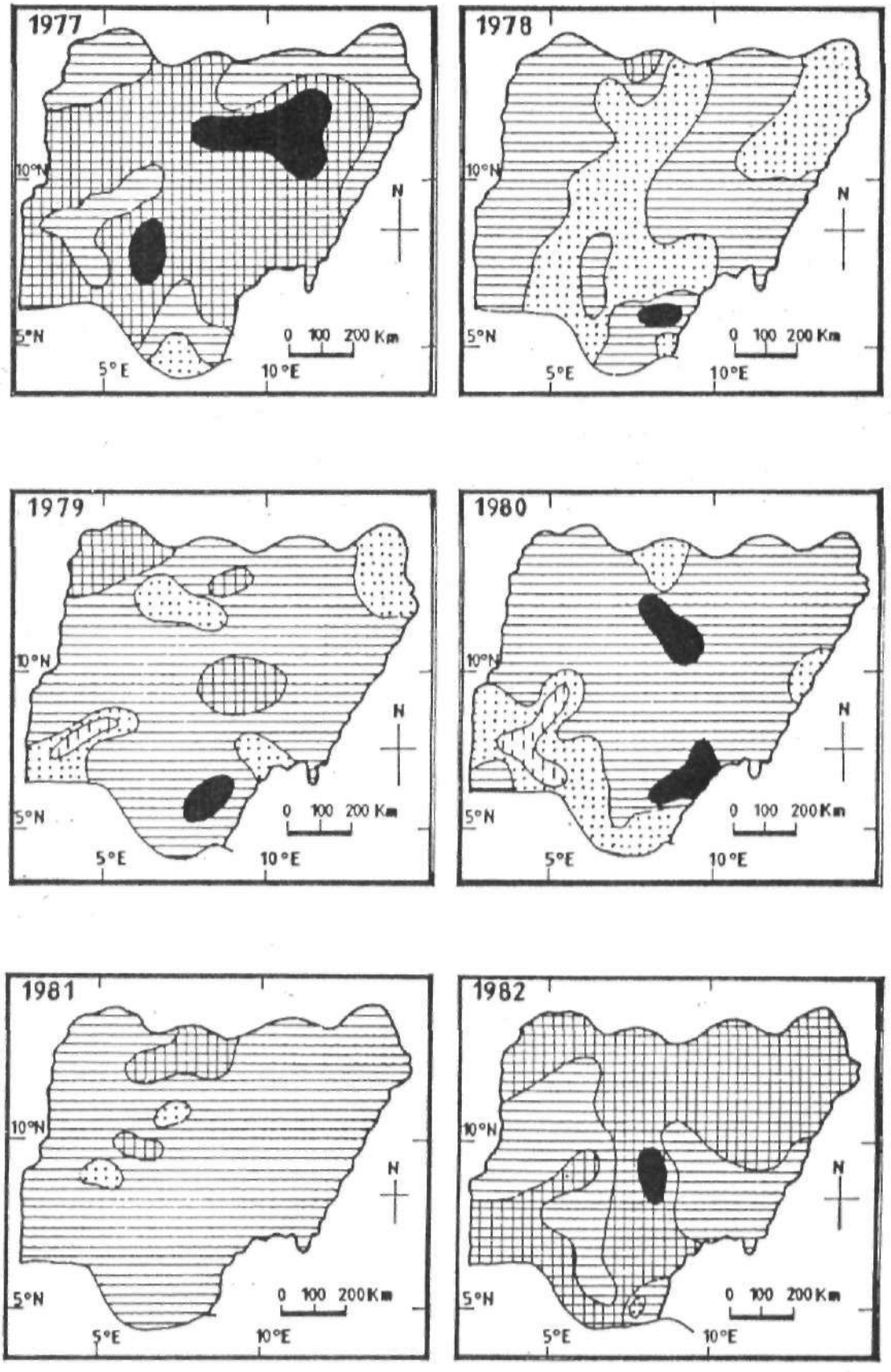
Appendix B (cont.) (1959-1964)



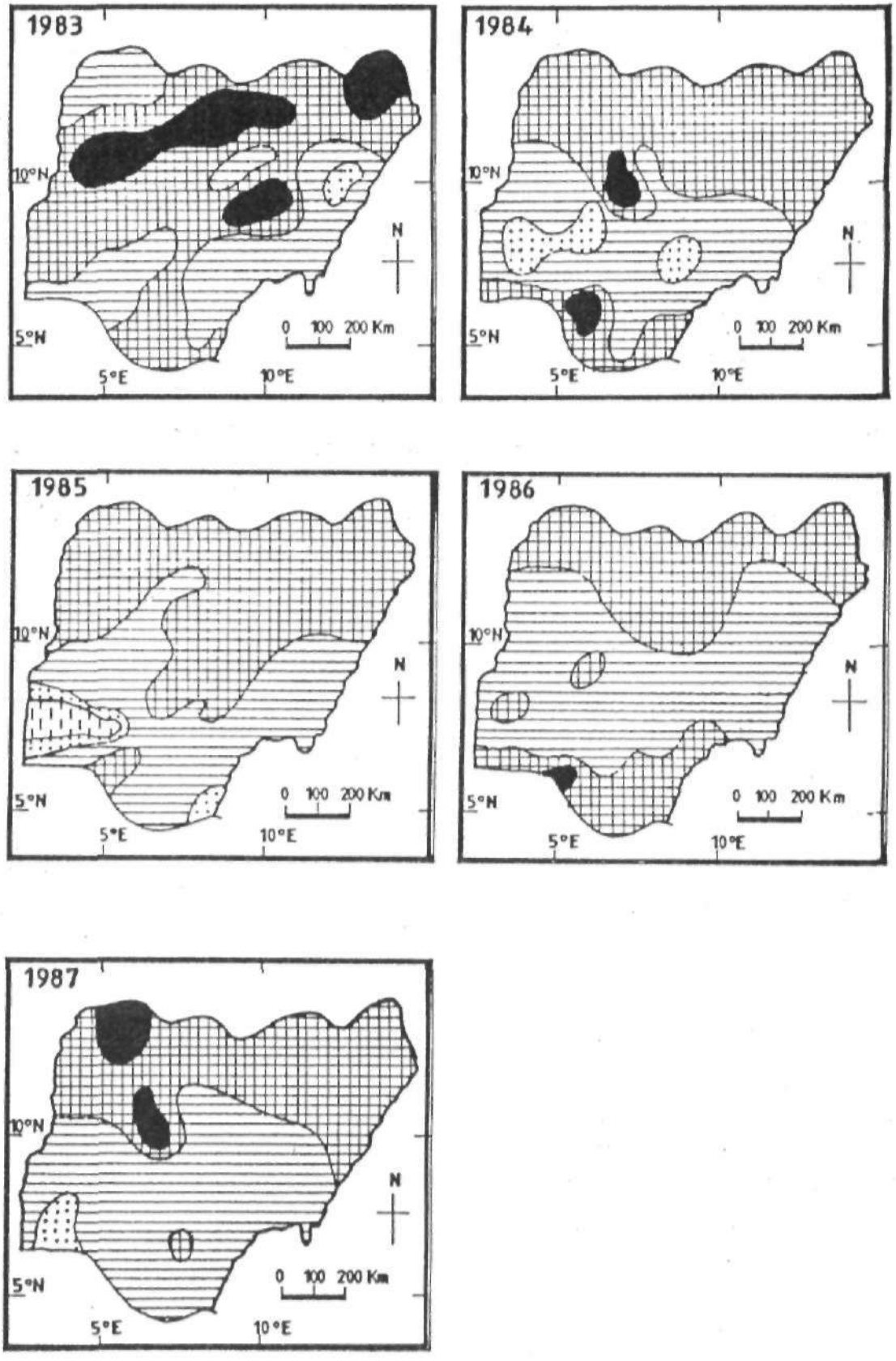
Appendix B (cont.) (1965-1970)



Appendix B (cont.) (1971 - 1976)



Appendix B (cont.) (1977-1982)

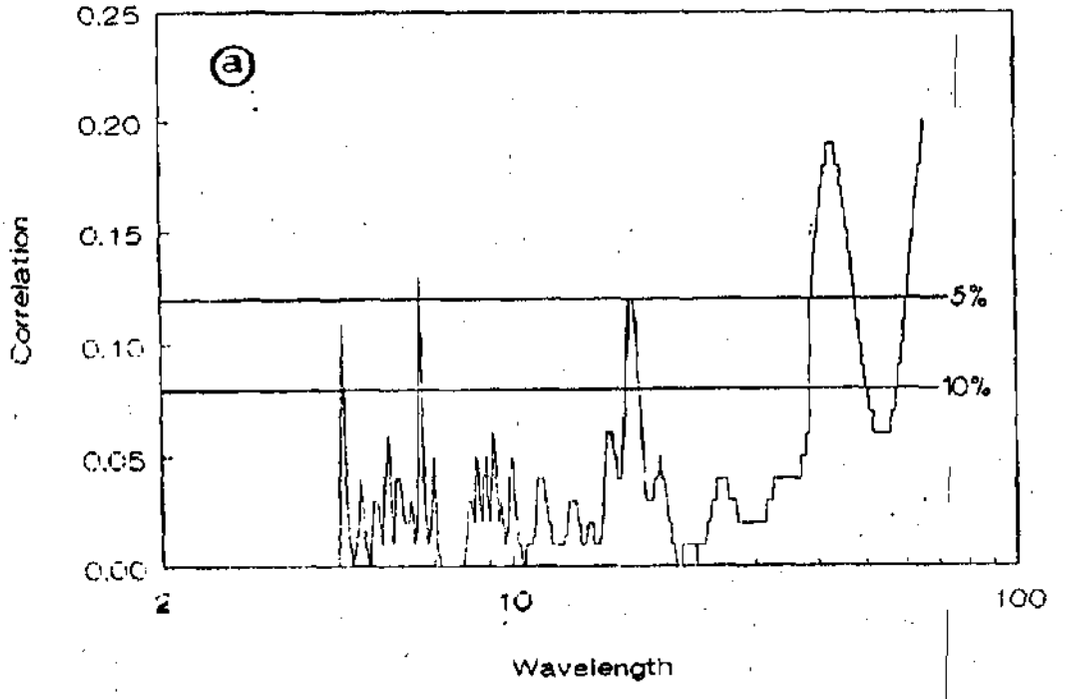


Appendix B (cont.) (1983-1987)

APPENDIX C

NON-INTEGER SPECTRA OF DROUGHT FOR SELECTED STATIONS IN NIGERA

PLOT OF SOKOTO STATION



PLOT OF BIRNIN KEBBI STATION

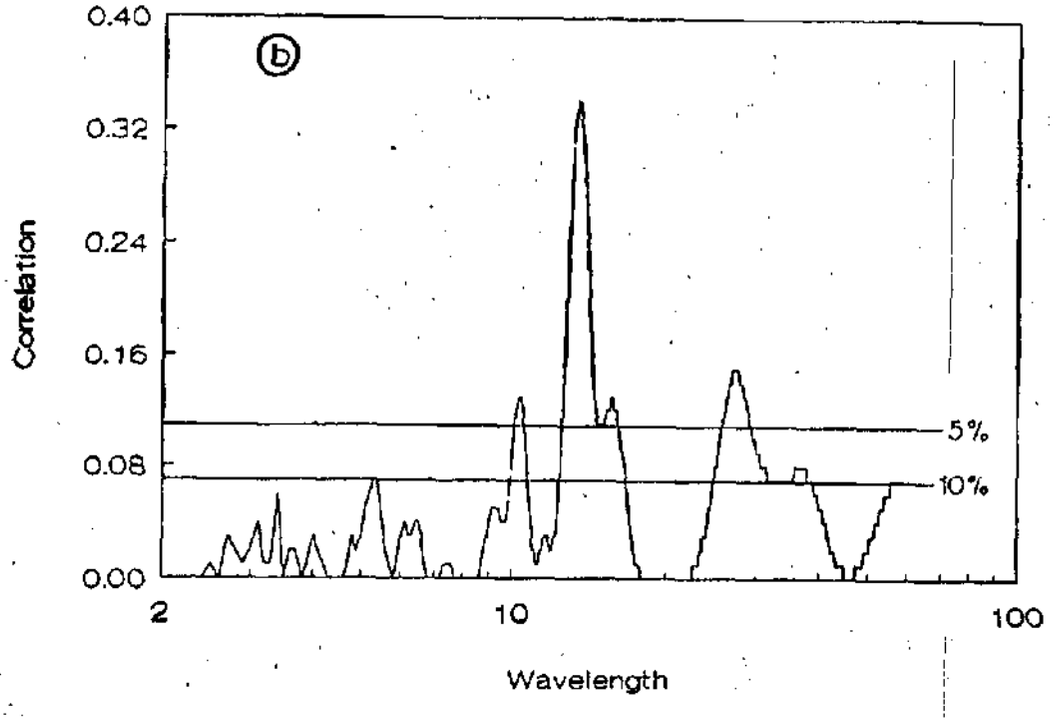
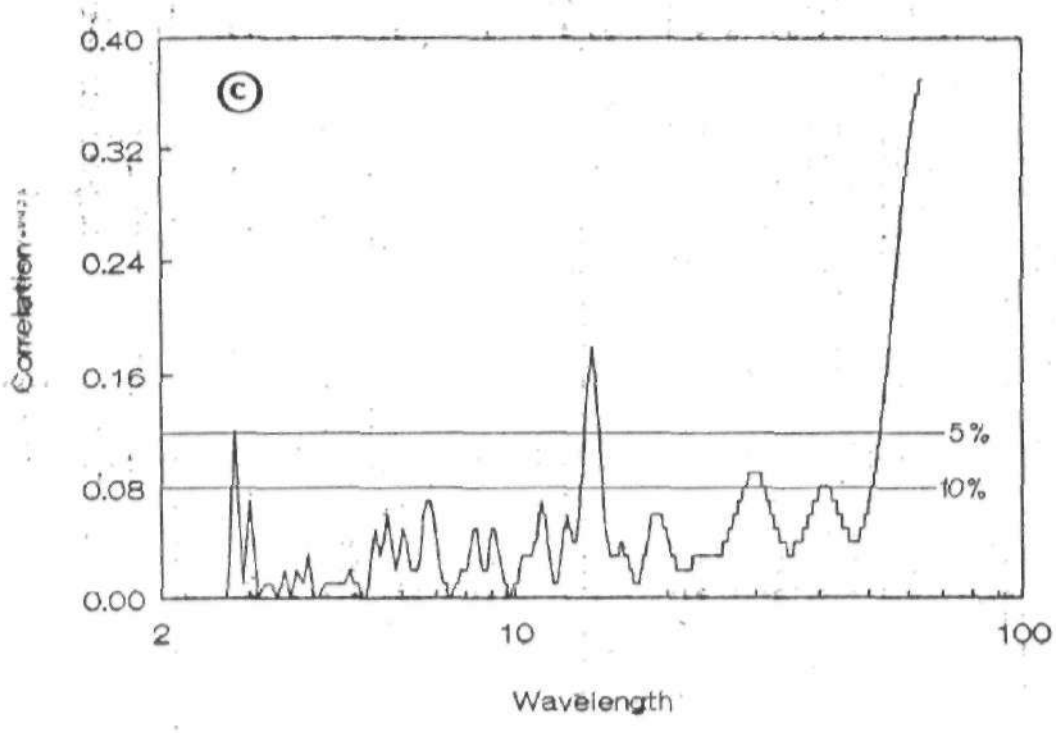


Fig 1 (a-f). NON-INTEGER SPECTRA OF DROUGHT FOR SELECTED STATIONS IN THE SAHELIAN ZONE a) SOKOTO b) BIRNIN-KEBBI

PLOT OF KATSINA STATION



PLOT OF KANO STATION

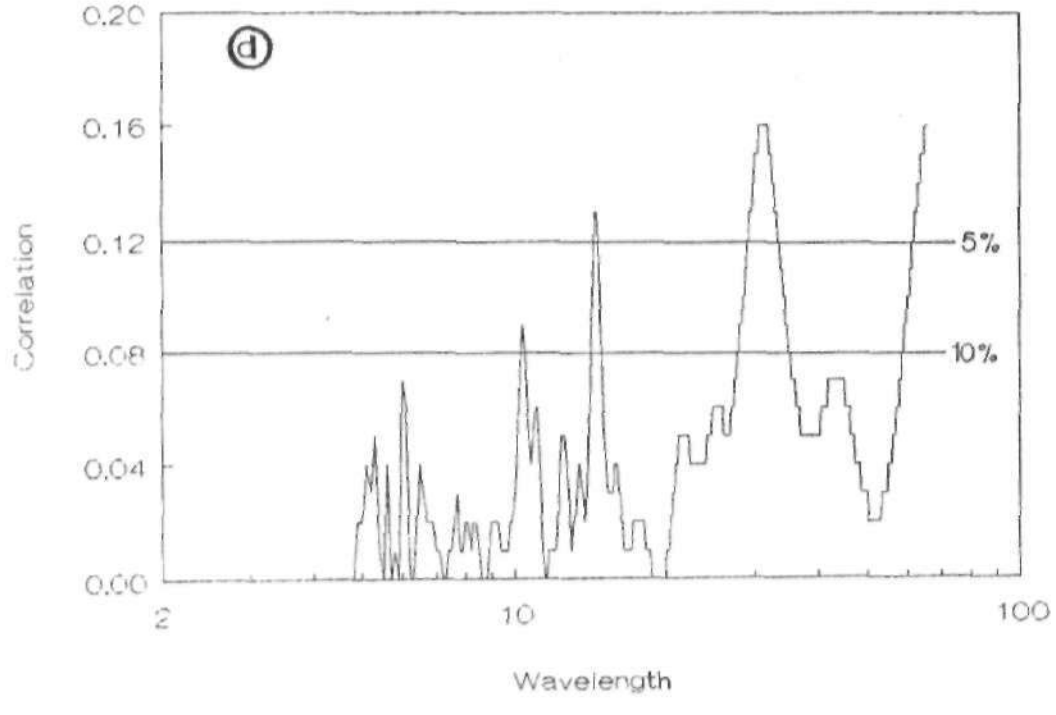
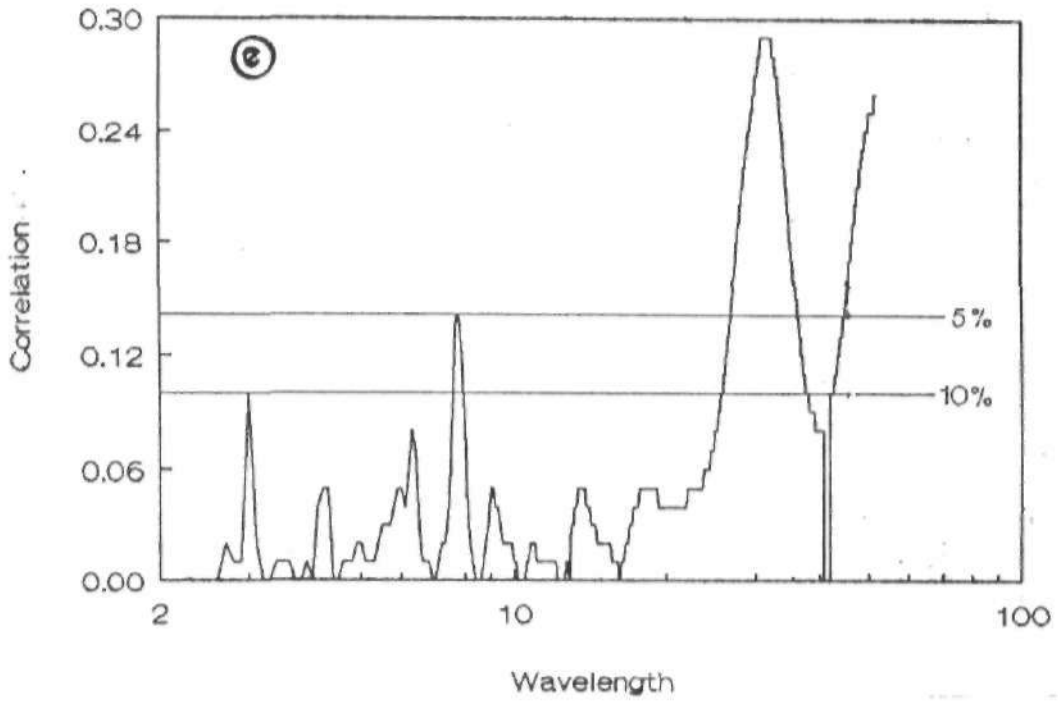


Fig. 1 (cont). for ⓒ Katsina ⓓ Kano

PLOT OF POTISKUM STATION



PLOT OF MAIDUGURI STATION

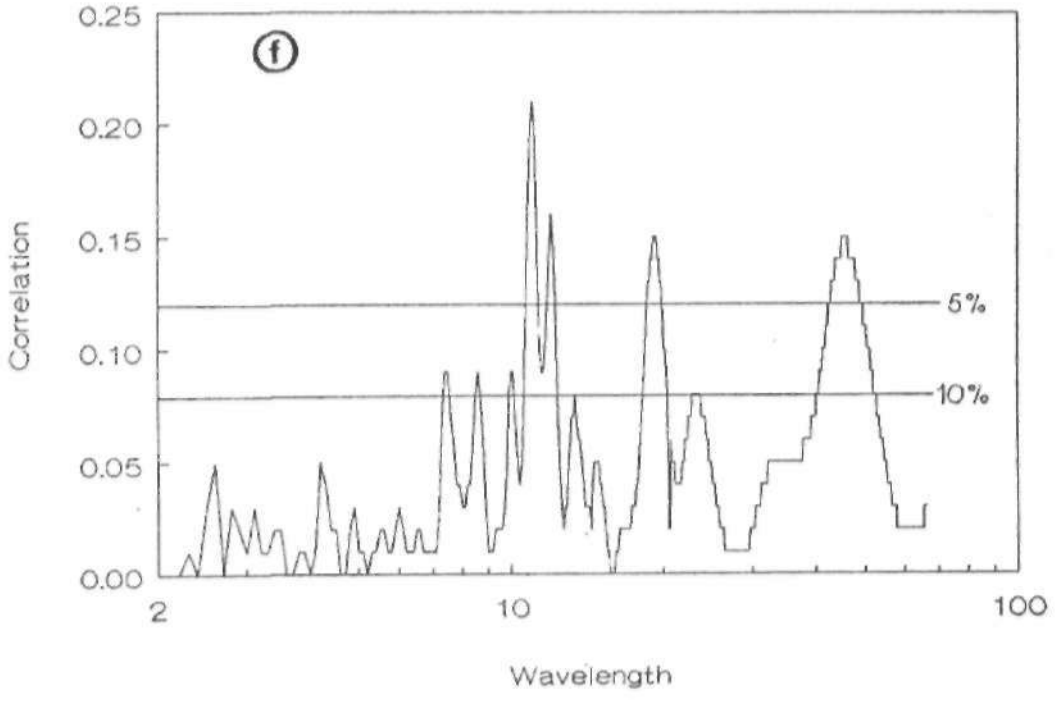
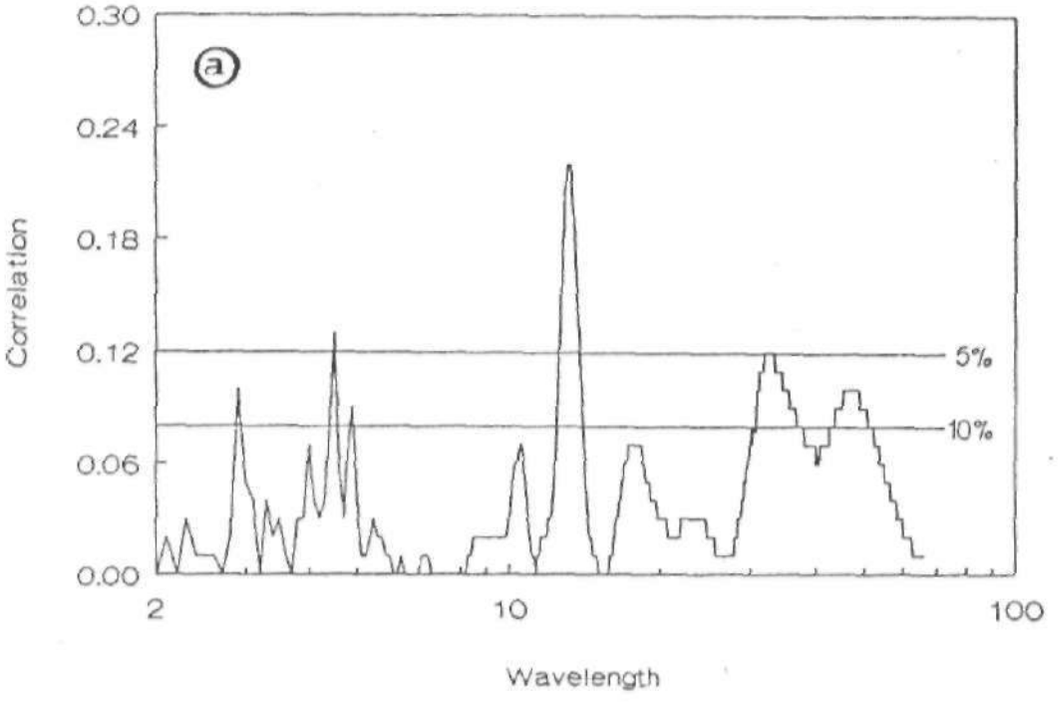


Fig. 1 (cont) for (e) Potiskum
(f) Maiduguri

PLOT OF KADUNA STATION



PLOT OF JOS STATION

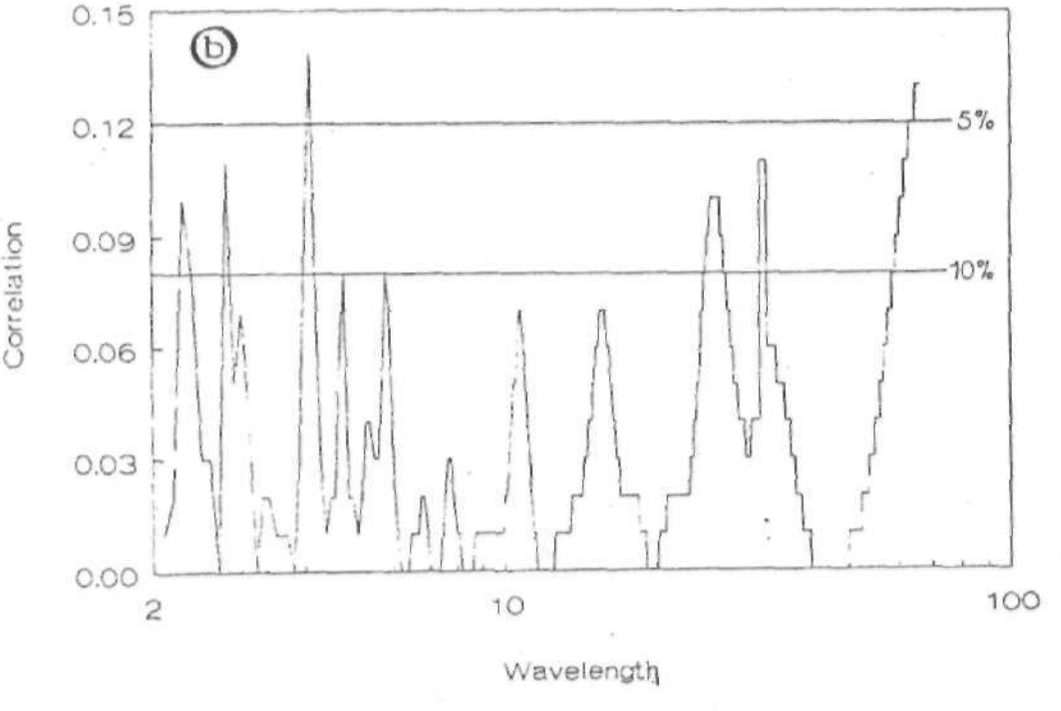
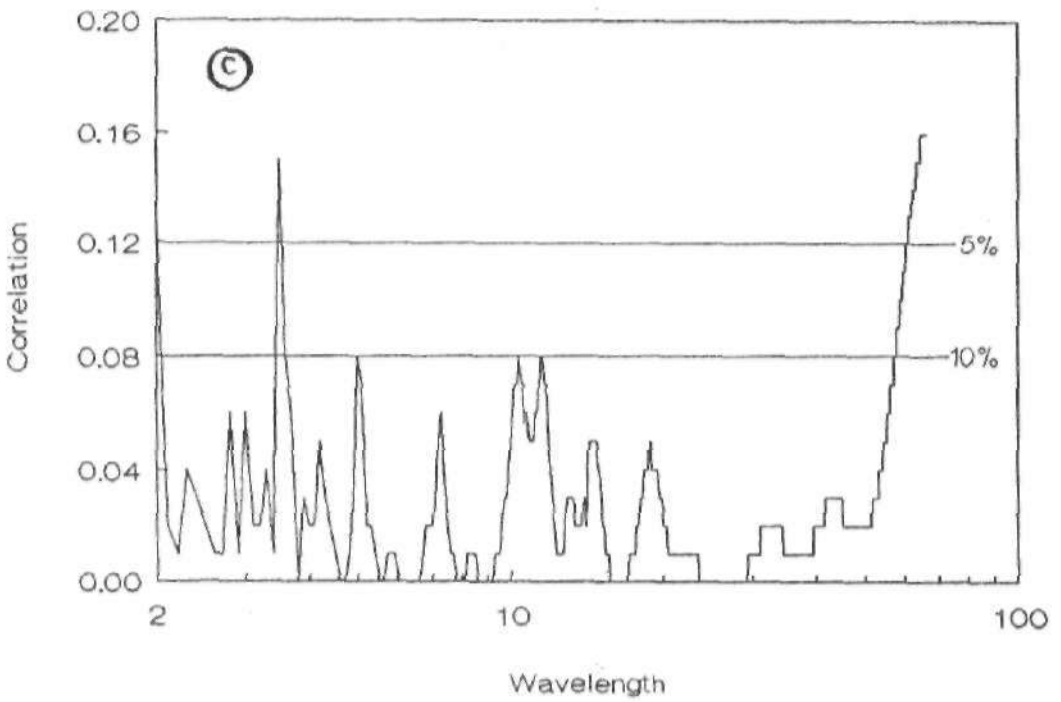


Fig. 2 (a-h). NON-INTEGER SPECTRA OF DROUGHT FOR SELECTED STATIONS IN THE MIDLAND ZONE a) KADUNA b) JOS

PLOT OF BAUCHI STATION



PLOT OF YOLA STATION

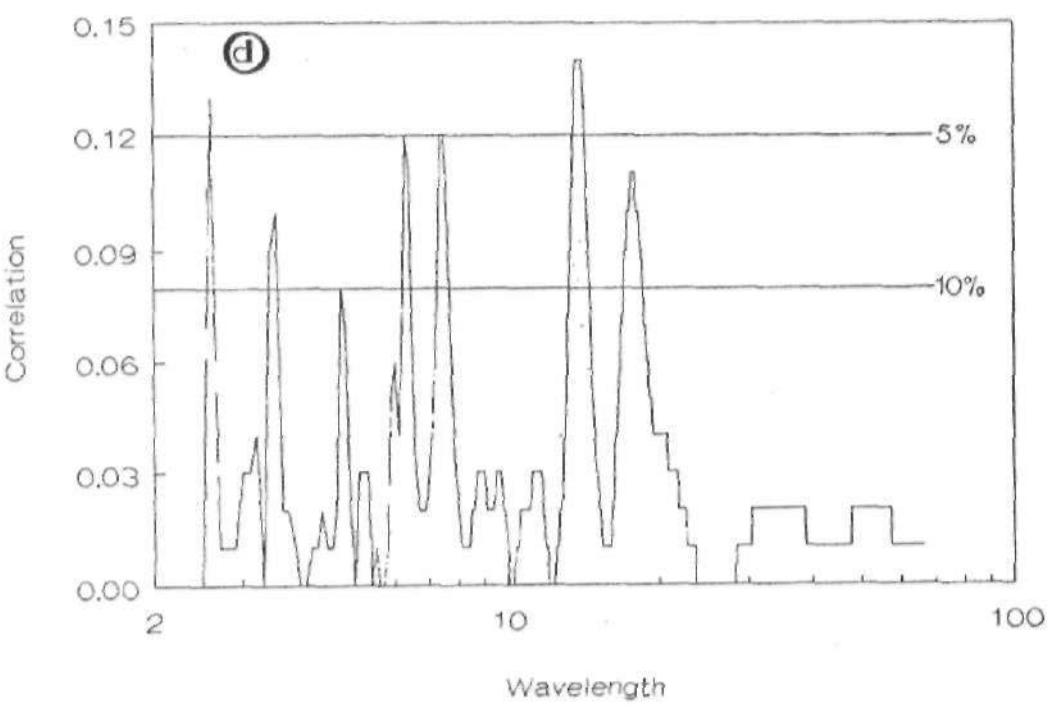
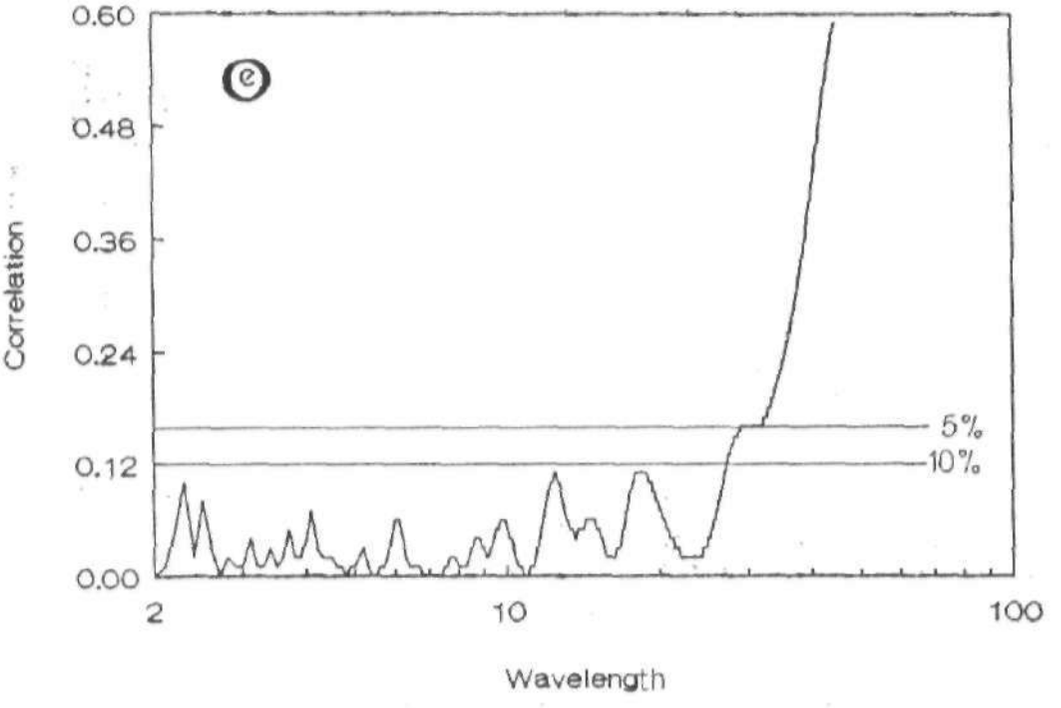


Fig. 2 (cont). for © Bauchi ⓓ Yola

PLOT OF YELWA STATION



PLOT OF ZARIA STATION

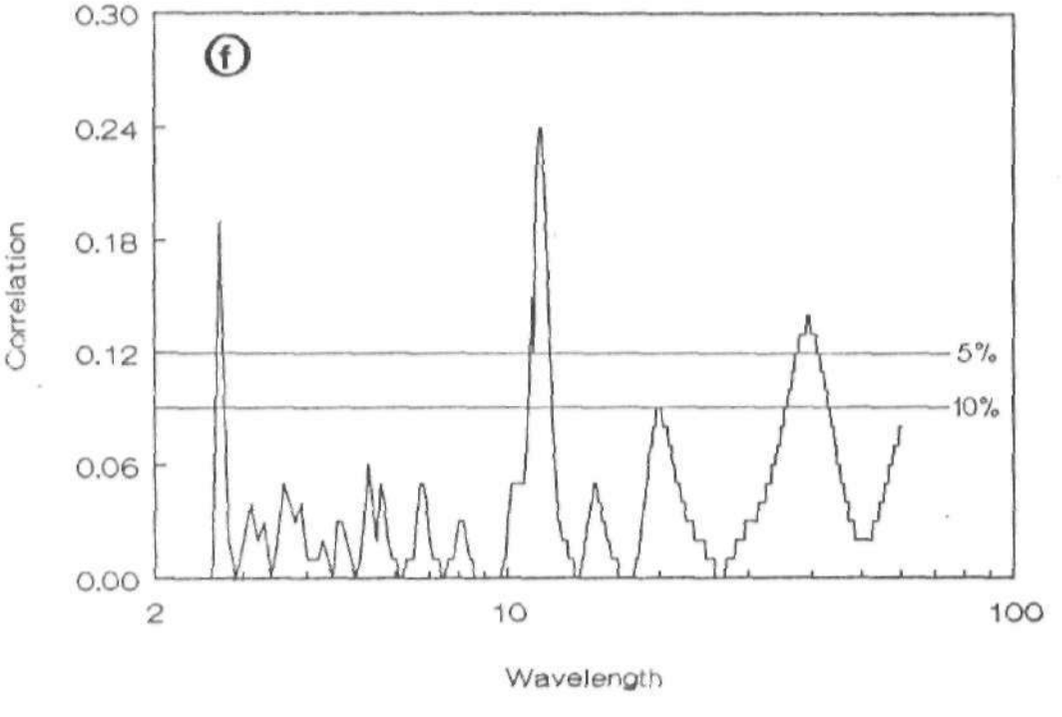
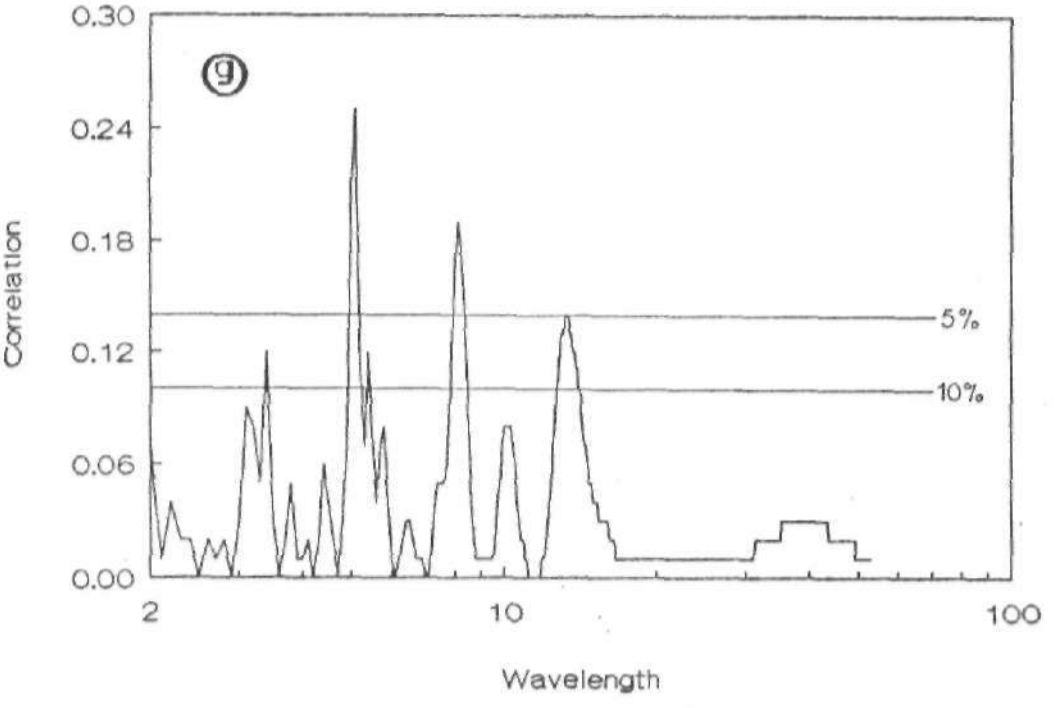


Fig. 2 (cont) for (e) Yelwa (f) Zaria

PLOT OF BIDA STATION



PLOT OF LAFIA STATION

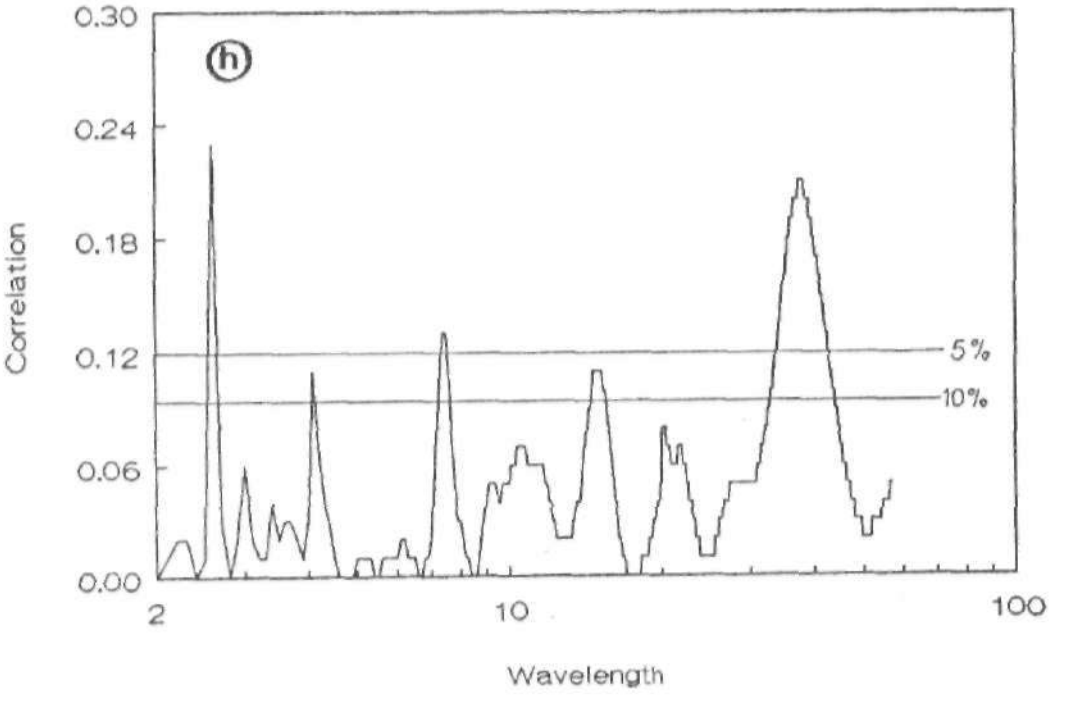
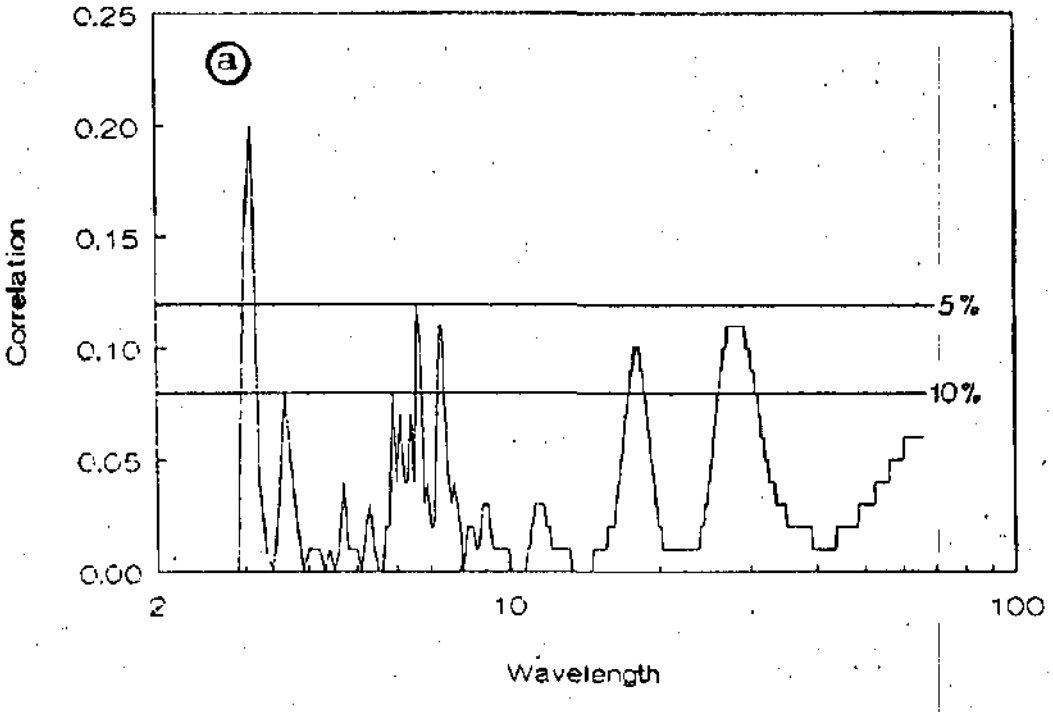


Fig. 2 (cont). for (g) Bida (h) Lafia

PLOT OF IBADAN STATION



PLOT OF OYO STATION

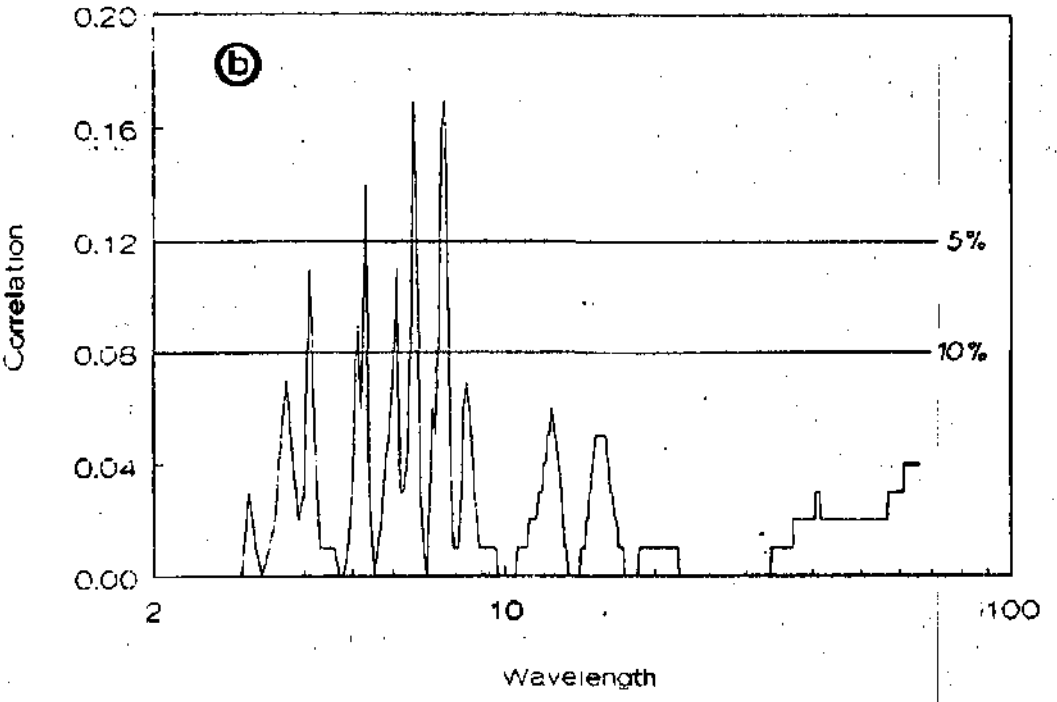
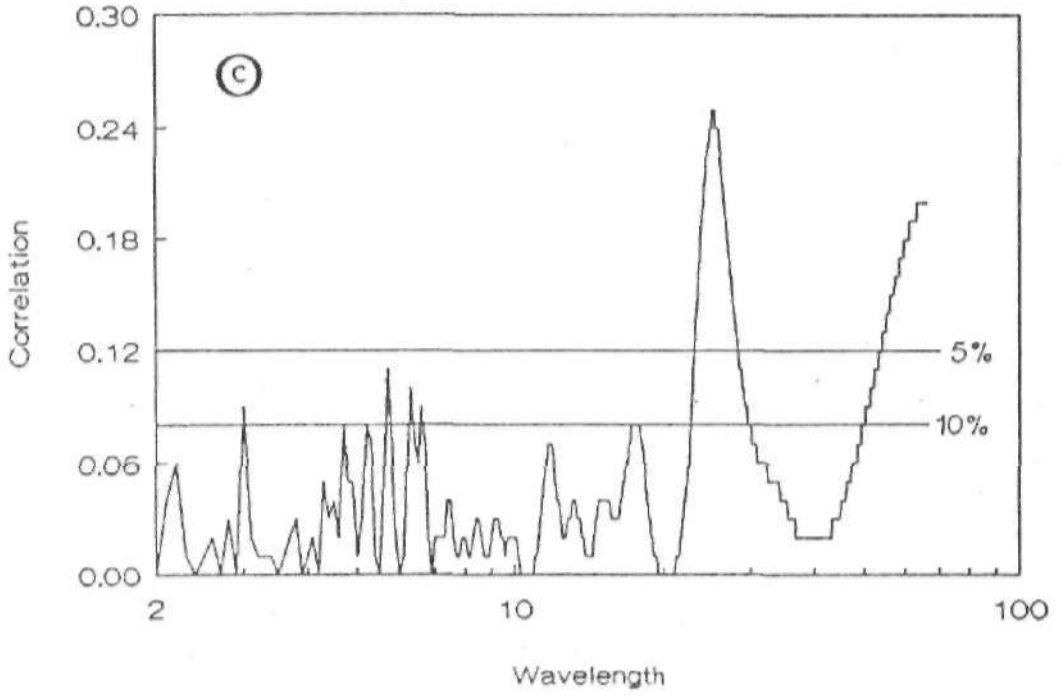


Fig. 3 (a-f). NON-INTEGER SPECTRA OF DROUGHT FOR SELECTED STATIONS IN THE GUINEA-SAVANNA ZONE a) IBADAN b) OYO

PLOT OF OFFA STATION



PLOT OF LOKOJA STATION

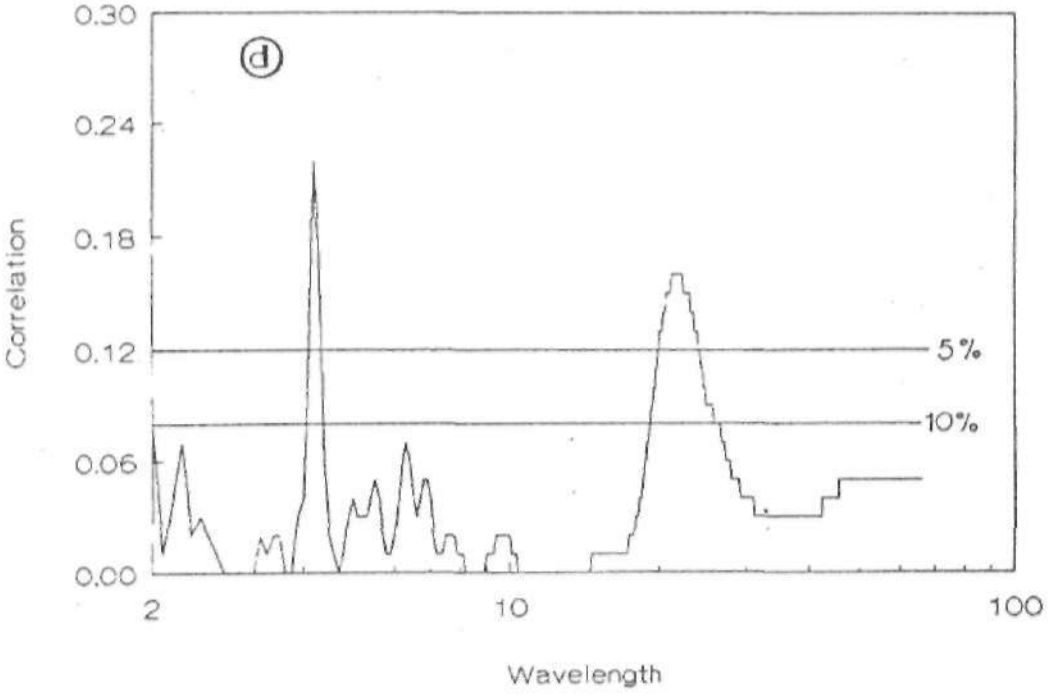
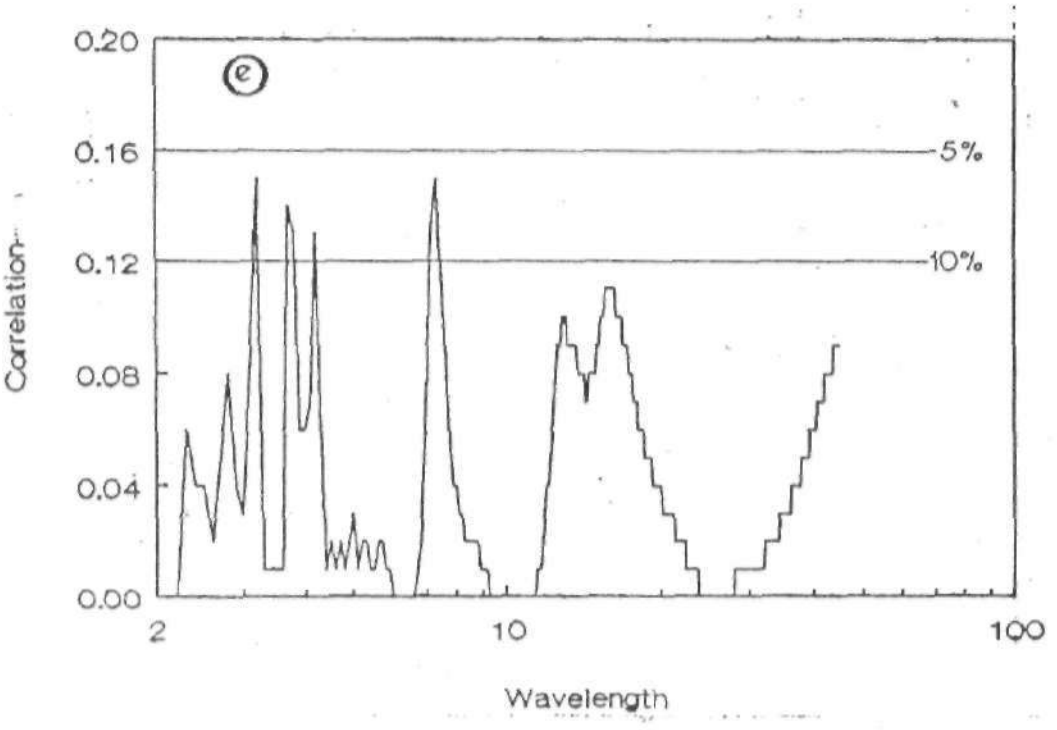


Fig. 3. (cont). for ⓒ Offa ⓓ Lokoja

PLOT OF MAKURDI STATION



PLOT OF YANDEV STATION

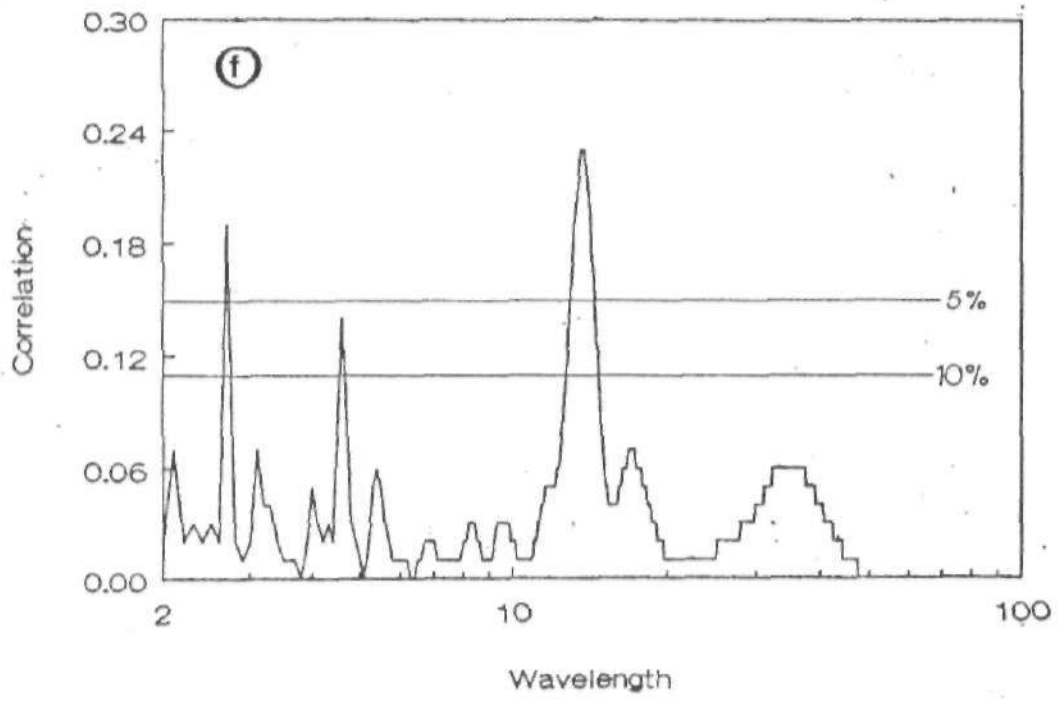
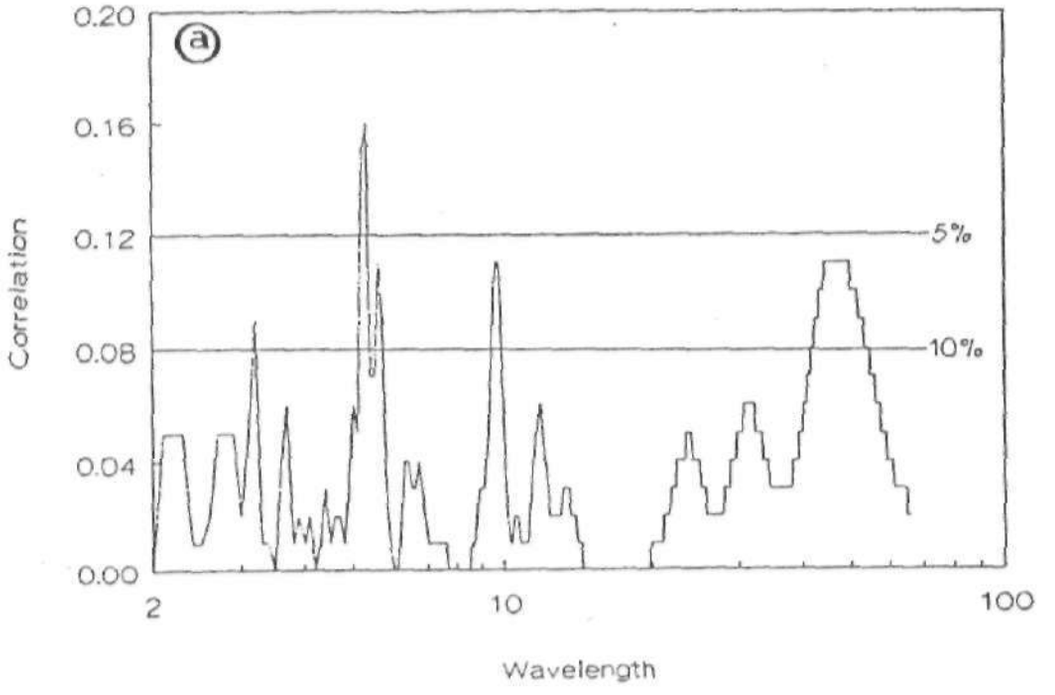


Fig. 3. (cont). (e) Makurdi (f) Yandev

PLOT OF LAGOS STATION



PLOT OF KWALE STATION

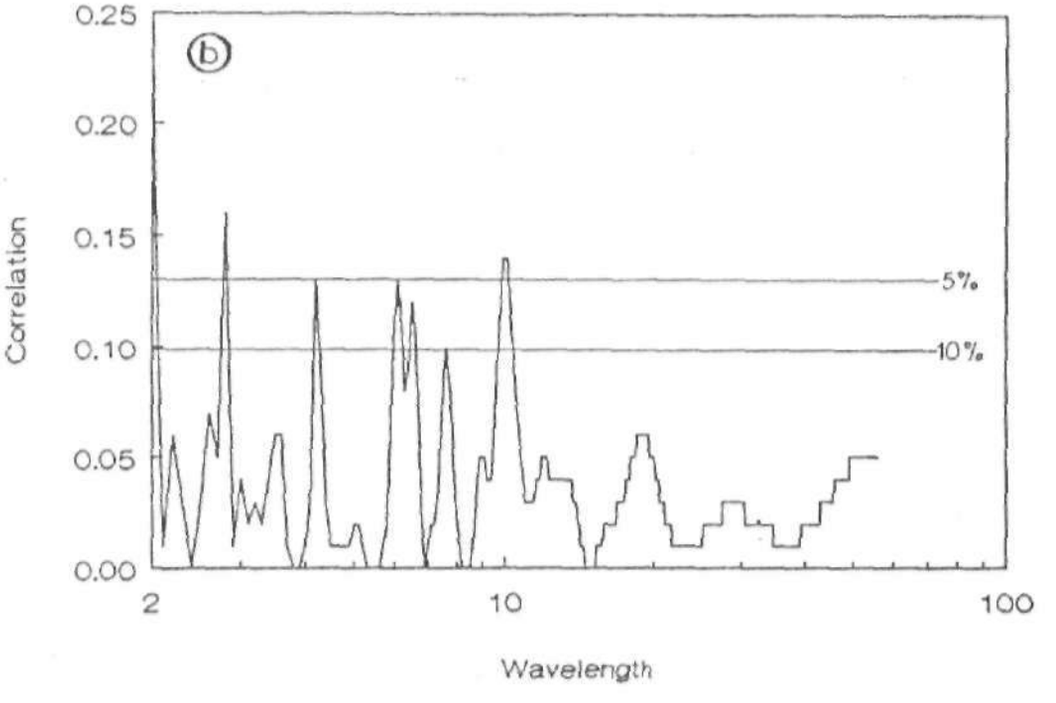
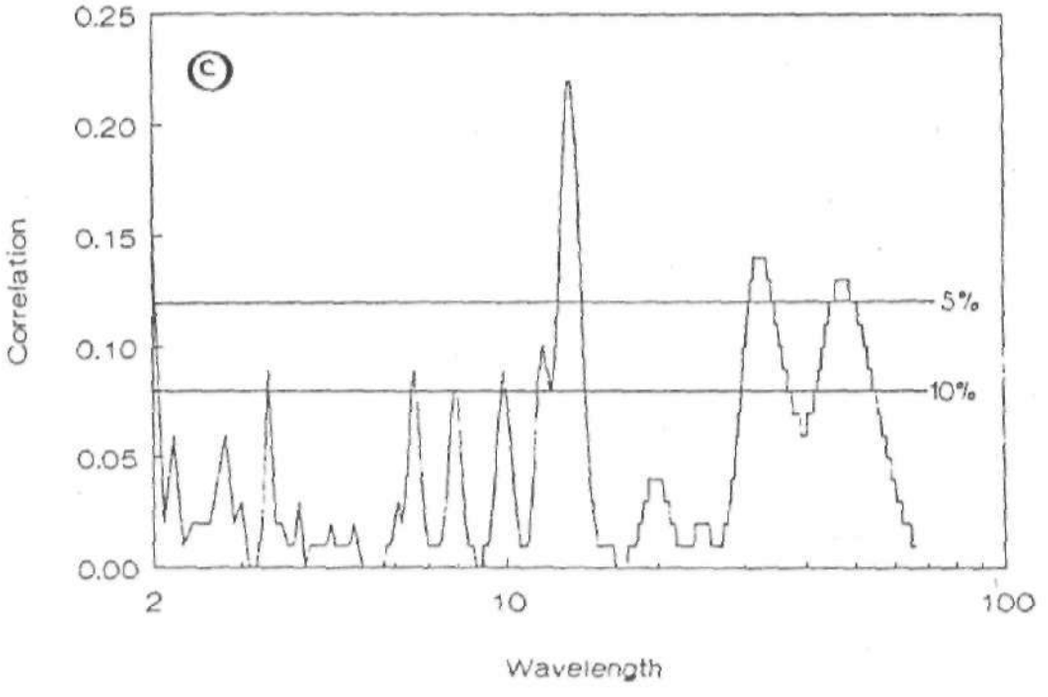


Fig. 4. (a-h). NON-INTEGER SPECTRA OF DROUGHT FOR SELECTED STATIONS IN THE COASTAL ZONE

ⓐ **Lagos** ⓑ **Kwale**

PLOT OF OWERRI STATION



PLOT OF PORT-HARCOURT STATION

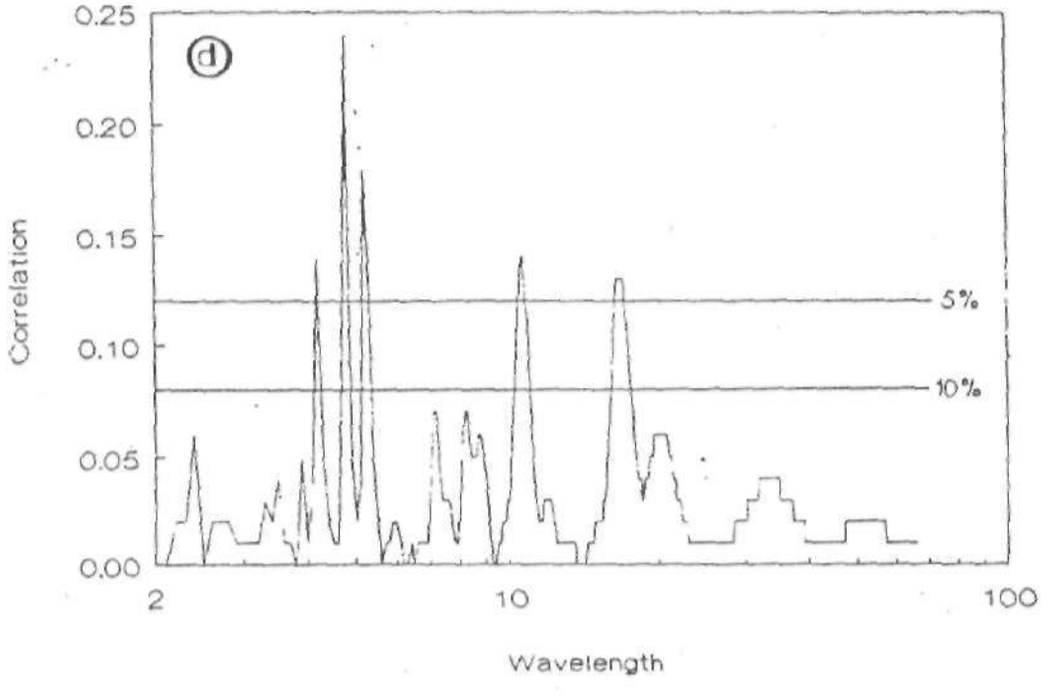
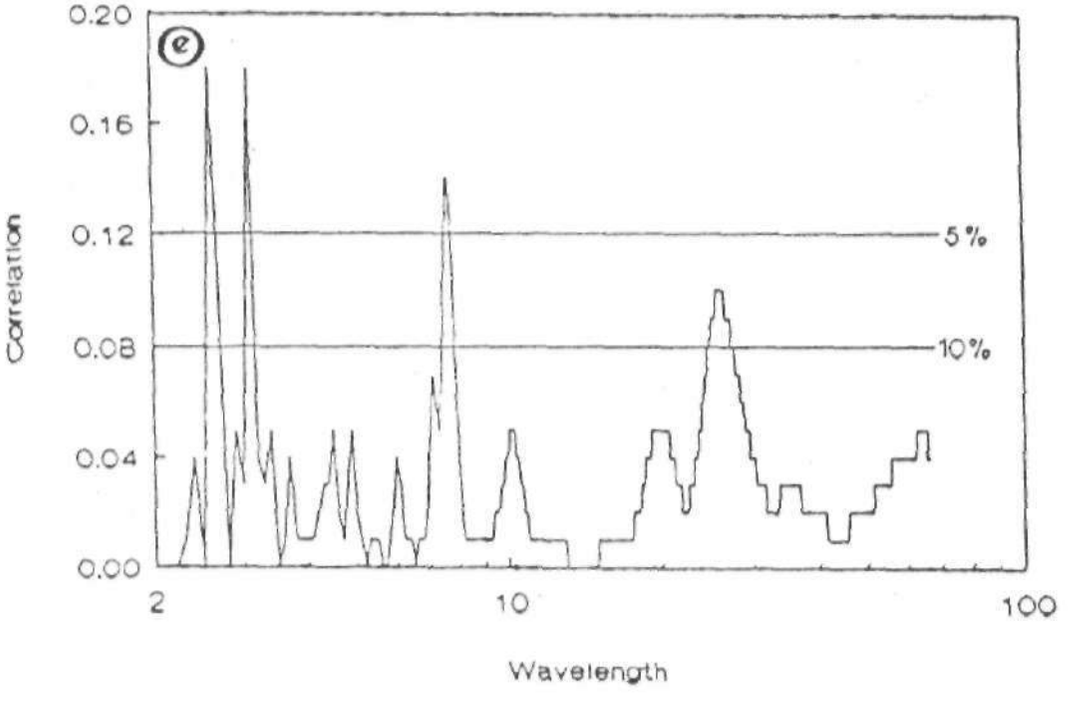


Fig. 4. (cont). for (c) Owerri (d) Port-Harcourt

PLOT OF ENUGU STATION



PLOT OF AGBOR STATION

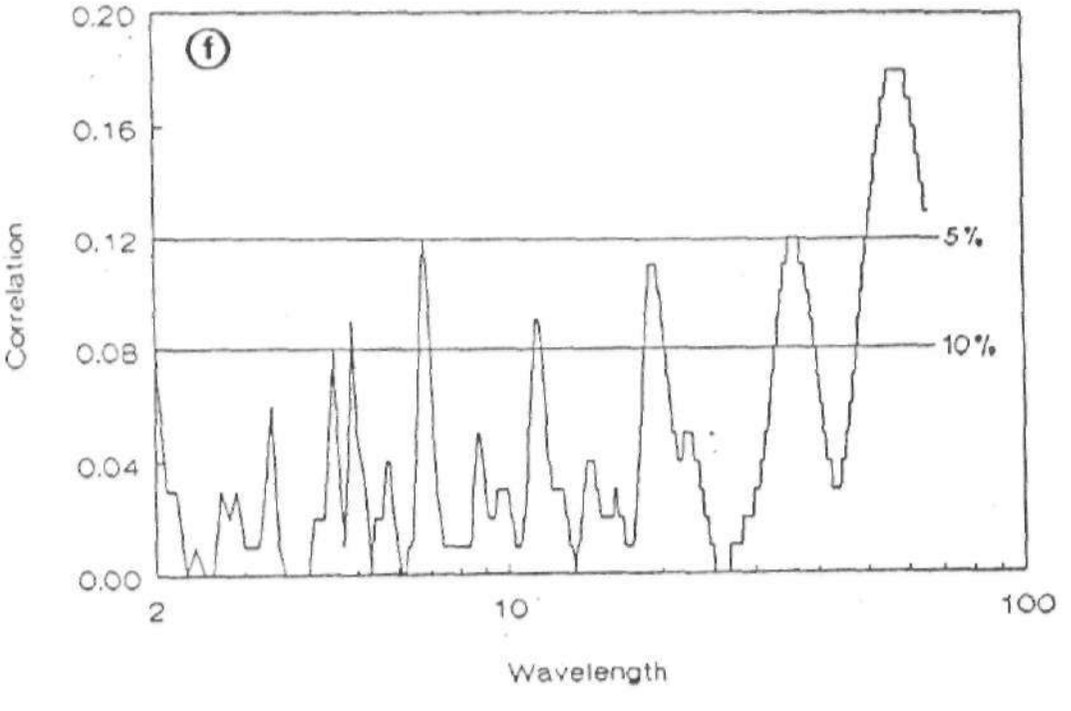
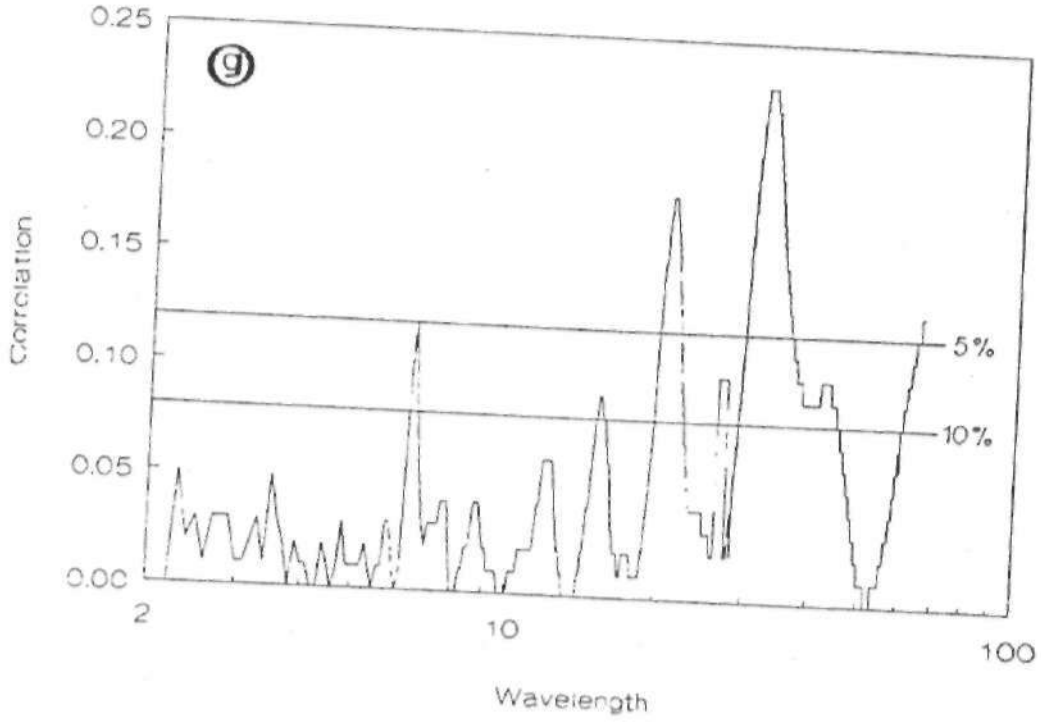


Fig. 4. (cont). for ⓔ Enugu ⓕ Agbor

PLOT OF AFIKPO STATION



PLOT OF CALABAR STATION

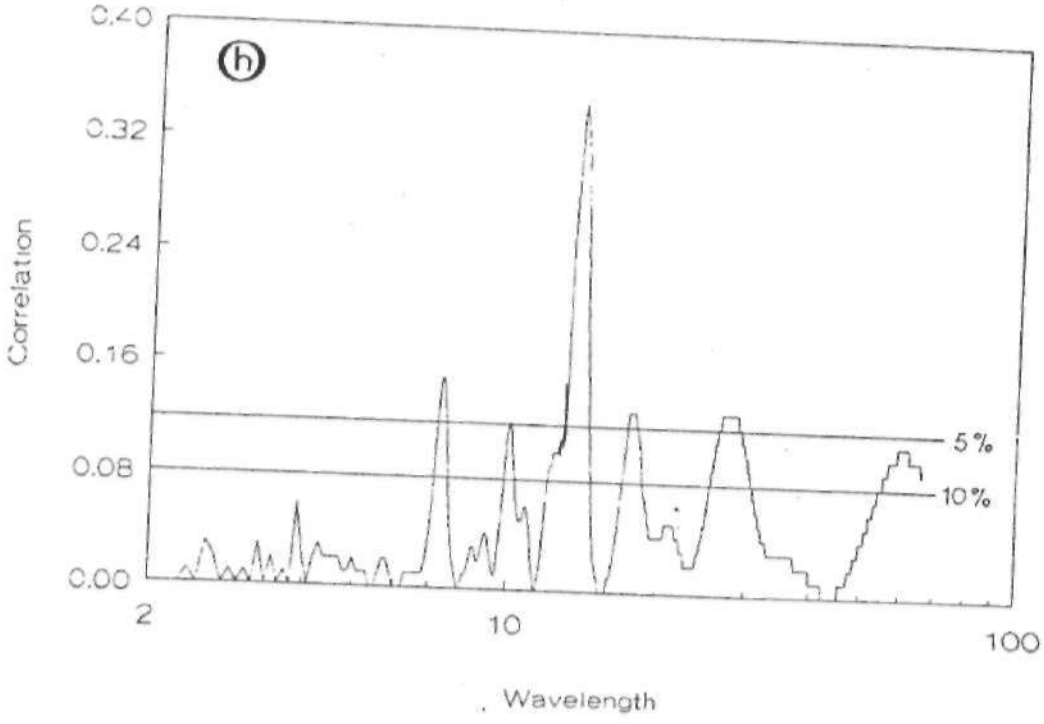


Fig. 4. (cont). for ⓖ Afikpo ⓗ Calabar

APPENDIX D

SIGNIFICANT PERIODICITIES (YEARS) OF MEAN SEASONAL (APRIL-OCTOBER) BHALME AND MOOLEY DROUGHT INDEX IN NIGERIA

*significant at 10% level

NDIX D

	Station	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37	38	39	40	41	42	43	44	45	46	47	48	49	50	51	52	53	54	55	56	57	58	59	60	61	62	63	64	65	66
SAHEL	Sokoto			4.0										14.8																											42.9																												63.9		
	Birnin-Kebbi									10.3			13.4														27.3																																										
	Gusau				5.5						11.2																																44.3																										
	[illegible]ma												13.2															28.7													40.0																									66.0		
	Kano						7.9							14.1																				33.5																																66.0		
	Hadejia	2.0		4.3	5.3	6.3								14.8																		32.0																																66.0		
	Potiskum						7.2														31.7																																		52.0														
	Maiduguri						7.9	8.4	9.9	10.8	11.9		13.5						19.2																										45.6																										
MIDLAND	Yelwa																																								44.9																														
	Zaria										11.7									20.0																			39.3																								59.7						
	Kaduna	2.9		4.9																												32.8																47.2																		
	Minna	2.6	3.4			6.1			9.8				13.0			16.8						22.6																																																
	Bida		3.4		5.4			8.0																																																													
	Lafiagi		3.0	4.3							11.2																																																									
	Jos	2.8		4.8	5.8																					26.2																																								65.5		
	Bauchi		3.5		5.0					10.4	11.6																																																									67.9		
	Pankshin	2.1			5.9									14.1																																																						
	Numan	2.8				6.9																						29.0																																						
	Yola	2.9		4.1	5.7	6.9											17.2											28.0																																						
	Lafia	2.9		4.4			7.8								15.3																										38.0																										
	Gombe			4.2										14.7																																																			
GUINEA-SAVANNA	Ilorin		3.2																																												46.1																								
	Ibadan	2.6		4.9	5.6		7.4											18.1										28.2																																						
	Oyo		3.1	4.1	5.6	6.6																																																														
	Offa	2.9		4.6	5.6	6.5											17.5								25.1																																						64.8		
	Lokoja		3.0		5.2																	22.7																																											
	Makurdi		3.5	4.0			7.2																																																													
	Yandev	2.1	3.0										13.8																																																						
	Ado-Ekiti							8.9									17.5																	30.3																														
COASTAL	Lagos		3.1		5.7				9.9																																						46.9																								
	Ondo				5.7	6.6	7.2				11.3																																																									
	Benin	2.7					7.9						13.9							20.3																																																
	Agbor			4.9		6.8					11.5								19.3																																																
	Kwale		3.1	4.5		6.8	7.9			10.3																																																									
	Auchi	2.1	3.0																																																												
	Owerri		3.4			6.7		8.1		10.0		12.0	13.5																																																						
	Enugu		3.1				7.7																			26.1							32.6																														
	Port-Harcourt			4.6	5.0					10.5						16.7																																																
	Uyo				5.5				9.6		11.0				16.4																																																			
	Afikpo														15.0					20.6													32.8																															65.6		
	Calabar						7.3			10.0			13.5				17.8											23.0																																						