THE EFFECTS OF HUMAN-INDUCED WATERSHED CHANGES ON STREAMFLOWS

By

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A Doctoral Thesis
Submitted in partial fulfilment of the requirements for the award of
Doctor of Philosophy of Loughborough University of Technology

July 1995

Loughborough University of Technology

SYNOPSIS

The aim of the study was to establish the effects of human-induced watershed changes on streamflows. The research hypothesised that land use change influences base flows. Enjoro river in Kenya was used as the case study. In the 1940s, the watershed was characterised with a sparse population, forestry and large scale conservative agriculture. The river regime was naturally perennial. Between 1960 and 1990 land subdivision, intensive cultivation, urbanisation, and deforestation changed such stable ecosystem

Several approaches were used to evaluate the perceived cause-effect relations in the watershed. The time series of the flows, rainfall, and other climatic records were used to infer effects of changed physical characteristics in the watershed. Quantitative evaluation of the changes was accomplished by simple graphs, homogeneity tests, satellite imagery and model simulations of hydrologic variables. Analysis of the data series before and after the presumed changes provided an understanding of the variability masked in the hydrologic system. These comparisons allowed for the determination of the period in which the watershed changes influenced the river regime.

The combined effect of human and natural factors decreased the river base flows. A 30% increase in deforestation, 20% in agriculture and 10.4% in urbanisation was observed· Water availability decreased from a runoff coefficient of 22% in the 1960s, 10% in the 1970s and 8% in the 1980s. This progressive decline in runoff developed into hydrologic drought regime in the 1980s. Normalized difference vegetation index (NDVI) predicted well the flow changes in the watershed. Simulations of rainfall and flow supported the changes observed in the hydrologic variables. The optimised parameters with HYRROM showed 'store' parameters (SS, RDEL, GDEL) to be sensitive to changes in vegetation cover especially during the dry years of 1965, 1973 and 1984. The model simulated some parameters in the watershed which could be used to infer changes in streamflows due changes in land use. It was however, difficult to estimate and to validate long-term model parameters because of limited data and the contrasting geography of the region which induced hydrologic variability. The model did not isolate effects of specific land uses, although it predicted the observed flows.

There is evidently, a need for future research on the problem. The investigation demonstrated the difficulty in identifying differences in streamflows from watersheds undergoing simultaneous physical changes and human interventions. Since a specific effect of a particular land use change could not be isolated independently, continued research on the development of an integrating watershed coefficient is recommended. Remote sensing techniques should be incorporated in the development of integrating watershed coefficients.

ACKNOWLEDGEMENTS

The success of this work was due to a combined effort of my supervisors and the Director of the Research. Distinguished appreciation, and gratitude must go to the main supervisor, Mr. I.K. Smout who constructively criticised the work and gave me invaluable help and assistance. His gentle remarks contained an immense amount of professional knowledge, experience and skill. My special thanks also go to Dr. A.P. Cotton, the second supervisor, who vitalised the research methodology and provided research materials.

In the department and particularly WEDC, I owe the support staff a lot of thanks for the provision of photocopying and computer facilities during the entire period of the research. In particular, Patricia Jackson in WEDC Resources Centre who tirelessly and patiently searched for all materials and journals.

The financial support from Egerton University and the Government of Kenya through the Ministry of Education is gratefully acknowledged. Special thanks to Mr. P.H. Dido, the Education Attaché' in the Kenya High Commission, London. He was very helpful in securing funds from Nairobi and supported my stay so ably.

The members of staff at the Kenya Meteorological Department in Nairobi who patiently worked with me through volumes of files from their archives to extract climatic data used in this study. To also the staff at the Regional Centre of Remote Sensing, Surveys and Mapping in Nairobi for their kind support in the extraction of the NDVI data from the NOAA-AVHRR, GAC data bases. To the people at the Department of Resource Surveys and Remote Sensing of the Ministry of National Planning, Nairobi for their support in the provision of equipment for tracing and extracting information from the Land SAT, topographical maps and SPOT satellite imageries and finally to the staff at the Ministry of Water Development in Nairobi for the provision of river flow and rainfall data. In Nakuru, my sincere thanks go to the members of staff of the Lake Nakuru Conservation and Development project and the Senior Warden, Mr. Kisese for their support and permission to work in the National Park.

To my wife Olpha, and my children Kiptogoch, Jebiwott and Jeptum, my thanks for your patience could never be sufficient. I salute my brother Ezekiel who kindled my interest in academic excellence since my youth.

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CHAPTER I

INTRODUCTION

1.1. BACKGROUND TO THE RESEARCH

Research into the effects of watershed changes on streamflow is an area currently under urgent global attention. Agenda 21 (UNCED, 1992) recommendations and principal objectives in Chapters 10 and 18 emphasized the application of integrated approaches in the development and management of land and water resources. In chapter 31, the scientific and technological community are urged to make a more open and effective contribution to the decision-making processes concerning the environment and development. This global attention came about because land and water resources are the two very important resources whose drastic depletion is causing the greatest effect on a continued survival of humanity. A specific action of immediate concern is how the existing fresh water reserves could be conserved, managed, and developed in a manner that will ensure its continuous use and sustainability. There is therefore a need for a better understanding of land and hydrological processes to make a better use of scarce water resources now and in the future. Consequently this research attempts to contribute to this common goal.

Macro-scale studies have been carried out into the effects of human activities on hydrological regimes. There is however, a need to establish the actual trends and magnitudes of the changes and to quantify the cause-effect relation of man and his environment at small watershed level. It is against this background that this study was conceived. A single rural watershed in Kenya was chosen as the case study. The watershed for years underwent tremendous ecological and environmental changes which have implicitly influenced its hydrological regime to some extent.

Watersheds are extremely complex systems and hence any disturbance therein is bound to have certain resultant adverse effect. These effects are not instantly recognized by the inhabitants of the watershed, instead, it takes several years and even decades to be observed and felt. The term watershed in this case is used interchangeably with catchment to denote all those areas enclosed by a continuous hydrologic surface drainage divide and draining into a single natural or artificial body of water. Within this system are found natural ecosystems and human activities that tend to alter and modify it.

Moreover, it is generally accepted that modifications of these hydrologic systems have been going on throughout several past centuries. The author's impression however, is that the rate of modification has of recent times been drastic and its extent and magnitude especially in small rural and peri-urban watersheds have not been adequately addressed and quantified. Besides, watersheds being the natural sources of surface and ground water to a multitude of flora and fauna, there should be concerted efforts to minimize these human-induced changes that would ultimately alter the quantity, quality and the uniform distribution of the resource.

Many studies carried out on the subject have tended to be regional and global in nature. The results from these studies have not adequately depicted the actuality in small rural watersheds under intensive human activity as a result of increased human population. Therefore, the issue under urgent concern in tropical Africa, and East Africa, in particular, is how watershed managers could deal with such dynamic human-induced changes. Watershed researchers and hydrologists are thus required to narrow their studies to answer such questions as: whether land use changes really cause any quantifiable changes in streamflow?; how significant are the changes in terms of extent and magnitudes? and which patterns produce the major effects?

It is difficult to prescribe wholesome and holistic solutions to these questions, especially as the analysis is often based on imperfect observations in a complex and sometimes discontinuous domain. Therefore an integrated approach may be the only approach that will enable land use planners and managers to direct their energies and resources to specific problems at hand.

What further complicates the matter, is that human activities are known to have influenced hydrological characteristics for thousands of years and in many parts of the world. Thus what are popularly thought of as 'natural environments' are infact the end result of generations of land uses (Arnell, 1989). These interventions have also had impact on water quantity and quality at varying levels. A study is required that distinguishes different types of impact on the basis of which part of the hydrologic system they affect and how they arise.

In addition, not all human influences on hydrological regimes are through land use changes. The generation of hydropower, supply of water for domestic, industrial and agricultural development and river channel modification also lead to hydrological changes. A study of this nature therefore, has to deal with the 'deliberate' and inadvertent human impacts on the hydrological regime. It will also be useful to distinguish different orders of impact. Immediate or first-order impacts of an activity may be the most obvious but these effects may trigger other changes resulting in other conditions which may further complicate the question under study (Arnell, 1989).

1.2. STATEMENT OF THE PROBLEM

Previous studies on small rural watersheds in Kenya have often been descriptive in character. Most reports provide a rather general and abstract analysis of hydrometeorological variables without detailed examination of their cause-effect relations. Limited attention has been given to temporal and spatial analysis that isolates the sequence of events contributing to the current scenerios. The studies further fail to pinpoint clearly where human impact is more profoundly felt or the hard hit spots which require immediate and appropriate remedial action. With out such studies coupled with a well planned engineering design input, the problem of extreme hydrological changes occurring due to human interventions may never be identified and addressed adequately.

Further, human activities are known to have enhanced natural land degradation and even enhanced micro-climatic changes in tropical Africa (Hamilton, 1985, Suliman, 1990, Smout et al., 1993). The climatic changes in the continent have recently been established as accelerating at alarming rates away from its historical patterns. The 1984 devastating drought in Africa, for example, is a clear proof of the changes through misuse and uncontrolled user rate of its natural resources. Findings by Farmer and Wigley (1985), Jackson, (1988), and Parry, (1992) have shown that in the last four decades, drought of even minute magnitudes could not be absorbed and sustained any longer by the newly created fragile ecosystem.The present ecological system in the continent has even affected its agricultural production patterns. In addition, these changes are no longer limited to discrete ecosystems or African tropical regions. Rather, they have become global in nature (Ogendo, 1990) and a global attention focused in the 1991 Rio Earth Summit confirms the importance of the problem to the world community.

Other than the natural phenomena, the majority of the changes can at the moment, be considered anthropogenic in character (human-induced). The anthropogenic ones include the progressive encroachment on the watershed in the form of heavy human settlements, intensive cultivation, overgrazing, wanton destruction of woodlands for timber, fuelwood and charcoal and an enormous pressure on land and water resources. These are difficult to control because the abject poverty which characterizes most developing economies is a major constraint to the achievement of sustainable development. In particular, Kenya views underdevelopment and poverty as the greatest threat to the sustainability of the environment (Moi, 1992). Therefore, the examination of the causes of poverty at the watershed level is important in studying the causes of anthropogenically induced hydrologic changes. This research however, is more concerned with the physical processes.

The activities are further accelerated by the ever increasing human population in Kenya which depends solely on subsistence farming. Furthermore, the indigenous land production systems coupled with widespread stagnation in agricultural technology in most semi-arid areas of Kenya has created disequilibrium between patterns of resource use and availability.

The other human activities which modify hydrological regimes are the rapid growth of industries and urbanisation. As human population migrates from rural to urban areas, there is increased occupancy and use resulting in increased impervious areas and overloaded drainage systems, which modify the urban hydrological and natural environment. Particular physiographic conditions brought by these changes are: 1) the increased impervious areas which affects strongly the flood production levels while 2) infiltration is reduced 3) decreasing the baseflow of the natural streams and 4) reducing the time of concentration.

These events are observed to also cause a change in water quality and quantity. Sediment transport is increased downstream and sewage and factory effluents discharge into the natural bodies of water. These and other human interventions damage land and water resources which the population needs in abundant supply, in good quality and in uniform distribution.

The problem therefore calls for consideration, examination and evaluation of the changing trends and magnitudes of hydrological variables. This will enable the development of strategies and guidelines for total watershed planning and rehabilitation of the watersheds so they may act naturally as shock absorbers to these dynamic human interventions. This study therefore, concentrates its efforts on the evaluation and assessment of the magnitudes and extent of both hydrologic and land use changes and proposes recommendations that will render the small rural watersheds self sustaining and ecologically sustainable.

1.3. OBJECTIVES AND HYPOTHESES

1.3.1. The Hypotheses

As a basis of the research, the foregoing discussion on the nature and scope of the problem under investigation led to the proposal, formulation and testing of four (4) condensed hypotheses. The first deals with the relationship between changing land use and streamflow; the second is on the occurrence of hydrologic droughts (low flows) resulting from changes in rainfall characteristics in the region and particularly in the case study watershed.

The third hypothesis tests the effects of the changing watershed vegetation cover on streamflow characteristics. It tests the statement that large scale vegetation changes (for example, due to increased deforestation, agriculture, population density, and urbanisation) that tend to modify the rivers' flow regime can be established from a remotely sensed data.

Hypothesis number four examines a possible adaptation of existing conceptual and lumped hydrologic models to simulate (optimize) hydrological processes (parameters) that will allow an evaluation of the effects of different land uses on streamflows.

Hypothesis 1: The consensus scientific opinion that land use patterns modify streamflow characteristics can be shown to be true for small rural watersheds in Kenya.

Hypothesis 2: It is possible to analyse limited data records from changing watersheds, to identify the contributions of the regional rainfall regime and of human-induced factors and observe changes in streamflows.

Hypothesis 3: The normalized difference vegetation index (NDVI) derived from satellite imagery can used to monitor and predict streamflow responses to changes in vegetation cover in ungauged watersheds.

Hypothesis 4: That even with limited and discontinuous hydrologic data, existing conceptual hydrologic models could be calibrated and optimized to simulate the hydrological processes that will allow an evaluation of the effects of different land uses on streamflows.

1.3.2. The Objectives

The general aim of this study was stimulated by the concern that human-induced watershed changes could have adversely affected the amounts and distribution of land and water resources in Kenya. Therefore, to test the above hypotheses, Enjoro River watershed in Kenya was chosen as the case study. The watershed was particulalrly selected because its surface runoff had reduced over the years and affected the water balance and hence the survival of Lake Nakuru (Kimani et. at. 1992). The study thus focused on the detection and quantification of the hydrological impacts of land use changes by studying historical hydrologic records. It is hoped that results from the study will assist land use planners and engineers to select and design practices that are environmentally and ecologically sustainable. Consequently the specific objectives of the study are:

1. To achieve from the literature review a detailed understanding of the hydrological processes operating in rural watersheds to provide an adequate basis for examining and predicting the consequences of human-induced land use changes.

2. To develop a framework for analysing the relationship between land use and streamflow in rural and peri-urban watersheds by collecting hdrological data, conducting ground-truth field surveys and studying satellite imageries.

3. To identify the temporal changes in hydrological response characteristics of the chosen watershed that may be attributed to human-induced changes in land use (agriculture, deforestry and urbanisation).

4. To identify for the chosen watershed the spatial and temporal rainfall characteristics that may have contributed to the changes observed in the streamflow regime.

5. To investigate the possible selection and adaptation of existing conceptual models for simulating the hydrologic processes that will enable an evaluation of the effects of different land uses on streamflows.

1.4. APPROACH AND SCOPE OF THE THESIS

Chapter I introduces the subject matter, provides the backgound information into the aim of the study, and explores the scope of the problem through a series of objectives and four hypotheses. The chapter concludes with a brief summary of the Thesis in Figure 1.1 and the rest of the content is introduced in chapter II.

Chapter II reviews literature on the subject matter and investigates the extent to which influences of human activities on hydrological regimes have been studied before. It concludes by restricting the area of the current research.

Human interventions in the hydrological cycle are extensively covered in Chapter III. The Chapter examines the nature, elements and composition of a human-modified hydrologic system. It discusses the system inputs and outputs, and relates to the principal areas of human modifications and how the systems approach can be used in studying the effects. The chapter further explores relevant analytical methods in the study and suggests areas requiring adjustments.
The methodology, data collection and validation procedures are dealt with in Chapter IV which provides a detailed description of the case watershed in terms of its location, geology and history of the land tenure system. It explains the setting up of the data collection programme in the field, retrieval of aerial and topographic maps, use of land

SAT and SPOT satellite imageries, use of Normal difference vegetation indices (NDVI), and screening of the data. Data quality control and checks were carried out with the aim of eliminating other causes of streamflow changes other than the land use.

Data analysis, results and discussions are covered in chapters V, VI, and VII. These chapters analyze: precipitation, streamflow and hydrometeorological data; NDVI, land use patterns and demographic changes. Chapter VII used the modified version of the HYRROM model to optimize the watershed parameters which were then used to infer changes in streamflow patterns. Suitable parameters related to different land uses in selected periods were identified and optimized for streamflow regimes during these periods.

Chapter VIII discusses the overall results of the study with respect to the stated specific hypotheses and objectives and summarizes the results, conclusions and recommendations. It reviews and makes observations regarding the adequacy and depth of the study and draws recommendations for immediate action and finally cites areas requiring short and long term further research. The Thesis report ends with a list of references, bibliography and appendices.

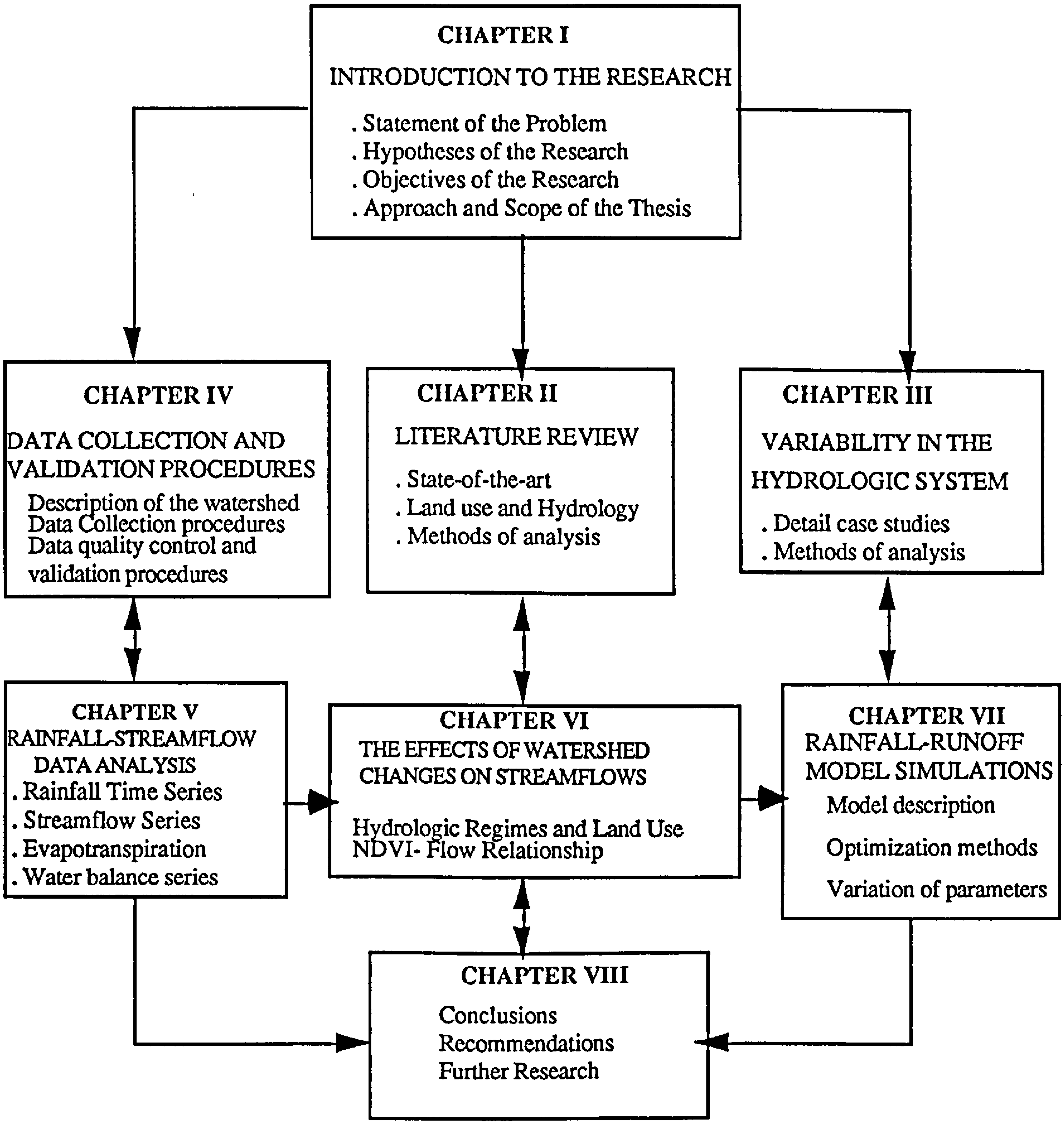


Figure 1.1. The approach and scope of the Thesis

CHAPTER II

LITERATURE REVIEW

2.1 BACKGROUND TO THE SUBJECT MATTER

This chapter consists of a literature review of the effects of human activities (anthropogenic) on hydrological regimes, with particular emphasis on agricultural and forestry development, urbanisation and hydraulic structures. The effects are examined specifically at the watershed level. A watershed is a term used interchangeably with drainage basin and catchment area to denote all areas enclosed by a continuous hydrologic surface-drainage divide and draining into a body of water. The only interactions, are the migration of mobile biotic populations and some ground-water transfers (Petts, 1984). Watersheds are also the natural sources of surface and ground water supplies. Within it are found natural ecosystems and human activities that tend to alter the quantity and quality of water in the respective bodies of water.

The constantly increasing demand for wholesome water has forced engineers and planners to propose and contemplate complex and ambitious plans for water resource systems planning (Loucks et. al., 1981). Such plans include integrated and multifaceted watershed development and management so as to maintain consistent river flows. These ambitious plans were recognised by the World Meteorological Organisation (WMO), and World Health Organisation (WHO) by establishing the Global Environmental Monitoring System (GEMS) whose mandate was to create data banks on the world's river flows and quality.

Developed countries in the North realised this insidious threat early enough, probably because of the impacts of industrial revolution in the 19th century. In Europe for example, United Kingdom, Finland, and Germany research on the effects of deforestation and urbanisation was carried out during the first half of this century but quantative analyses were not common until the 1970s (Meinzer., 1942). Clearing of tropical forests and a rapid change in land use in most African nations reduced the forest area by more than 50% in less than three decades.

The rate of human population growth in Africa is linked with increased deforestation in the continent. Population rose from 438.79 million in 1980 to 592.13 million people in 1990; equivalent to an annual growth rate of 3.2% (World Bank, 1991). The rate is even higher in individual countries. During the same period, Kenya had an annual growth rate of 3.8%, Tanzania, 3.0% and Uganda 3.4%. Since 1975, Kenya had the highest growth rates in Africa of between 3.5 and 4.1%. Most African nations had increased growth rates especially immediately after independence in 1960s. The consequences were increased land fragmentation, increased human migration to urban areas, increased human settlements and continuous encroachment of population to the fragile arid and semi arid lands. Alongside these activities, livestock overgrazing, intensive cultivation coupled with unpredictable tropical weather patterns, complicated the development and conservation of watersheds in the continent.

Human activities of concern that are influencing hydrological regimes in tropical Africa are deforestation, urbanisation, intensive agriculture, and natural resource development. Deforestation is a resultant effect of the need of timber for construction, fuelwood and creating space for settling increasing human population. Urbanisation is attributed to the movement of the population from the rural to urban areas, industrial development and the need to develop services to meet the changing living standards of the urban population, and poverty-driven rural-urban out migration.

By the late 1970s, the Kenya Government however, recognised this urgent need for the conservation of the environment. As a result an elaborate conservation strategy was emphasised through economic, social and political measures that ensures reasonable trade-off between resource use and the deteriorating impact on the environment (GOK, 1979, 1986). These measures included the initiation of intensive soil conservation and afforestation programmes, water conservation and flood control, and the establishment of the Rural Afforestation and Extension Service (RAES). These measures were planned to ensure a balanced ecosystem and thus provide a continuous supply of forest products to meet the 72 per cent of the national energy requirements currently derived from fuelwood.

The Kenyan overall picture depicts a rather low level of forest depletion (of 1%) per annum (GOK, 1989). This rate is much higher in small watersheds in the high and medium agricultural potential areas of the country (most of which are the main catchments to the national drainage system). The hydrological effects of these changes are felt faster in these important agro-ecological zones; the main source areas of food and fuel energy in the country.

LandSAT imagery survey data for 1976/79 period (GOK, 1989) showed Kenya had a forest cover of about 3.4 per cent of the total land mass. These forests are located in the arable land which is presently undergoing unabated intensive cultivation and encroachment into the forests. Further, the land tenure system and cultural norms coupled with increasing population, led to an intensive subdivision and fragmentation of the arable areas into small uneconomic and ecologically unstable units. The population pressure in the high and medium potential areas comprising of 18 per cent of total area (of 582000 sq.km) induced uncontrollable out-migration into the arid and semi arid fragile environments.

The movement into these fragile ecosystems resulted in wanton clearing of vegetation cover, exposing and accelerating soil to wind and water erosion. Overstocking accompanied these activities leading to serious land degradation (GOK: 1980, 1984, 1987, 1989). The extent and magnitude of these activities are yet to be established; a challenge to policy makers, engineers, environmentalists and hydrologists in the country. Thus, there is a need to continuously monitor these human-induced phenomena because the rate at which the hydrologic changes occur is also rapidly changing and may assume geometric rates in the next century.

Agricultural development on the other hand continues to expand in Kenya through increased hectarage, intensive cultivation, multi-cropping systems and mechanisation. To increase production per unit hectare, use of agricultural chemicals was intensified (GOK/MOA, 1989). The steady increase in the use of these chemicals especially in the coffee, tea and wheat plantations increased considerably, the pollution rate of water sources in major watersheds. Again, the magnitudes, trends and distributions of these effects are not yet established. Studies by the Tea Research Foundation of Kenya and reported by Othieno (1979) for example showed increased nutrient loads on the Sambret River due to agro-chemical inputs from the Kericho tea plantations.

2.2 CONTEXT AND PERSPECTIVES

2.2.1 The Historical Perspective

As early as the 1930s, there was a widely held view that forests were essential to maintaining streamflow (Nicholson, 1936). A considerable debate about the influence of forests regulating streamflows continued through the 1940s. By the 1950s, however, there was a growing new-and-contrasting scientific evidence that forests consume more water than shorter vegetation (Law, 1956). Thus, controlled catchment experiments were required to evaluate the effects of land use changes in forested areas and in areas whose developments incorporated good soil and water conservation practices. The experiments would then prove either hypothesis before any deforestation and subsequent agricultural development was implemented.

By mid 1950, a programme of intensive commercial tea farming in East Africa was planned by the British government. The conflicting scientific hypotheses had to be incorporated in the planning of the agricultural developments. A series of experiments was started in Muguga, Kericho and Kimakia in Kenya, Mbeya in Tanzania and Atumatak in Uganda in 1956 to investigate the hydrological effects of land use changes. Detailed results are reported in Pereira and Hosegood (1962) and Pereira (1965, 1973, 1979, 1981)

The findings from these East African studies as summarised by Edwards and Blackie (1979) and Blackie, Edwards, and Clarke (1979) provided actual data necessary to test the widely held theories and hypotheses on the effects of changed land use on hydrological regimes, particularly in the tropics where hydrological information was scanty. Their findings were:

- That replacement of rain forest by tea estates at Kericho, had an overall reduction in water use, combined with no significant runoff or sediment loss. The result applied only on soils experiencing similar rainfall distribution and with an equally efficient soil and water conservation measures.
- That replacement of bamboo forests by pine softwood plantation at Kimakia initially decreased the water use, but once the pine canopy had closed, no significant differences in water or sediment yield were detected.
- That at Mbeya, replacement of evergreen forest by smallholder cultivation on very steep slopes resulted in large increases in water yields. The dry seasonal base flow doubled but the maintenance of the seasonal flow patterns and water quality depended on soil type.
- That bush clearing followed by several years of cattle exclusion resulted in a remarkable grass recolonization of the severely overgrazed rangelands at the Atumatak catchments. The recolonization increased infiltration rates and drastically reduced peak flows. Subsequent controlled grazing did not affect the hydrological stability of the improved regime.

Similar findings of increased water yield downstream as a result of reduced vegetation biomass was reported in the Coweeta watershed research in North America, a temperate catchment, some 10 years earlie (Odum, 1953). The 50-year continuous and long-term Coweeta ecological studies contributed to the understanding of man-made perturbations on major inputs, outputs and internal functions of ecological systems.

Two underlying philosophies guided the research approach at Coweeta: (1) that the quantity, timing, and quality of streamflow provides an integrated measure of success or failure of land management practices, and (2) good resource management is synonymous with good ecosystem management. Response to watershed disturbances in the two case studies was used as a research tool for interpreting ecosystem and hydrological regime behaviour. The use of perturbation or disturbance allowed the specific hypotheses to be tested with subsequent revision and development of theories and application of results when and where appropriate.

Effects of human activities on the environment however, have of recent times assumed adverse proportions. Conflicts of interests and survival of mankind became the central issue as put forward precisely by Penman (1979) as:

"Even for a primitive biological existence at subsistence level mankind needs air, water, food, and energy, adequate in quantity and in quality. There is no special order of importance, but putting water first or second could be defended. What does matter is that getting access to any of the four may produce conflict with attempts to ensure access to the others, and in the progress towards civilisation there has been occasional political recognition that controlled management of the environment is desirable to minimise the effects of conflicts".

Such recognition brings forth the issue of land use and catchment or watershed management especially in the tropics where the impact of traditional land use on water resources provokes serious conflicts. The challenge thus is the vast scale on which mis-use of land and water is destroying the basic agricultural and environmental resources of developing tropical countries.

Coupled with these changes is the consensus scientific opinion of changing global and regional climates (Milankovitch, 1948, Mason, 1976, Hays, Imbrie and Shackleton, 1976, Farmer and Wigley, 1985, Parry, M.L, 1992). Although the global climate shows no contemporary evidence of change, the mis-use of land can produce severe changes in local environments (Pereira, 1979). The rate of climate change is enhanced by the effects of changes in rainfall-runoff-soil-plant relations where human intervention is constantly taking place.

The distribution of heat and moisture over the earth's surface is the major determinant of climates both on the large and small scale (meso-and-microscale). Whereas large-scale transport of properties on the global scale determines the climatic features, thousands of kilometres downstream, climate on the meso-and microscale is tied to the local surface characteristics because of the profound and immediate impact on the local energy and moisture budgets (Lawson, et. al, 1979). Modification of the microclimate is readily brought about by changes in surface cover because of the increased 'harshness' of the environment due to the misuse of the cleared land (Goodland et al.,1984). Hence human activities that degrade a watershed characteristic in the end will modify its local micro-climate which in turn influences its hydrologic regime.

Studies on the effects of human activities on hydrological regimes have however been at the heart of hydrological sciences for many years. Because of the complexity and integrated nature of the problem, it has been difficult to particularly isolate specific causal effects. This is because one human activity is rarely performed in isolation (Arnell, 1989) and the hydrological characteristics of drainage basins are thus an integration of the effects of different human activities operating at different scales.

It is now accepted that decreased infiltration and increased runoff can result from the destruction of protective vegetation, the tramping or over-cultivation of the surface soil and its resulting exposure to radiation and to rainfall of high intensities. The recent drought disaster in Africa (1973/75, 1980/82 and 1984/85) for example, may have been enhanced by the impacts of excessive human population and livestock in a fragile ecosystem (Farmer and Wigley, 1985), otherwise droughts of similar scales have occurred in the past, but without such a devastating effect.

A clear and very recent consequence of rapid deforestation is that reported by Zaimeche (1994) in which Algerian widespread deforestation was connected to the drier conditions affecting the region. A low social and economic levels induced deforestation on such a large scale that erosion and soil losses are reaching unprecedented levels.

2.2.2. The Sustainable Development Perspective

Watershed management is concerned with the control of water transfer from the upper to the lower reaches of a river's catchment area. The amount of water transferred from the upper reaches depends upon the difference between rainfall and evapotranspiration. Evapotranspiration itself depends on the vegetation, the soil depth, soil water holding capacity, and surface run-off (soil-water-plant-relationship continuum). Hence both the amounts of water leaving the watershed, and the seasonal flow of the river is dependent upon human interventions on the watershed characteristics.

The amounts of water entering and leaving the lower reaches of the river system also depend on the management and user rate of water in the upper reaches. The management practices hence, influence the hydrological regime of the river in terms of frequency and magnitude of flooding, sediment yields and its streamflow seasonality. Theoretically, therefore land use systems in watersheds should be designed in a manner that minimises the transpirational demand, increases dry season flow, and sustains the watershed ecological and hydrological regimes.

Against this background, the concept of sustainable development was devised. In its report, the World Bank Commission on Environment and Development (1987) defined sustainability as:

" the development that meets the needs of the present without compromising the ability of the future generations to meet their own needs"

In other words, the satisfaction of human needs and aspirations is the major objective : of farms; of factories, of housing to provide shelter; of water and sewage services to improve the quality of and healthiness of life (Clarke, 1994). After the satisfaction of these basic and legitimate needs, they should be able to aspire towards a higher quality of life. To achieve these aspirations the Chairman of the World Bank Commission on Environment and Development, Brundtland stated that:

"...a world in which poverty and inequity are endemic will always be vulnerable to ecological and other crises" hence the actual cause should be rectified first.

The idea of sustainability was the central theme in the United Nations Conference on Environment and Development (UNCED), the Earth summit held in Rio de Janeiro, Brazil in 1992 which produced several resolutions among them: Firstly, adoption of Agenda 21-a forward looking and comprehensive global action plan leading to the 21st century with the needs of the developed and developing countries taken into account. Secondly, the establishment of a Commission on Sustainable Development under the aegis of the United Nations, whose role is to monitor the implementation of the resolutions.

On the basis of these events, hydrologists and ecologists should provide information for sustainable development of watersheds. Historically, hydrologists studied the effects of forestry in streamflow generation in a restricted scientific sense. The issue of land degradation causing changes in flow regimes was initially left to agriculturalists. Since the research now cover the whole spectrum of land use, hydrologists, ecologists, agriculturalists and land developers should work together for a sustainable development of the natural resources especially at the watershed level.

As scientists and policy-makers become more conscious of this trend of environmental degradation, methods and actions of alleviating the situation are being studied. To scientists, the first task is to use all the resources of science and humanity to abate it.

The second alternative is to strive for a radical improvement in the management of the soils, vegetation, and water supplies in the affected regions and thus arrest the problem. These tasks are all full of uncertainty because it depends not only upon the competence and appropriateness of the technology chosen but also on the winning of hearts and minds (Pereira, 1979).

The most plausible approach is perhaps to return the land-use in these regions to hydrological stability by exploiting their ecological role as demonstrated by Edward and Blackie (1979) in their analysis of 16 years of data on land use changes in East African catchments. Their studies demonstrated that a complete commercial tea estate in Kericho, Kenya, with roads, houses, factories, offices and workshops can be developed in a stream source area of tall forest without long-term effects on soil stability or on the flow regime. This was achieved by meticulous planning for soil conservation and runoff control, followed by highly competent execution and subsequent management.

It is however, still difficult technically, socially and economically to tackle land mis-use in the semi-arid climates of the tropics. These climates are characterized by high rainfall variability and extremes of temperature which make the area prone to rapid deterioration under mis-use. For example, Kenya's Machakos area had an intensive mechanical conservation network from 1930s and yet aerial photographs taken in 1948 and 1972 showed sheet erosion from overgrazing had increased by one third in the 24 year-interval and many terraces had failed through soil erosion from the overgrazed areas (Thomas, 1974). It was only after an intensive integrated conservation approach between 1972 and 1990 that hydrological stability was achieved (Mortimore et al., 1993) evidenced by increased unit agricultural production and increased dry seasonal flows; a measure of success of soil and water conservation practice.

The introduction of fruit tree crops in terms of agro-forestry has been advanced as a model in sustainable social forestry development. The International Council for Research in Agro-forestry (ICRAF) in Kenya has produced very useful findings on the subject (Cheatle et al., 1986). In theory, land in arable crops and under a good soil and water conservation management should allow rain water to percolate into the groundwater and to seep slowly into the river (Lal and Russell, 1981) and hence sustain stream minimum flows.

Draining of swamps in the upper reaches of watersheds is yet another human intervention in watersheds in the tropics. These swamps hold considerable volumes of water that are slowly released as inflows. But with the draining for crop production and for grazing during the dry seasons, this reduces the recharge rate of rivers. Evidence of massive drainage and human interference of these wetlands in tropical Africa is abundant. For example, Kenya is addressing this problem in the 1990s through Kenya Wetlands Working Group (KWWG) established to undertake research on sustainable development strategies in the wetlands.

2.3. METHODS OF STUDYING EFFECTS OF HUMAN ACTIVITIES ON HYDROLOGICAL REGIMES

Impacts of human-induced watershed changes can be studied in a number of ways: the analysis of long time series, the use of regional comparisons and transfer functions, the experimental basin studies (control method) and the use of mathematical modelling techniques. The suitability of any of these methods varies according to the availability of data, capital costs, time and technological manipulation. But generally the assessment of human effects is often complex and difficult (Petts, 1984).

2.3.1. Analysis of Time Series Method

This involves the examination of the time series of hydrological data to search for evidence of change for example using double mass curves (WMO, 1988). A bend in the cumulative curve of a variable would indicate a change which would be quantified from the angle of bend. The cumulative curve is studied to make sure that the change was actually due to the suspected human influence. Further, the method involves the searching for periods of persistent trends and periodicities as well as the statistical behaviour of the historical data over time and space.

Arnell (1989) and Seuna (1989), however, caution on wholesale use of the method because of the difficulty in attempting to infer a causal mechanism from the end result alone. They pointed out that unless the human activity was sudden and of large impact, it would be difficult to distinguish the effects of the action from other activities or from normal background climatic variation, hence other methods may be used as a check.

Further evidence of this difficulty is reported by Higgs and Petts (1988), in their study of the Upper Severn Catchment in UK. They found variations in flood occurrence over time could be explained by climatic variations but the effects of reservoir impoundment could not be isolated while using the analysis of time series method. The method has however, worked elsewhere as reported by Stoddard (1991). Its success depends on the accuracy required, the time availability and amount of funds allocated (Seuna, 1989).

2.3.2. Regional Comparisons and Transfer Functions Method

The method combines data from several catchments supposedly experiencing different human activities to infer the effect of different land uses on hydrological characteristics (Blackie, 1973., 1974). Romanian Hydrological group used data from several catchments to illustrate the effect of land reclamation works on sediment yields. The limitation of this method is that unless the catchments are hydrologically similar, other characteristics may also be causing the observed variations in the hydrological behaviour. There is however, no such a thing as hydrologically similar watersheds, hence assumptions are included by generalising discussion of the results.

To incorporate the effects of a wide range of catchment characteristics, the approach uses multiple correlation and regression analyses. Regional transformation of information may be feasible only when the very characteristics influencing between-basin variability are relatively constant.

2.3.3. Experimental Basin Method

This is considered as the classical approach (Rodda, 1986). The areas under study and that under control are kept in their original state for the calibration period that is usually between 10 and 15 years. During this period, hydrological parameters (e.g runoff, infiltration, peak flows and water quality etc.) are measured, correlated, and quantified and those of the study area are correlated to those collected from the control area. After the treatment (human disturbance e.g clear cutting or chemical application), regression equations obtained are used to study how large the study parameters would have been, if the experiment had not been carried out.

The difference between the measured and the calculated values reflects the treatment. Rodda (1976) reviewed the development of the method and several researchers have used it extensively. The method assumes that all catchment controls such as topography, geology, and climate for practical purposes would remain constant across the basin during the study period, but of course there is always a likelihood of an event taking place in which case the experiment may be abandoned or results from it generalised. Most researchers prefer to use this method. It is an expensive method because the watersheds are instrumented individually and data recorded and analysed at the same time. The study period however, is too long for a research whose results are needed urgently to solve an engineering problem.

2.3.4. Mathematical Simulation Modelling Method

A model is a simplified representation of a complex system that may be either physical, analogue or mathematical. The mathematical models are those in which the behaviour of the system is represented by a set of equations, perhaps, with a set of logical statements, expressing relations between variables and parameters. A variable in this case is understood to be a characteristic of the system which may be measured and which assumes different values when measured at different times, while a parameter is a quantity characterising a hydrological system and which remains constant in time (e.g.. area of a watershed). Clarke, R.T (1984) provides an intensive work on the use of mathematical models in hydrology.

The impacts due to changes of watershed characteristics can also be studied using these mathematical models (Clarke, R.T., 1984). The models work very well with shorter periods of observation and a control area is not required. Use of the models has been applied to several hydrologic problems for a long time. They range from relatively simple conceptual models to complex ones and require a large number of variables.

Use of models reduces costs considerably compared to the other methods reviewed. They are however limited in scope because of the risks inherent in applying a model with input data outside the range of the inputs used during the model derivation and calibration (Arnell, 1989). Therefore, predictions of change will be associated with the varying degree of uncertainty (Rogers and Anderson, 1987). However, models should not replace field observation wholesale and they should only be used where unavailability of data is the problem.

This review has thus shown that, changes in hydrological regimes of an area can in fact be assessed with some degree of accuracy- all of which depend on the purpose, extent and the availability of funds for the study. Selection of any method depends upon the above factors. However, developing countries with limited financial resources may find the adoption of this method difficult.

On a general observation however, several studies have been carried out using a mix of the methods depending upon situations and local conditions. For example, Seuna (1989) used experimental basins in Finland to investigate the influence of man's activities on forests runoff and water quality. Results from the the studies were meaningful and reliable.

Murthy (1980) estimated the hydrological changes caused by human activities in the Damodar Valley in eastern India using methods ranging from simple graphs and double mass curves to deterministic and stochastic models depending on the nature of the problem under investigation. He cautioned, however, that the success of the methods used depended on the availability of quality data.

Studies have also been carried out by interpreting basin characteristics using landsat digital multi-spectral scanner and interpretation together with aerial photographs and topographic maps (Kuittinen, 1980). This method involves comparing maps obtained at different periods and interpreting the extent and magnitude of change of the basin characteristics. Such maps contain a great deal of information for hydrological purposes.

The methods used in the detection of the effects of human activities on the whole depended upon the character of change. For gradual changes Matalas (1963) recommends that trend analysis may be applied and homogeneity tests carried out for sudden changes. Investigation of trends in a time series is often done using the moving averages procedures described by Mcuen and Snyder (1985).

2.4. HYDROLOGICAL CHANGES DUE TO AGRICULTURAL DEVELOPMENT

2.4.1. General Hydrological Changes

Several studies have shown that land use and various land management activities have substantial effects on hydrological regimes. Changes in land use further alter the time to peak of streamflow hydrographs, thus modifying the flood frequency characteristics, due to the changed amounts of runoff and/or the response time of the watershed (David, 1983). Sangvaree and Yevjevich (1977) for example compared unit hydrograph characteristics of agricultural and forested watersheds in the eastern United States and found the unit hydrographs peak for agricultural watersheds was 2 to 4 times runoff in comparison to forested watersheds confirming the long held opinion that agricultural activity encourages increased surface runoff than forested lands.

Good agricultural development practices also influence the hydrologic behaviour of a hydrologic system in a positive manner, by contributing to water retention, increasing soil porosity and permeability, increasing infiltration and by reducing the overall surface runoff. These hydrologically favourable practices include the protection of fields against erosion, application of different cultivation methods that encourage maximum water retention, and crop rotation. Thus, agricultural conservation practices can have significant impacts on the quantity of water that originates from agricultural watersheds.

These agricultural practices substantially affect direct runoff by 1) changing the volume of runoff and 2) changing the peak rate of runoff. Changes in runoff volumes generally change the peak rate in the same direction, however, peak rates can be changed without necessarily changing the volumes. Infiltration rates are increased by agronomic practices that provide dense vegetative cover, abundant mulch or litter, high organic matter content, good soil structure and good surface drainage.

Mechanical conservation practices such as contouring, contour furrowing, graded terraces and level terraces substantially increase surface storage (Lal and Russell, 1981) and will alter the hydrological regime of the area. Any form of agricultural development in a watershed will to a certain extent influence the hydrological behaviour of the ecosystem. Thus agricultural practice in a catchment lead to changes in the physical and chemical characteristics of land surface and ground water at downstream of the impact points.

2.4.2. Effects of Agricultural Practices on Streamflows

Processes causing massive deterioration of tropical and temperate land and water resources are the use and mis-use of mechanical and subsistence farming. Pereira (1981) attributed pressures on the tropical African forests due to increasing human population growth as the major factor and reason that have forced subsistence farmers to intensively cultivate the land. The causes of both agricultural, and hydrologic droughts in Africa, for example the 1980 and 1984 disaster, may have been enhanced by the impact of excessive population and livestock in a fragile ecosystem(Darkoh, 1993) and Farmer and Wigley (1985).

Several investigations have found that conservation tillage (no-till) reduces surface runoff as compared to conventional tillage (Baker and Laften, 1983, Edwards and Amerman, 1984). Burwell and Kramer (1983) analysed 24 years of runoff records for plots in Missouri (USA) and found that direct surface runoff for conventional tillage was 120 mm versus 100 mm for conservation tillage; a reduction of 13 per cent. Wendt and Burwell (1985) found that a winter cover crop reduced surface runoff as compared to standard no-till for corn but that surface runoff was about 20 per cent less for reduced tillage than for conventional or other methods. Thus, while farmers will never stop cultivating, a proper choice of good methods can still positively alter the behaviour of soil and water movement characteristics.

A significant trend in land management has been the rapid adoption of conservation tillage as a soil erosion control measure (Spomer and Hjelmfelt. 1986., Langdale et al. 1979). Hinkel (1983) reported that these methods received high adoption rates and were expected to be used in about 65 to 85 percent of US croplands by the turn of the century. This adoption rate will definitely stimulate a large number of investigations into the hydrologic effects of these conservation methods.

The response of surface runoff to conservation tillage shows a substantial variability because surface mulch reduces surface crusting and reduces runoff velocity. Soil evaporation is also reduced because the residue changes the wind profile and albedo of the soil surface (Shane, 1983). Kramer (1986) further confirms that conservation tillage reduces runoff and soil loss on cropland. Crop stage periods of rough fallow, rapid growth, reproduction and maturation and residue were analysed with significant differences found only for crop stages, seed bed and residue. The most substantial difference in runoff between the tillage treatments was crop stage seedbed, where the mean runoff for conservation tillage was 33 per cent less than that for conventional tillage. The mean soil loss in crop stage seedbed for conservation tillage was 67 per cent less than the conventional tillage. It can therefore be inferred that runoff peaks would be reduced by conservation tillage with reduction being greater for small runoff events.

Cropping patterns and systems also influence the total peak runoff rates and soil erosion. Thomas et al (1981) using simulated rainfall in Iuini catchments in Machakos, Kenya, found runoff percentages were higher from grazed than from the cultivated lands. Earlier, Thomas and Barber (1978) using simulated storms of 25 mm/hr intensity for two hour duration on a degraded grazing site and an old pasture found the runoff percentages to be 81 and 24 per cent respectively. Barber et al. (1979) reported closely similar results from the same catchment. Thus, the amounts of runoff are much greater from grazing land than from cultivated land for both high and low intensity storms.

Soil type and distribution influence runoff rate and soil erosion. Kampen et al (1981) in the semi arid tropics of India, studied runoff and soil loss during high intensity-long duration storms. The total peak runoff rates and soil erosion were much lower on vertisol soils in cropped broadbeds than for other management treatments. The rainstorm intensity and cropping systems would then affect the hydrology of the area. Unfortunately, man has little influence on the soil type and intensity and duration of rainstorms in the tropics, and thus can only alter his mode of farming.

2.4.3. Summary on the effects of Agricultural Development

The review has shown that agricultural development influences the watershed hydrological cycle. It intensifies water use rates, reduces river base flows, accelerates soil and land degradation and increases river bed siltation. Opening of new lands for agriculture removes vegetation cover, which changes the energy balance of the area, thus increasing or decreasing the evapotranspiration (depending on the availability of moisture). The alteration of the water balances ultimately influences the patterns of the local micro-climates. Without moisture, aridity and land degradation is worsened. Over a time the combined effects will alter the hydrological regime of rivers in the areas being developed. However, a well-planned agricultural development (with complete soil and water conservation measures) in certain watersheds in the humid tropics (for example the Kericho case in Kenya) can sustain agricultural development without influencing adversely the hydrology of the area.

2.5. HYDROLOGICAL CHANGES DUE TO FORESTRY DEVELOPMENT

2.5.1. General Hydrological Changes

The case of Kenya probably provides an example of recent changes due to human population pressure on forestry resources. Wood provides approximately 72 per cent of Kenya's energy demands (GOK, 1989). Large populations mainly in the rural and semi-urban areas of the country depend on fuelwood and charcoal for household cooking and space heating. By 1980, the rate of forest depletion was estimated by LANDSAT surveys as 1 percent per year (GOK, 1979). On the basis of these findings, the Kenya government emphasised elaborate forestry conservation strategies in the 1980s, through the initiation of soil conservation strategies of afforestation and restricted felling of trees so as to ensure a balanced ecosystem and abate the rate of environmental degradation.

Fuel wood consumption in Kenya for example rose from 23,402 cubic metres in 1980 to 33,884 in 1989 with a 15 year mean annual consumption rate of 3.5 % (WB, 1990) while sub-saharan Africa annual wood consumption was 3.0 %. Sub-saharan Africa population rose from 438.79 in the 1980 to 592.13 million people in 1990, a 3.1% average rate of growth up from 2.6 % in 1975. During the same period urbanisation in the region rose at annuall rate of 5.3% with Kenya having a higher rate of 7.1%. This increased population increased the use of forestry and woodland products for energy both in the rural and urban areas.

By 1986, there was about 3.4 per cent forest cover in Kenya (GOK, 1988) with an annual afforestation rate of 0.28 % (1610 sq.km/yr). With a 1% rate of forest depletion it is expected that the total forest cover will reduce to 3.03 % in ten years. This is a worrying trend because of the inherent threat of environmental degradation through increased peak flows, reduced low flows and changes in local micro-climates. The rate of depletion may appear relatively small compared to the afforestation rate on a national perspective, but the effects on small river watersheds like that of Enjoro river is significant. Continuous monitoring on specific local watersheds should therefore be emphasised by identifying environmental pressure points and danger signals upon which to base the corrective measures.

The African continent depends mostly on farming. To feed the increased population, the agricultural area was increased from 17062 million hectares (ha) in 1980 to 17685 million ha. in 1988, with Kenya increasing by 15 million ha during this period(WB, 1990). Deforestation however was the main activity, to open new lands for food production and to supply fuelwood. During the period under review, Kenya's annual deforestation rate stood at 39000 ha. per year while the sub-saharan Africa countries lost a total of 3846,000 ha. of forest per year.

2.5.2. The Effects of Deforestation on Streamflows

Modifications of land surface characteristics involving the clearing harvesting and burning of forests for the purpose of crop production or animal husbandry ignore the need to preserve the delicate balance between the environment and the maximum vegetation. The modification reduces the transpirational surface area, reduces interception and re-evaporation and decreases soil water storage capacity (Lawson et al., 1981, Pereira, 1981). These activities further affect the soil micro-floral and faunal activity (Cummings, 1963, Lal and Cummings, 1979) and upsets the soil texture thus accelerating soil erosion.

Studies at Coweeta watersheds in the US and other sites have demonstrated that logging increases streamflow (Likens et al. 1970). Increased flow was shown to last 20 to 30 years following logging (Kovner, 1956; Hewlett and Hibbert, 1963), hence it can be seen that long-term disturbance can result in long term streamflow disturbance. Hibbert's (1966) survey of experimental catchments throughout the world showed that deforestation increased and afforestation decreased mean annual runoff, but the magnitude of response was highly variable and unpredictable to a high degree of accuracy.

Deforestation is acknowledged as the most widespread human impact in watersheds. It is also extensively studied both in the temperate and tropical regions (Swanson et at., 1987) and its effects recognised as increased annual runoff due to reduced infiltration rates (Arnell, 1989). Forests and vegetation cover have a direct effect of increasing evapotranspiration (ET) in temperate regions where soil moisture is not limiting, but on clearing, the result is an increase in short duration runoff (Seuna, 1989). These changes take up to 25 years for the watershed to return to its original hydrological regime (Swank and Douglas (1974) after the disturbance assuming reforestation occurs.

Similar runoff increases due to changes in forest cover have been studied in United States (Douglas, 1983., Troendle, 1983). Hibbert (1983) reports that under certain conditions ET was reduced and water yields were increased by the removal of trees and replacement of grass cover.

The increased rate of deforestation in tropical Africa in recent years can be attributed to natural and socio-economic stability in the region. First, forests are cleared in the tropics to create additional land for farming, grazing, fuelwood, and to settle and absorb pressures of the ever increasing population . Kunkle and Dye (1981) attributed land shortages to changes of traditional shifting cultivation methods in Africa, to intensive land cultivation and mechanisation. Because of frequent intensive cultivation, land degradation cycles are repeated annually and the opening up of lands more often through forest clearing completes the cycle.

Studies of the effects of deforestation on hydrological regimes in tropical Africa however, have been on the general water balance area (Lal, 1981). Effects of the methods of deforestation and pasture management on streamflow and water quality have not been extensively studied and quantified due to the economic status of the countries in these regions, thus studies on changing patterns of water quantity and quality are still scanty.

While extensive studies have been done in tropical forests, there is very little known on the effects of vegetation cover clearing in the arid and semi arid environments. Effects of land use in some East Africa experimental watersheds was however undertaken by Pereira (1962, 1965, 1973, 1979, 1981) and McCulloch and Dagg (1965). There has been no study of similar magnitude and quality in the region since then.

In Kericho, Kenya, Pereira (1965) reported that a clearing of 30 per cent of natural forest for tea plantation reduced streamflow by 11 per cent because greater care had been taken to minimise soil erosion during the period of plantation. By the time the tea bushes gave effective complete cover, Blackie (1972) reported that the annual water consumption was virtually unchanged compared to natural forest. It is to be noted here that the recovery period was relatively shorter due to good soil and water conservation measures undertaken by the developers during the initial stages. The results contradicted that of Hibbert (1966) possibly because of the pre-planned soil and water conservation works.

Later, Pereira (1973) measured 90 mm rainfall storm on a newly cleared land and before canopy cover that resulted in a maximum streamflow of 27 m^3s^{-1} per km^2 from cleared catchment compared to only 0.6 m^3s^{-1} per km^2 from the forest control. A substantial increase in streamflow due to the clearing and lack of canopy growth in the tea plantation resulted. Similarly, Othieno (1979) reports higher runoff rate in Kericho soils with less vegetative cover. The Kimakia catchment in the Aberdare mountains, under the same study showed similar patterns but the study in Mbeya, differed because of the nature of the soils and effects of intensive smallholder cultivation on the catchment (Edwards and Blackie, 1981).

Responses to forests clearing depict more or less similar hydrological patterns both in the temperate and tropical rain forests. Investigations conducted by Richardson et al (1979) to establish the hydrological effects of mechanical or chemical bush control mesquite in Blackland prairie, Texas,(temperate) resulted in an 8% reduction in annual evapotranspiration and a 10% increase in annual runoff. Lawson et al. (1981) studied the water balance, meso and micro-climatic parameters over a forested area at Ibadan, Nigeria (tropical), in which cleared and cropped watersheds were used. Meteorological parameters were monitored in the 1979 cropping season to observe the rainfall distribution and micro-climatic changes in the watershed. 73 % of the incoming rainfall reached the forest floor as throughfall, the respective values of stemflow and interception were 10 % and 17 %. Surface runoff from the cleared area was 23 % of the rainfall compared to the negligible pre-clearing value and similarly negligible runoff from an adjacent forested watershed.

Model simulation studies have confirmed the changes in hydrological regimes due to forestry management irrespective of its geographical area. Studies and modelling by Lockwood and Sellers (1981) found clearing of forests to lead to a generally wetter soil moisture regime and higher runoff under a constant climatic condition.

2.5.3. Summary on the Effects of Forestry Development

In summary, forestry development is seen in two perspectives. Deforestation changes the land vegetation cover decreasing infiltration rates and capacities, and increasing peak runoff. On the other hand, afforestation encourages production of good canopy that increase moisture interception and thus increase infiltration and moisture storage in the watershed. There is however an increase in evapotranspiration, reduced continuous river base flows and a humid micro-climate. Hence, a forestry development that balances the two aspects is appropriate for sustainable use of the resource especially in the fragile tropical environments.

2.6. HYDROLOGICAL CHANGES DUE TO URBANISATION

2.6.1. General Hydrological Changes

The migration of human population from rural to urban areas is growing in all continents involving changes in land occupancy and use. Half of the world's population will be living in urban areas within the next few decades (Uehara, 1980) mainly working in commercial and industrial enterprises. In the 1960s, Savini and Kammerer (1969) had recognised the consequence of urbanisation associated with hydrologic aspects of water management, as caused by complex merging of social, economic and physical problems. The resultant effect is a superimposing of various human activities, increases of water abstractions, increases of sewage discharges, and reduced minimum streamflows. The general effects of urbanisation include the altering of the drainage system, increase of impervious areas that reduces infiltration. Therefore the interrelationships of man, his use and development of the land and water resources in these urban areas must affect the urban hydrological behaviour with time.

Studies by Arnell (1982a) and Richter and Shultz (1987) found that the impacts of urbanisation increased flood peaks and reduced lag times; and reduced minimum base flows because the precipitation falling on the catchment is fast removed. These hydrological effects result in decreased water tables in the area. Richter and Shultz (1987) studies on the Schwippee River in Germany showed increases in flood wave peaks from 0.8 to 2.9 m^3s^{-1} when the degree of urbanisation increased from 6.5 to 25 % of the total urban area. The increase in peak flows of approximately 1 m^3s^{-1} for every 9% increase in urbanisation changed drastically the channel formation of the river.

2.6.2. The Effects of Urbanisation on Streamflow Patterns

Poverty is recognised as the main cause of human-induced environmental degradation. For example, by 1985, Kenya had about 10% of its population below poverty level in the urban areas and 55 % in the rural. Fourteen per cent of the population in the urban and 20 % in the rural areas in sub-saharan Africa were below poverty level. These increased poverty levels and imbalance in rural areas forced the population to emigrate to urban areas in search for employment and food. These socio-economic and other political factors directly and indirectly thus influenced the urban hydrological environment.

Most sewerage and drainage systems in old urban cities in Least Developed Countries (LDCs) are of the combined type designed originally to deliver small quantities of surface runoff, domestic and industrial wastes (Colenbrander, 1971). As the cities increased in size, there was a corresponding increase in impervious area, sewage and industrial effluent. The inadequate drainage facilities resulted in raw sewage overflows into the receiving waters and thus altered the peak rates' patterns. United Nations Development Programme, Dosiour (1976) for instance reported on the rapid growth of the Nairobi city in Kenya and its surroundings by 600 per cent between 1950 and 1970. This growth rate overloaded and nearly paralysed the drainage system in the city resulting in occasional unpredictable flooding (Krodha, 1986). Several cities in Africa and in effect in most of the LDCs experience similar degree of hydrological changes.

Czja and Janowski (1986) studied the impact of urbanisation and industrialisation on 18 selected rivers in Poland during the 1961-1990 period. Their conclusion was that anthropogenic factors in the catchment caused large disturbances in the seasonal runoff character of the rivers regardless of the location of the hydrometeorological situations.

The effects of industrialisation are sometimes treated as an integral part of urbanisation during transition from middle urban to late-urban stage. Extensive exploration of groundwater resources to meet the industrial water demand in these areas produce cones of depression in most cities and water intakes. The abstracted quantity of water decreases the base flows in medium to small rivers around the cities and some smaller ones dry up.

Studies in Long Island, New York by Simmons and Reynolds(1982) showed that urbanisation reduced the total annual streamflow occurring as base flow from 95 to about 20 per cent of total streamflow and attributed the cause of the reduction to an integration of first the increased sewered area, secondly the routing of storm runoff directly to the streams through sewers and thirdly the decrease in infiltration of precipitation as a result of reduced permeable area.

Uehara (1980) attributed the lowering of ground water tables in urban areas to the tendency of concentrating demands for water supply and concentrated ground water withdrawals, and found that the smaller the catchment area, the more pronounced the impact can be discerned and the greater the probability of a particular impact becoming predominant. These and other hydrological effects are also common in cities and municipal urban areas of developed and developing countries.

2.6.3. Summary on the Effects of Urbanisation

The review on urbanisation has demonstrated its effect on the environment and increased user rate of the natural resources. As large numbers of humans migrate from rural to the urban areas, there is a reduced land occupancy, increased construction of buildings, roads and other infrastructure and an increase industrial development. The result is an increased water use rate that triggers larger surface and groundwater development. Urbanisation also increases impervious surfaces that reduce infiltration and hence increased frequency of flooding. The reduced infiltration reduces recharge of rivers within and bordering large cities resulting in low flow patterns and rendering them hydrologically unstable .

2.7. HYDROLOGICAL CHANGES DUE TO HYDRAULIC STRUCTURES

2.7.1. General Hydrological Changes

The construction of water storage, conservation and control works for stream and water level regulation is unavoidable so long as mankind must satisfy the requirements of irrigation and drainage, hydro-power development, inland navigation, water supply and other purposes. Such man-made works eventually lead to environmental and hydrological changes. The effects range from changes in streamflows and its quality to alteration of the hydrological regimes of the affected watersheds.

Inundation, is another effect of storage reservoirs and of river impoundment (Ackerman et al., 1973). The loss of habitat due to this effect is especially severe when reservoirs are situated close to mountains in dry areas. In these areas, the ecological and geomorphologic formations are fragile and susceptible to abrupt changes. Nilson et al. (1991b) found large reservoir dams to reduce river discharges downstream of the dam as a result of increased evaporation losses. Major parts of former river shoreline remain unflooded and in the long term, drained river banks develop terrestrial vegetation, whereas new shoreline vegetation establishes along the new waterline. The amount of the new shoreline vegetation is smaller relative to the former because of reduced range of water level fluctuations and a relocation of the shoreline from the edge of a wide flood plain to the channel margin. This alteration of the river habitat takes years to return to re-establish a self-sustaining status.

Because of these related effects of hydraulic structures, many countries introduced legislation to include environmental impact assessments (EIAs), in their planning processes to identify projects that pose unacceptable environmental impacts (Bolton et al.,1990) This has however, created a friction over the future development of water resources, and their impact on the environment, between engineers and planners on one hand and environmental groups on the other. Such a conflict from an environmental point of view sounds an excellent mode of training natural resource developers and users alike on the importance of sustainable development.

2.7.2. Hydrological Changes Due to Irrigation

Irrigation is practised to enhance agricultural production. Most irrigation projects however, have been planned, constructed, and commissioned without adequate consideration of their far-reaching effects on the environment. These effects influence the operation of the irrigation scheme itself and the wider area of human and natural environments. The mostly acknowledged hydrological impacts include: low flow regime, increased flood regime, fluctuations in groundwater levels and a host of other effects that changes the water-plant-soil-relationship.

Large irrigation schemes may also cause a hinterland effect, loss of natural vegetation, land degradation and increased erosion in the surrounding areas. These are due to population pressure, changes in animal husbandry, increased dryland farming, deforestation, infrastructure development and economic activities stimulated by the irrigation project.
The International Commission on Irrigation and Drainage (ICID), recognising the effects of irrigation and drainage on the environment , set up a Working Group on Environment: Impacts of Irrigation, Drainage and Flood Control Projects in 1987. The objective of the group was to gather and synthesise information available on environmental effects. The results provided guidance to project designers and managers in identifying and minimising environmental impacts.

Results from a survey reported in Bolton and Brabben (1989), on impacts of irrigation showed socio-economic effects as the most significant. Irrigation schemes affected human population in the surrounding area (in terms of resettlement, health and socio-economic conditions). A quarter of the population interviewed perceived the resettlements of people from reservoir inundated areas as a major impact. Disturbance to settlements or existing farmland was also a major concern. Through unplanned opening of new lands and associated intensive resource use with time, changes in hydrological regimes were apparent. Health impacts were of socio-economic importance with the spread of water borne diseases being a dominant environmental concern especially in developing countries.

In addition, a large proportion of water diverted from rivers for irrigation purposes is re-distributed through various routes depending on the local climatic, soil and geological characteristics. The greatest route is through increased evapotranspiration because of moisture availability, and through seepage and deep percolation. As a result, the local micro-climate and groundwater conditions are altered. Accompanying the changes are, the alteration of the soil infiltration rates and capacities, changed subsurface flow patterns, and accumulated salt content The increased salinity degrades soil structure, increases soil swelling and wilting points and lowers permeability and water holding capacities of the soil. Because water is diverted and redistributed, there is an increased time and distance for the same water to complete its hydrological cycle. Hence, the irrigation would modify the temporal and spatial hydrological behaviour of the watershed through this induced low flow pattern (hydrologic drought).

Blidaru et al., (1989), carried out comparative studies on the effects of irrigation on hydrological patterns in Romania. Seven basins were analysed with catchment areas of between 17 and 588 km^2. The precipitation data record covered the periods' 1921-1985 and the discharge data over 1955-1985 period. The study also considered the assumption that irrigation canals influenced streamflow over the period 1975-1985. They correlated the ratio between the average discharge over the period the hydrologic cycle was supposedly being influenced and the non-influenced period and found that hydrological changes due to irrigation were more pronounced in steeper and higher grounds than the flatter areas.

Excess irrigation water raises groundwater table levels. With improper drainage system, the water table rise continues until the soil mass becomes water-logged and produces practically no yield. The general phenomenon in most cases is that of changed ground water flow and generation mechanism. For example, the hydrograph of groundwater levels at some locations in upper and middle Egypt after the construction of the Aswan dam, show much more uniformity with time than before the construction of the dam (Attia et al.,1983). Most of the hydrographs shows a rise in the water table in what used to be the low-season and a fall in level in what used to be the flood season before the construction of the dam. This was attributed to the application of irrigation water in amounts far more than the actual need of the soil and the seepage from the canals running at high levels for a long time.

2.7.3. Hydrological Changes Due to Reservoir Construction

The effect of water storage, control and conservation works in the Nile river has been extensively studied. Shahin (1985), reports extreme changes in water quality and quantity along the river as a result of the construction of the High Aswan Dam. There was increased sediment loads, and degradation of the river channel during the construction of the dam. The hydraulic structures led to instability in the amount, rate and temporal distribution of silt transport to the sea and resulted in the disturbance of the quasi-dynamic equilibrium

Similar results had earlier been reported by Wassing (1964) that the rate of retreat of promontories at Damieta and Rosetta was 29m/yr in the period 1898 to 1960, but sediment transport to the sea reduced after the construction of the dam in 1964. This reduction of the annual sediment load produced an imbalance in the near-coast sediment budget, thus making the coast more vulnerable to considerable soil erosion. This led to rapid change in the estuarine circulation pattern from two-layered to one-layered flow, which increased surface salinity in the post-dam period (Shahin, 1985).

Floods and human-induced hydrologic drought in downstream watersheds resulting from dam building were cited as a human factor of major concern in Egypt. Ati (1993) examined the ecological changes that the Khasm el Girba Dam in the Atbara river brought in the downstream areas over the last three decades and their socio-economic implications. Prior to the construction of the dam, the lower Atbara economy was predominantly agricultural, dependent on a well-planned community based irrigation on the Atbara river. The ecological changes caused by the dam under a steady declining rainfall situation in the area since the 1950s included the following: severe drop in the amount of water passing downstream, erosion of the river channel downstream, siltation of the reservoir, deforestation of the catchment, disturbance of the breeding and feeding systems of the fish population and a sequential process of environmental degradation. The reduced water supply affected the main economic activity of the population. As a result, unemployment, under deployment and out-migration occurred at very high scales. These findings evidently illustrate the disastrous consequences of dam construction without consideration of their environmental and accompanying socio-economic impacts downstream.

Bliddaru et al., (1985) quantified the influence of hydraulic structures on hydrological processes, using a modified version of the Stanford IV mathematical model type. Using additional subroutines and variable steps, he modelled processes involved in runoff generation considering the interdependencies of the elements of the hydrological cycle. Predicted results showed that river training and levelling reduced concentration times by approximately 30%, due to the modification of the slopes and roughness. Total runoff coefficient increased and the overall coefficient under floods increased by 15% with hydraulic structures. These simulations emphasised a relative decrease of storage and modified time variation as compared to the natural regime. The surface storage decreased and slope rose from 0.5% to 0.65%.

2.7.4. Summary of the Effects of Hydraulic Structures

The reviews on the effects of hydraulic structures have shown large dams and large-scale irrigation schemes to influence adversely the environment and its inhabitants. Large population concentration in irrigation schemes increases water demand; hydraulic structures on rivers change the river's surface and subsurface characteristics. There is a high variability of groundwater tables because of fluctuating levels of water in reservoirs and because of the intervals of irritations. All these combine to reduce river flows, to increase silt and sediment load and to degrade the river shore lines. Because of moisture availability in irrigated areas, increased moisture loss occurs through evapotranspiration, and the local micro-climate is varied so that the general flora and fauna ecosystem are affected. Overall, hydraulic structure influences a watershed hydrologic regime for the worse as resource use rate and quality and quantity are altered, and may be depleted in the future.

2.8. CONCLUDING REMARKS

2.8.1. General Observations

The literature review has brought out a clear message and challenge to researchers. It appears that after the 1930 - 1950s testing of hypotheses on the effects of forestry on streamflows, 1990s in re-examination should be the period of holistic research and an integrated approach to changing hydrological regimes due to human-induced changes.

The principal effect of forests on the hydrological cycle is the reception and disposal of precipitation. Forests provide the greatest surface area for the interception and re-evaporation of water and are effective traps for the adsorption of solar radiation.

Alteration of the ground cover was found to affect surface albedo and runoff, to change the ratio of sensible to latent heat transport, and greatly modify surface wind patterns. Therefore, continuous monitoring of the elements of the hydrologic cycle was cited as of paramount importance to determining magnitudes of global climatic and hydrologic change and for the understanding of the consequences of changing environment. A summary of the findings is expounded in the following sections.

2.8.2. Reasons for Excessive Human Interventions

Several factors were cited as the primary causes of excessive human interventions on the environment and in particular, the local hydrological regimes. Increased population and shortages for farm land, grazing, poverty, fuelwood, urbanisation, roads, towns, infrastructures, chemicals, deforestation and afforestation, conservation works, drainage of wetlands and control of natural bodies of water for irrigation are the major causal activities. Darkoh's (1993) assertions and hypothesis that poverty is the main cause of environmental degradation in tropical African nations put the picture more clearly:

" ....faced by rapidly rising population, recurrent droughts, inadequate and unreliable rainfall, hunger and famine, it is to the river basins that most people in drylands of Africa are turning for salvation. Given the food security situation in African countries, the vagaries of the climate, the continuing movement of people from drought-stricken drylands, and the escalating demands of irrigated agriculture, urbanisation and industrialisation, one question to be asked is whether African river basins can cope with the mounting pressure on their resources".

Many large-scale mechanised irrigation schemes, resettlement and large dams altered the historical and natural conservative farming systems in Africa. The large scale hydroprojects took away lands traditionally used by pastoralists during droughts to alleviate pressure on the fragile dryland environment. This was enhanced by the food security polices of African nations that justify expansion of large scale development hydroprojects without considering its environmental consequence this is a major contributing factor of human-induced hydrologic changes in the continent.

2.8.3. Changes in Hydrologic Regimes

What comes out more evidently is that watersheds and indeed large river basins are complementary ecosystems highly responsive to natural and man-made disturbances and ecological degradation that have potential repercussions on the watersheds. These can also exacerbate, as well, hostilities between various groups of the natural resource users in small catchment areas and extending from larger basins, to other countries and inter-country conflicts. In particular the hydrological changes cited are:

- Increased urbanisation reduces land occupancy, increases local water demand, increases surface and ground water exploitation, reduces infiltration and hence reduces river low flows, and enhances flooding in poorly drained cities.

- Deforestation changes vegetation cover that results in decreased infiltration rates and capacities of the soils, increased surface runoff and hence peak flows in rivers. The reduced infiltration reduces subsoil moisture storage and finally reduces recharge rate into the rivers. A prolonged moisture deficit leads to reduced low flows and drying up of small streams in semi arid lands; changed rainfall-runoff regime, response times, times to peak, concentration times and increased hydrologic droughts.

- Agricultural development and practices were found to have approximately the same effects as forestry. These included: increased water use rates reduces base flows; intensive cultivation accelerates soil erosion and degradation and increases river bed siltation; large scale clearing and burning reduces vegetation cover that changes the solar energy dissipation and distribution, hence increased temperatures, resulting in increased evapotranspiration and a changed micro-climate.

- The constructions of control, storage and conservation structures cause major adverse changes to the local hydrological regime. For example, the higher population concentration in the irrigation schemes, leads to increased water demands and hence conflicts in the overall management of water resources. The introduction of hydraulic structures changes the river's surface and subsurface flow regimes, changes the ground water generation mechanism, and ultimately changes the low-flow behaviour of the rivers; alters the rate of silt and sediment loads and deposition at lakes and sea shores, and encourages river channel and sea shore degradation and erosion. The overall hydrological changes are the changed micro and meso-scale climatic characteristics, habitat, and river flow regime.

2.8.4. Weather modification

This refers to all advertent and purposeful human activities that have the effect of changing the weather. Examples of inadvertent modification include that caused by urbanisation, deforestation, land clearing and creation of artificial water bodies. The purposeful weather modification includes all activities undertaken to alter locally unpleasant and disastrous variations of weather or to improve on a local basis. These include fog dispersal, hail, lightning and tornado suppression, frost prevention and artificial rainmaking. Modification of the precipitation and evapotranspiration patterns would have impacts on the other parts of the hydrologic cycle by influencing the magnitude and time of runoff, soil and groundwater storages and finally the streamflows. Except for experimental cases, purposeful modification of the weather is not widely practised.

2.8.5. Methods of Analysing Changes in Hydrological Regimes

This review has shown that, changes in hydrological regimes of an area can be assessed and studied with some degree of accuracy depending upon the purpose, extent and the availabililty of funds for the study. Selections of any of the assessment methods (analysis of time series, regional comparisons and transfer functions, experimental basin, and mathematical simulation methods) depend more upon the financial capability of the researcher. Developing countries are at a disadvantage in this case because their financial resources limit the amount, extent and quality of research. Use of a mix of the methods however, were recommended as satisfactory in a multi-disciplinary and applied research of this nature.

2.8.6. Rehabilitation of the Watersheds

In order to reverse the adverse trends of environmental degradation and rehabilitate the watersheds, several measures were advanced. Integrating individual research efforts into holistic concept of watershed response was emphasised. Several alternative systems exist which can maintain sustainable ecosystems and hence a viable hydrological regime in the watersheds:

- Developing ecologically viable agro-forestry systems in the rural sector and afforestation in the highlands. These will encourage infiltration rates and hence naturally recharging the groundwater reservoir that can sustain the water demand.

- Providing good agricultural farming practices that encourage soil and water conservation and at the same time increase agricultural yields to feed the increasing population and thus reverse the rural-urban out-migration.

- From a socio-economic viewpoint; provide a pattern of land use that incorporates production systems enabling the resource users to sustain an acceptable standard of living while protecting the resources.

- More integrated efforts are needed to permit wiser use of forest and land resources and help prevent large scale destruction of the environment by emphasising the importance of forest cover in proper management of soil and water resources.

- Adaptive research is needed to evaluate the performance of different land development techniques for soil and water conservation in diverse agro-ecological environments so that their effects on hydrological and ecological parameters can be quantified.

Admirable examples of the successful planning, development, and management of watersheds justify the above recommendations. The works at the Coweeta watersheds in the USA and the studies by Pereira (1961, 1962, and 1973) and (Edwards and Blackie, 1979 and 1981) in the East African catchments are but two clear illustrations in contrasting climatic situations. Hydrological effects were minimised because the developments were properly planned during the initial stages. Conservation measures were included that would hold the impact of water and energy balance to minimum.

Ecological and hydrological sustainability prevail in a natural vegetation with little disturbance, for this produces a soil surface capable of absorbing the rainfall and of allowing it to percolate into the deeper subsoil and seep out into the river as springs. These can be achieved by funding and implementing Agenda 21 action plans on natural resource conservation and development.

With all these done, however, the major solution in the future lies in the realm of rural and urban sociology and economy, since a recommended land use policy will only be willingly and effectively adopted by the land users if it increases the reliability and sustainability of their basic and legitimate demands with energy, water, food, shelter, and a clean environment for sustainable future.

On the basis of this literature review, the research topic and its extent was refined. The statements of the problem and objectives were consolidated and developed into hypotheses. Field surveys and data collection procedures were developed to quantify the formulated hypotheses.

CHAPTER III

VARIABILITY IN THE HYDROLOGIC SYSTEM

3.1 DEFINITIONS AND SCOPE

Hydrology is a term derived from the Greek words*hydro* meaning water, and *logos*, meaning science. In its broadest outlook however, hydrology is concerned with all water on the earth, its occurrence, distribution and circulation, its physical and chemical properties, its effects on the environment and life of all forms. It deals with alterations in the natural order of things. The US Federal Council of Science and Technology for Scientific Hydrology in1962 defined hydrology as:

"..... the science that treats of the waters of the earth, their occurrence, circulation, and distribution, their chemical and physical properties, and their reaction with the environment including their reaction to living things. The domain of hydrology embraces the full life history of water on earth".

Socio-economic developments raised standards of human life with a corresponding increase in water demands and hence its management. The fact that water is essential to life and that its distribution and availability are intimately associated with the development of human society means that it was almost inevitable that some development of water resources historically preceded a real understanding of its origin and formation. Thus hydrology as a science was developed fully after the realization that water resources depletion was leading to scarcity in the future.

3.2. THE HYDROLOGIC CYCLE

The interdependence and continuous movement of all forms of water provided the basis for the concept of the hydrological cycle. Since Meinzer (1942) this has been regarded as the central concept in the science of hydrology. It describes the circulation of water from the sea, through the atmoshphere, to the land and with delays, back to the sea by overland and subterranean routes, and in part by way of the atmosphere; and many short paths of the water returned to the atmosphere without reaching the oceans.

Fundamentally, this concept envisages that all water is involved in a cyclical movement which continues indefinitely. In one form or another, water occurs virtually everywhere, varying in quantity from almost unlimited supply in the oceans to nearly none in the desert regions. It occurs in the atmosphere as water vapour, clouds and precipitation. On the earth's surface it is found in streams, lakes and in the oceans.

Beneath the ground surface it occurs as springs and groundwater flows. The sequence of events, movement, and re-distribution of the water within the global cycle is presented in Figure. 3.1.

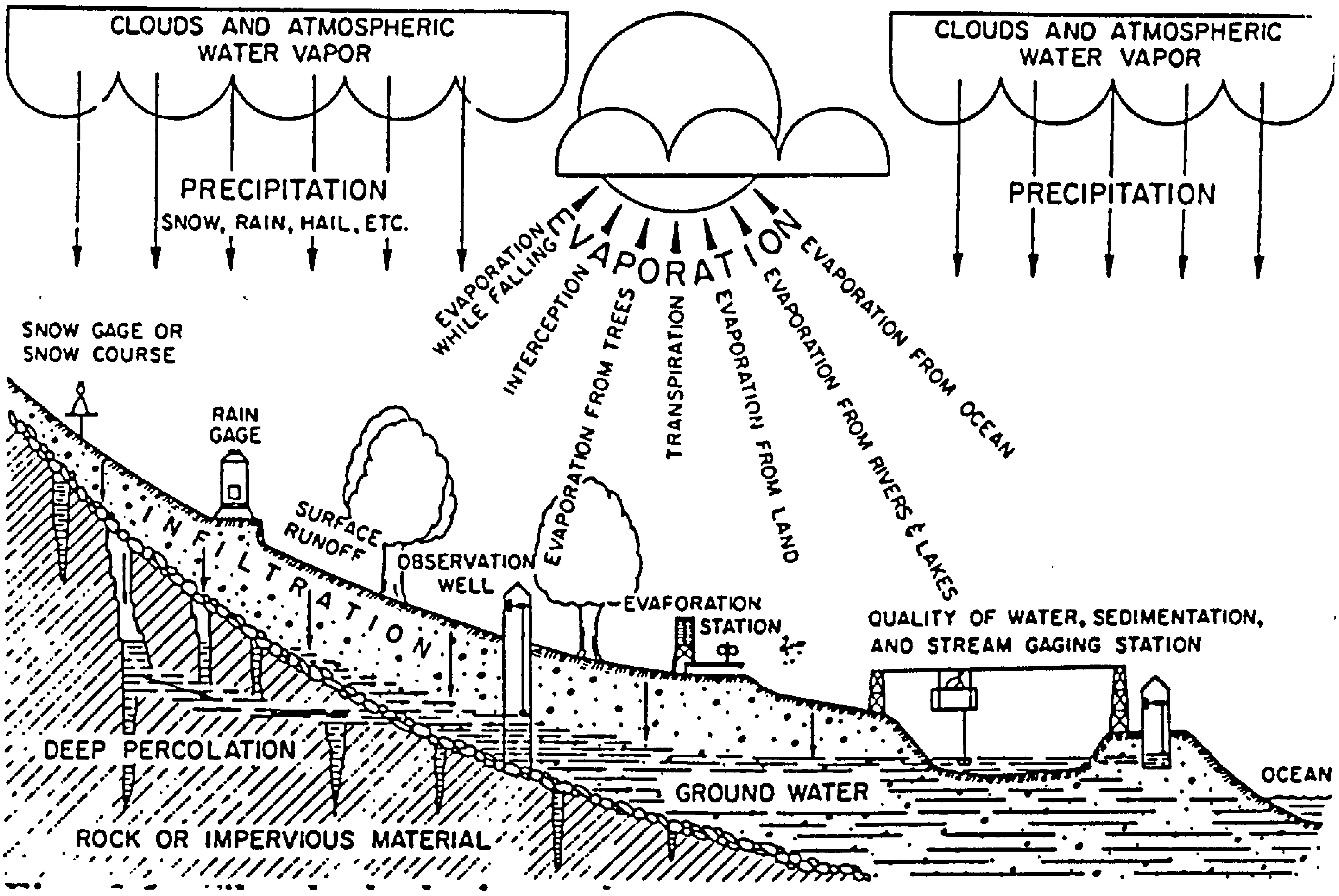


Fig. 3.1. The hydrologic cycle (after Meinzer, 1942)

At any instant, the largest portion of precipitation is stored in the oceans where evaporation is a continuous process. The moisture evaporated condenses and some falls directly on the ocean, a considerable portion is carried by winds or condenses as dew or frost on vegetation and other surfaces. The precipitation in the form of dew and frost is almost evaporated directly or consumed by vegetation and then transpired. A portion of is re-evaporated before it reaches the ground surface. Another part is intercepted by vegetation, buildings, and other objects and part is re-evaporated. The other portion may first remain on the surface as storage, and surface moisture and eventually evaporated into the atmosphere. It may flow into depressions and channels to become streams and lakes from which it may be evaporated back or seep into groundwater storage or flow on to oceans.

Finally, the precipitation may infiltrate into ground as subsurface soil moisture. This portion may be removed either by evaporation and transpiration from the soil and vegetation surfaces, or by interflow towards stream channels, or by downward deep percolation to become groundwater. The groundwater component eventually is returned by evaporation and transpiration into the atmosphere or by seepage into surface streams or by inflow to the oceans. Based on the circulation and re-distribution, the hydrologic cycle can be studied on various scales from global through continental, regional, national, and sub-national scale to the watershed scale where physical processes are examined in greater detail in Figure 3.2.

This rather smooth, uninterrupted, and sequential movement of water is however belied by the complexity of natural events (Ward and Robinson, 1988). The movement is short-circuited when precipitation on reaching the ground surface is evaporated immediately into the atmosphere without contributing to streamflow, soil moisture, or groundwater movement or to the oceans. Similarly, precipitation falling on lakes may be evaporated without reaching the land surface, or that falling upon the land could percolate into aquifers and take several years to contribute to springs and the oceans.

Secondly, there is an irregularity of the movement of water at global scale. Rainfall may occur one in ten or twenty years in deserts, while the other phases of the cycle such as evaporation and surface runoff may take a short period after a rainfall, or the slow redistribution of the groundwater. Further evidence of this irregularity is more pronounced in the subpolar regions, where precipitation in the form of snow, may take several months between the rainfall event and the active involvement in the complete hydrologic cycle. At small watershed scale, the hydrological processes are rarely operated completely uninfluenced by human activity. Hence, a detailed understanding and quantification of the movement and distribution of water in the hydrological cycle should be a prerequisite to the identification of points of human impact. The systems approach in most cases is used to analyze and synthesize the human interventions in the watershed.

3.2.1.The Watershed as an Hydrologic System

Hydrological phenomena are extremely complex. Hydrologists have represented them simply by using a systems concept. Since a system is a set of connected parts that form the whole, it can be used to synthesize and locate points of human impacts in the hydrological cycle (Figure 3.2). The watershed can then be defined as a structure or volume in space, surrounded by a boundary, that receives precipitation and other inputs. It operates on them internally to produce outputs in the form of evapotranspiration and streamflows. Since human beings are part and parcel of the system and use its resources, the term human-modified hydrologic system can be used to define that new system that has undergone human modification, and its operations reflect the human element and actions in totality.

In this way, hydrologists are able to differentiate locations that have changed, to analyse the inputs, and outputs and hence draw conclusions on transformations on the watershed. The elements of the hydrologic system which include precipitation, evapotranspiration, streamflow and other phases of the hydrologic cycle can also be grouped into subsystems of the overall cycle. To analyze the total system, the simpler subsystems are treated separately and results combined accordingly to interactions between subsystems. The basic theme and application of the concept was concisely stated by Wicht (1971):

" The catchment of a natural system... is an ecosystem in which the living organisms and their non-living environment are reciprocally related in an energy cycle...If the dynamics of water in the catchments are to be adequately understood, so that they can be managed to improve water yields, the ecosystem must be analysed into its component parts and the interactions between these investigated to gain a holistic conception of the catchment".

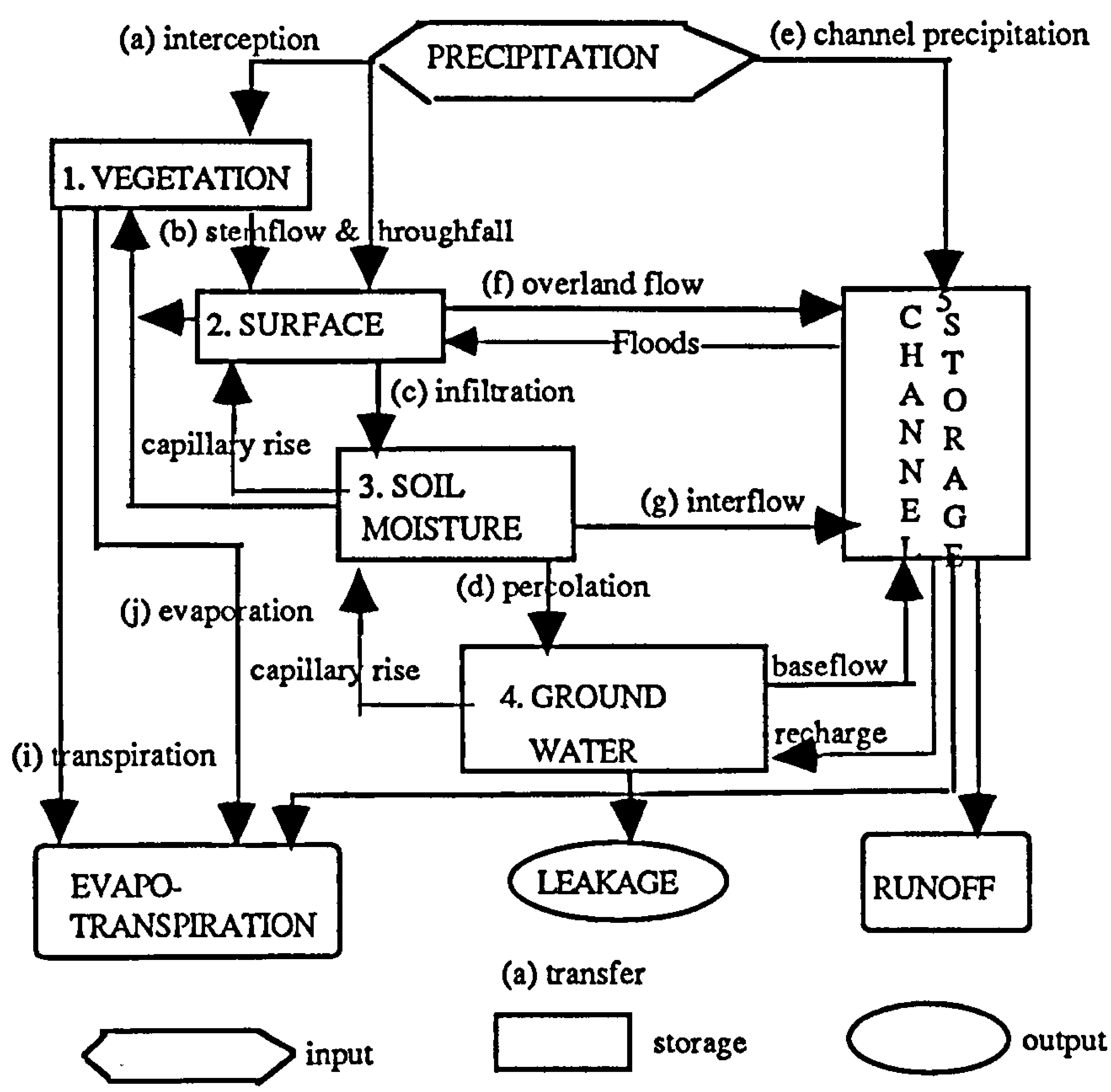


Figure 3.2. The system concept of hydrologc cycle representation at watershed level (from an original drawing by Lewin J., UCW, Aberystwyth, and quoted by Ward and Robinson, 1988)

The approach explains how and why events of different extent and magnitudes occur in the watershed. An hydrologist thus operates in a broad field of environmental conservation. Building on a sound understanding of the classical hydrologic cycle, effects of human development are incorporated in the system and interpreted as a whole. This means that legislation and its application, economic evaluation and interpretation in terms of hydrologic modification, can be carried out in order to predict the outcome of man-made environmental changes.

In the process of utilizing water and other resources in the watershed, the reflective properties of the area are changed which changes the total amount of heat retained in the global system, or interbasin transfer of water resources. This re-distribution has profound effects on the regional and local climates. This therefore implies that human activities on earth can influence the very definition of this subject. The importance and points of influence in the hydrologic cycle are precisely stated in Dooge et al. (1973) as:

...." so long as man was a relatively insignificant element in the whole balance of nature, his power to disturb this balance was small. Now, however, his technology is being so powerful that he can bring great changes on continental and even global scales, and the need to foresee the consequences of his actions has become a condition for survival"

Figure 3.3 summarizes the principal points of human deliberate attempts to modify the natural hydrologic system. The systematic representation shows that human interventions can occur practically everywhere and possibly at every element of the hydrologic cycle. The most important modifications are related to the following: (1) large-scale surface changes in terms of deforestation and removal of vegetation cover, which affect the surface runoff and hence modify channelflow and storage, (2) extensive irrigation and drainage, (3) construction of large dams and other control structures which influence concentration and peak times and (4) a massive abstraction of surfacewater and groundwater for domestic and industrial use. As water demands continue to rise, artificial stimulation of precipitation and groundwater recharge is likely to be reconsidered after a lull period since the late 1950s. Other inadvertent modifications however, may include interbasin water transfer, the use of transpiration and evaporation suppressants, controlled runoff and roof water harvesting.

The basic premise of the systems theory, that all things have connections of many other things and that the significance of any one depends upon its relations with others have long been adopted in hydrology (Domenico,1972). Three advantages of using systems approach in hydrology are widely acknowledged: Firstly, systems theory provides a valuable integrating framework which brings together in an orderly way a variety of theories, explanations, and mathematical methods that may otherwise have no underlying organization (Chow, 1969). Secondly, in a research context, it requires the research worker to look at the whole problem instead of adopting a piecemeal approach (Ward and Robinson, 1988). And thirdly, it makes it easier for the hydrologist to build simulation models whose behaviour approximates the real system and which are amenable to a mathematical analysis and solution (Domenico, 1972.Clarke, K.F., (1994)-It is the third and most important factor which has opened up a new field of investigation and modeling human-induced watershed changes on hydrological regimes.

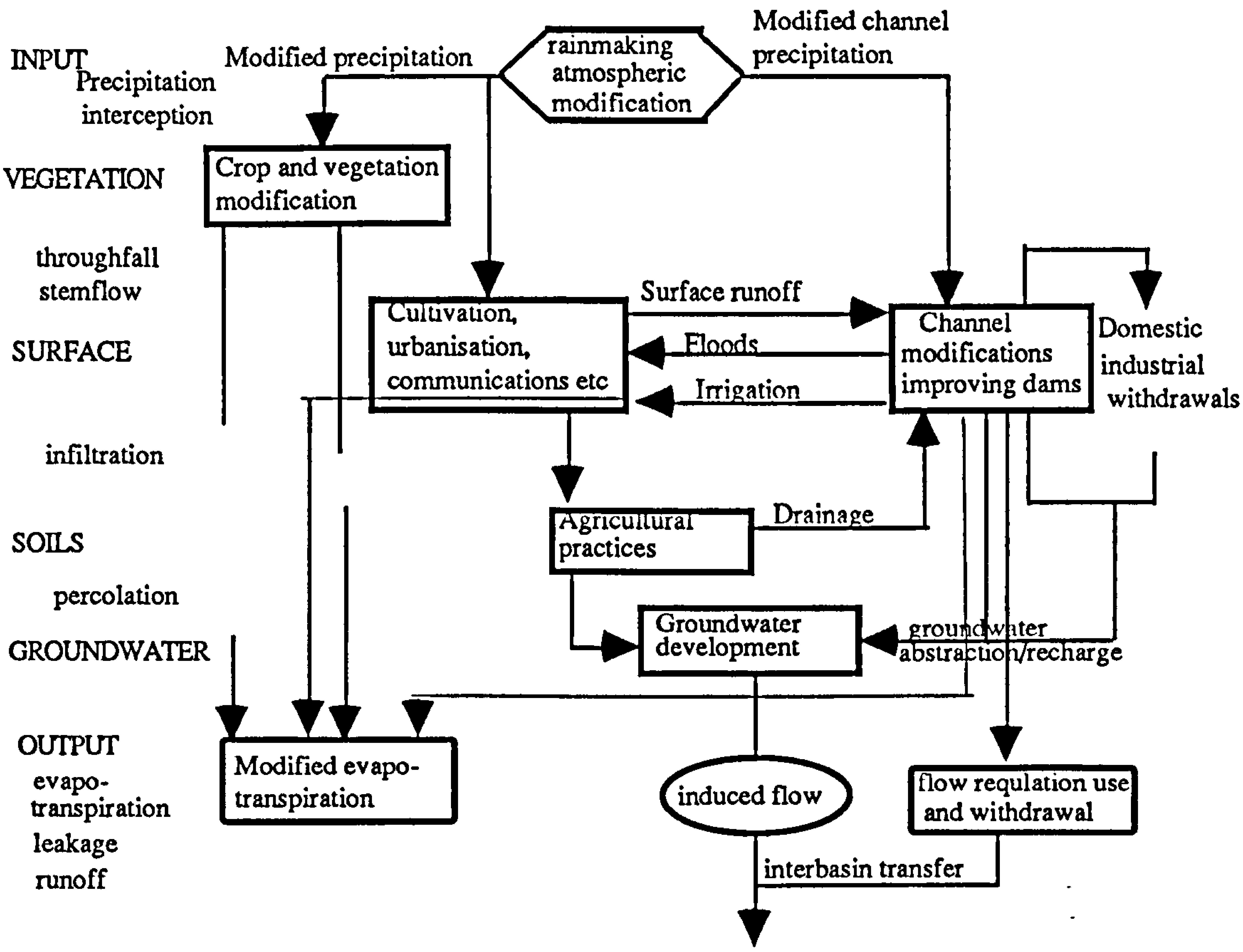


Fig.3.3.Principal points of human intervention in the hydrologic cycle (after Ward and Robinson, 1988 and Chorley, 1969)

3.3. VARIABILITY OF THE ELEMENTS IN THE HYDROLOGIC CYCLE

This section examines the spatial and temporal variations of the hydrological elements at global, tropical Africa and research findings in East Africa and Kenya, the location of the watershed under study. The hydrologic elements are affected by human activities directly or indirectly. In extreme cases, the consequent environmental changes includes effects upon channel pattern, riparian, floodplain, and delta or estuarine, fauna and flora (Petts, 1984). The impacts of these activities are assessed in several ways, including the examination of the destruction of amenity or recreational environment or by examining changes in the components of the water balance. Inspite of the difficulties in measuring changes in areal evapotranspiration, the water balance approach still presents a framework to guide the analysis.

Traditional land users in Africa for example have always adapted well to changing environmental conditions but human-induced changes in the hydrologic system, have weakened this adaptation. The changes made it more difficult for the population to adapt to the harsh conditions such as a series of dry years (Oguntoyinbo and Odingo, 1979, Jackson, 1988). Impacts of unplanned land use tended to be greatest in marginal areas, making the residents more vulnerable to rainfall variability and accompanying effects of droughts. Thus, assessing the variability of the hydrological elements at watershed level becomes complicated given the holistic approach to sustainable development recommendations in the UN Agenda 21 (UNCED, 1992). The following hydrological processes are examined in terms of: rainfall; evaporation, evapotranspiration, and streamflows and watershed characteristics.

3.3.1 Temporal and Spatial Variation of Precipitation

Water moving on land surface is derived either directly or indirectly from precipitation. The points of occurrence, amounts, and their distributions are required in order to effectively plan and design appropriate conservation measures. Its periodicity and variability is required for easy isolation of human interventions. Detailed hydrological aspects of precipitation are required by tropical land use planners because, the rainfall variability, climatic variation and the controls and origins of rainfall are extremely complex. The general circulation of the atmosphere, disturbances, and local factors in these regions further combine to produce extreme rainfall variability. Accordingly the literature on the variation of precipitation is reviewed in this perspective.

3.3.1.1. Historical Rainfall Variations

Temporal variation of rainfall in a given location is described by observing its long-term secular, periodic and stochastic patterns. The secular variations incorporate cyclic, non-cyclic or trend characteristics. The existence of cyclic variations on annual basis is useful for forecasting flood events and water planning. However, attempts have been made without success, to find cyclic variations in annual precipitation totals (Ward, 1975., Ayoade, 1988., Ward and Robinson, 1988, Jackson, 1988). Many cycles have been reported but few have been conclusively demonstrated (Farmer and Wigley, 1985), although the concept of a 35-year cycle, known as the Bruckner cycle, is widely accepted but without adequate proof. Analysis of thousands of annual precipitation and runoff series for more than 150 years by Yevjevich (1964) and Vines, (1985) failed to indicate any significant periodicities or trends. Changes in global winds and moisture fluxes related to sea surface temperature anomalies have been suggested as a possible cause of the failure in rainfall in recent years in the Sahel region of Africa (Folland et al., 1986, Ward and Robinson, 1988).

3.3.1.2. Rainfall Variation In Tropical Africa

Attempts have been made to analyse rainfall time series in East Africa to investigate past climatic conditions and their temporal variability. Notable works of Rodhe and Virji (1976) and Ogallo (1980) showed no statistically significant trends on the regionally averaged rainfall series, as well as in the monthly and dry season records, concurring with Lema's (1990) observation of no significant trend in East Africa rainfall during the 1880- 1980 period.

Studies of rainfall in Africa by Ayobande (1973), Ogallo, (1984), and Farmer and Wigley, (1985) using spectral analytical techniques were still unsuccessful. Several investigations have however emphasized that many of the non-cyclic secular variations of precipitation are caused directly by a combination of geographical and climatological factors. Studies of secular rainfall variations by Klaus (1978) in some temperate and tropical areas concluded that, the variations were largely the result of dynamic changes in general circulation of the atmosphere at these localities. These and similar investigations indicate that local geographical and general climatic factors are the causes of the variations.

Extreme periodic interannual rainfall variability, dominate the arid and semi-arid African continent where severe droughts are common. The meteorological causes of these droughts are still unclear, but numerous studies suggest major changes in the large-scale tropical circulation patterns are involved (Kruerger and Winston, 1975, Kidson, 1977, Kanamitsu and Krishnamurti, 1978). The rainfall pattern is characterised by the occurrence of a few extreme high values and a skew distribution in some stations (Jackson, 1988) as given in Table 3.1. Superficial examination of these values supports the concept of high variability between their monthly minimum and maximum averages.

An important aspect of variability is the tendency for sequences of wet or dry periods to occur. For example in Table 3.2 the occurrence of a sequence of dry years at stations in Tanzania is illustrated. Similar pattern is observed in Mali in West Africa. At Gao in Mali, Oguntoyinbo and Odingo (1979) reported that over the period 1966-1973, annual rainfall as a percentage of the 1931-1960 mean ranged widely from 55 to 97% in individual years. Similar patterns have also been reported by Swift (1973) for the Sahel. It is the tendency for such sequences to occur, perhaps more than the magnitude of departure from average conditions which has serious implications.

Table 3.1. Rainfall range for some stations in Africa (Griffiths, 1972).

Station & Country	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
Mogadiscio, Max	9	3	257	245	324	349	240	182	239	192	179	76	997
Somalia Average	1	0	9	58	56	82	58	40	23	27	36	9	399
Min	0	0	0	5	7	13	13	10	5	4	4	1	57
Harbell Max							1203						
Liberia Average							461						
Min							123						
Meru (Kenya)							Oct	ober ra	nge 1	386 -	5		
Quelimane Mozambique							A	nnual r	ange	530 -	2501		
Maintirano Madagascar							An	nual ra	nge 4	5-240	% of	mean	

Table3.2. Sequence of annual rainfall in some stations in Tanzania (Jackson,1988).

Narok Forest Station Year	1936	1937	1938	1939	1940	1941		
average, 1811 mm Total	1389	1202	1536	1429	1539	1428		
Mvumi Mission Stn Year	1948	1949	1950	1951	1952	1953	1954	
average, 544 mm Total	360	374	391	576	377	459	340	

However, Nicholson and Chervin (1981) studied in detail the spatial and temporal characteristics of rainfall fluctuations in Africa and suggested the major rainfall fluctuations were associated with factors modifying the intensity or frequency of disturbances. Figure 3.4 from Nicholson and Cherwin (1981) studies show negative regional departures of rainfall dominating the entire continent in the 1970s compared to 1950s where some individual years with positive departures prevailed over almost the entire continent. This may support the historical observations of large inter-annual variability in the absolute amount of rainfall over the continent. No conclusions however, can be drawn for tropical Africa because there may be equaly large sectors elsewhere where anomalies opposite to those over the continent prevail. Nicholson and Chervin (1981) suggests further detailed investigation.

Spectral analysis method is used to examine area-average or single-site rainfall series. It identifies periodicities, or cycles, present in the rainfall data. Nicholson and Cherwin (1981) study presented in Figure 3.6 shows the results from this method when applied to 84 African rainfall regions. Statistical significant peaks are common at periods of approximately 2.3, 2.8, 3.6, and 5 to 6 years.

However, a general observation can be made that the West African region exhibit a few significant peaks below 8 or 10 years which is in agreement with the spectral analysis results from Bunting et al. (1976). Ogallo (1978, 1979) using the same method on 69 individual stations in Africa found cycles of 2 to 2.5 and 2.7 to 3.3, while Rodhe and Virji (1976) found 2, 3.5 and 5 to 5.3 for East African rainfall. In contrast, Puterbaugh et al. (1983) found no statistically significant oscillations in these regions. Kraus (1978) using Maximum Entropy Spectral analysis found a nearly 3-year period dominating in West African rainfall fluctuations. In an attempt to reconcile these differences, Farmer and Wigley (1985) computed area-averages for the nine regions in Figure 3.6. The results are presented in Table 3.3. These findings agree with those of Kraus (1978) for West Africa (regions 1 and 3) and Rodhe and Virji (1985) for East Africa (region, 7). It is therefore clear that the rainfall data show some periodicities, but their significance is difficult to assess. Given this difficult and the widely varied results obtained it is equally difficulty, to confirm the findings.

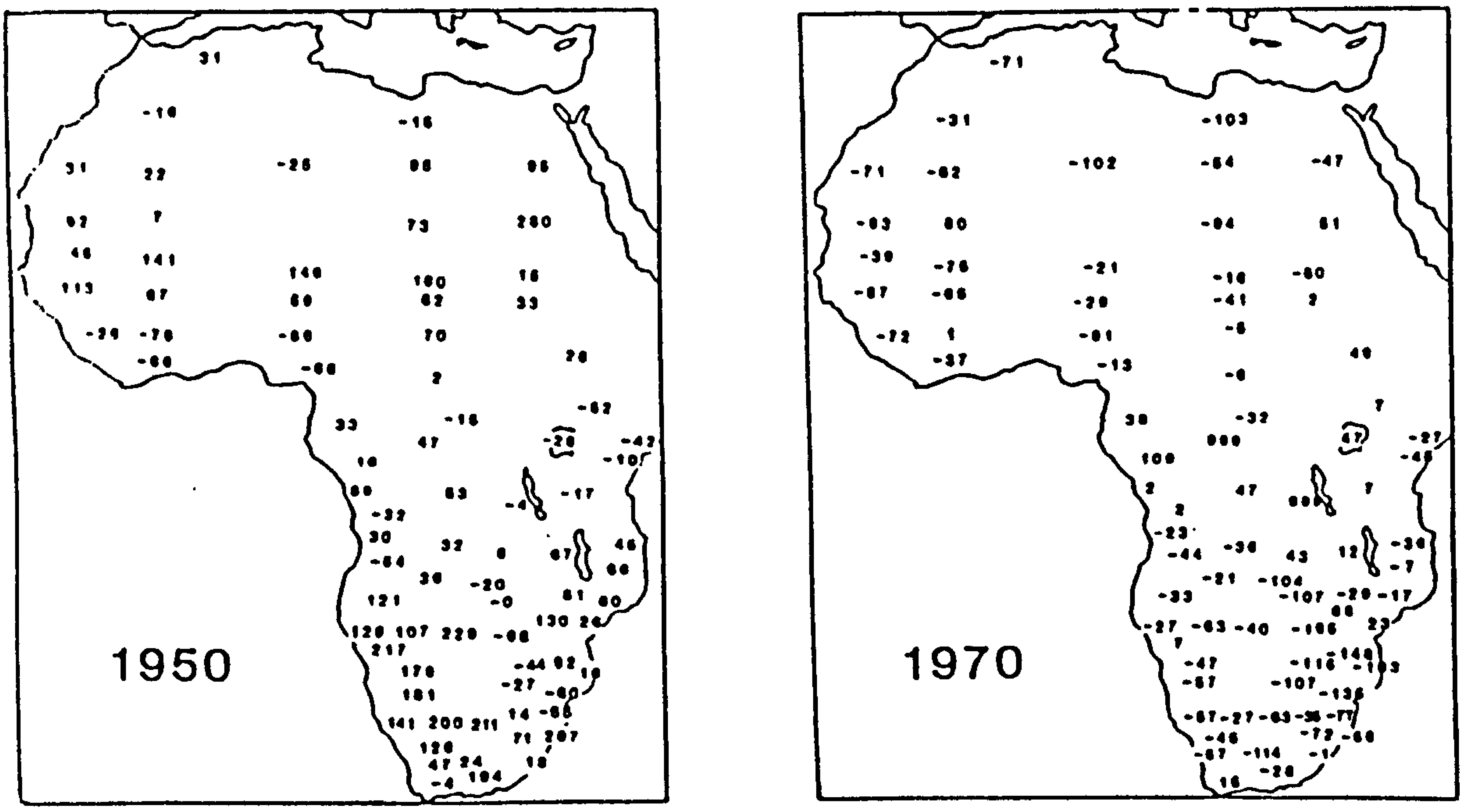


Figure 3.4. Regional rainfall departures from long-term mean for 1950 and 1970 in African continent (after Nicholson and Chervin, 1981). The numbered regions which divide the area are those used in the analyses of area averages (Farmer and Wigley, 1985).

Trend analysis is another technique used to examine rainfall variability. Linear trends of African rainfall reported in Sansom (1952) for the period 1920-1949 found a pattern of increases and decreases which imply a lack of a significant trend. Similar findings were later found by Rodhe and Virji (1976) in their analysis of annual rainfall records for East Africa. Ogallo (1978, 1979) analysed trends in annual rainfall data in the African continent and found none with significant increases. Winstantley (1985) examined longer time scale rainfall variations for Gambia and elsewhere in West Africa and found decreasing trends from Gambia to the west of Sudan and Ethiopia dating back to around 1880. Farmer and Wigley (1985) however cautions on the wholesale acceptance of these reported trends because of inadequate evidence for any century time scale rainfall trends in the continent.

Table 3.3. Significant (5% level) and other noticeable spectral peaks for African rainfall in the regions in Figure 3.7 (Farmer and Wigley, 1985).

Regional Number	Significant Periodicities (yrs)	Other Periodicities (yrs)
1	3.1	2.4, 45
2	-	7.5, 32
3	2.7	3.9, 6.1, 32
4	2.5, 8.0	3.5, 25
5	7.1	2.7, 3.0
6	3.5, 7.1	-
7	3.4, 5.3	-
8	-	2.7, 5.9, 7.1, 12
9	-	2.2, 2.7, 5.9

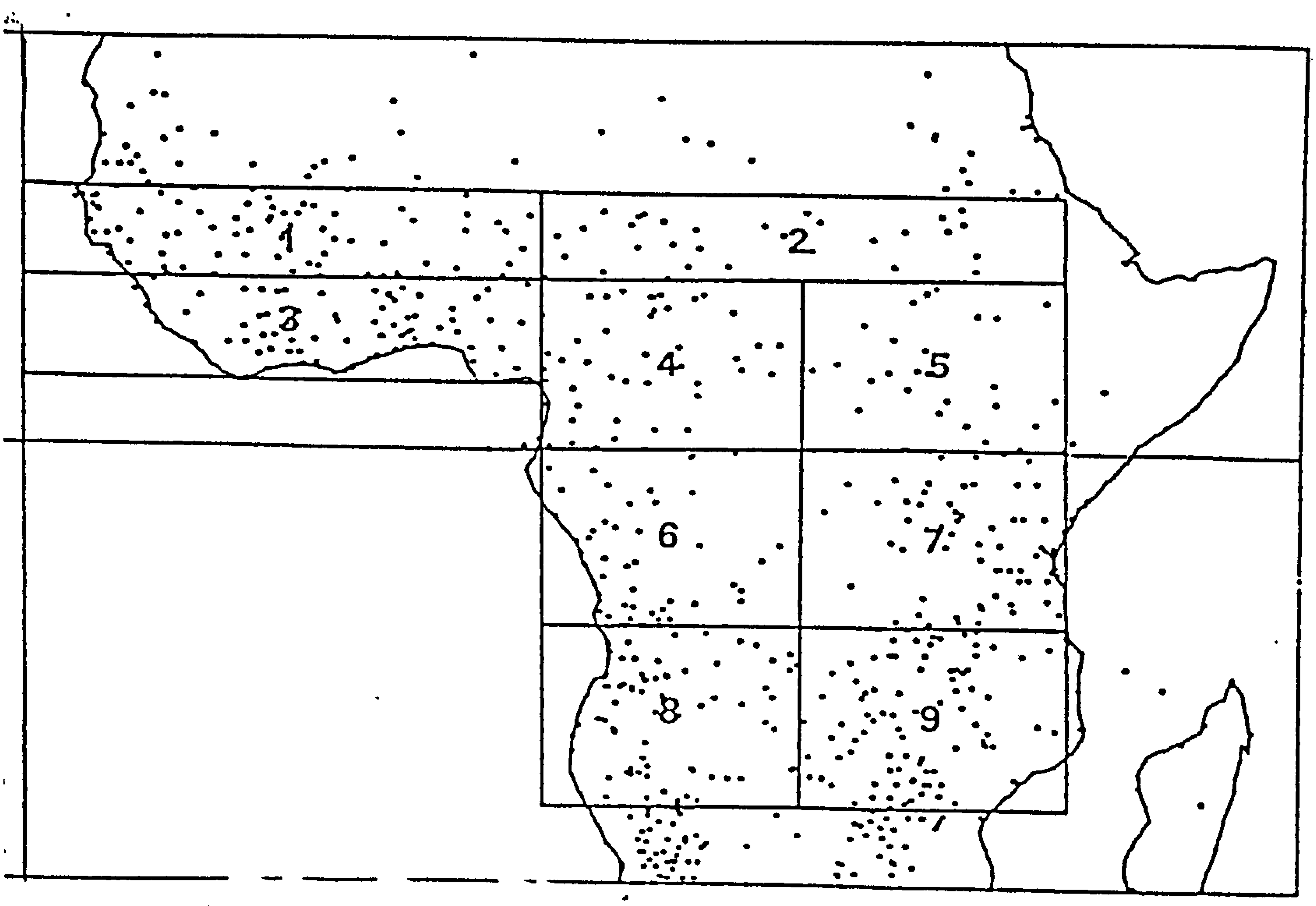


Figure 3.5 Region number (after Maley, 1981) as reported in Farmer and Wigley, (1985)

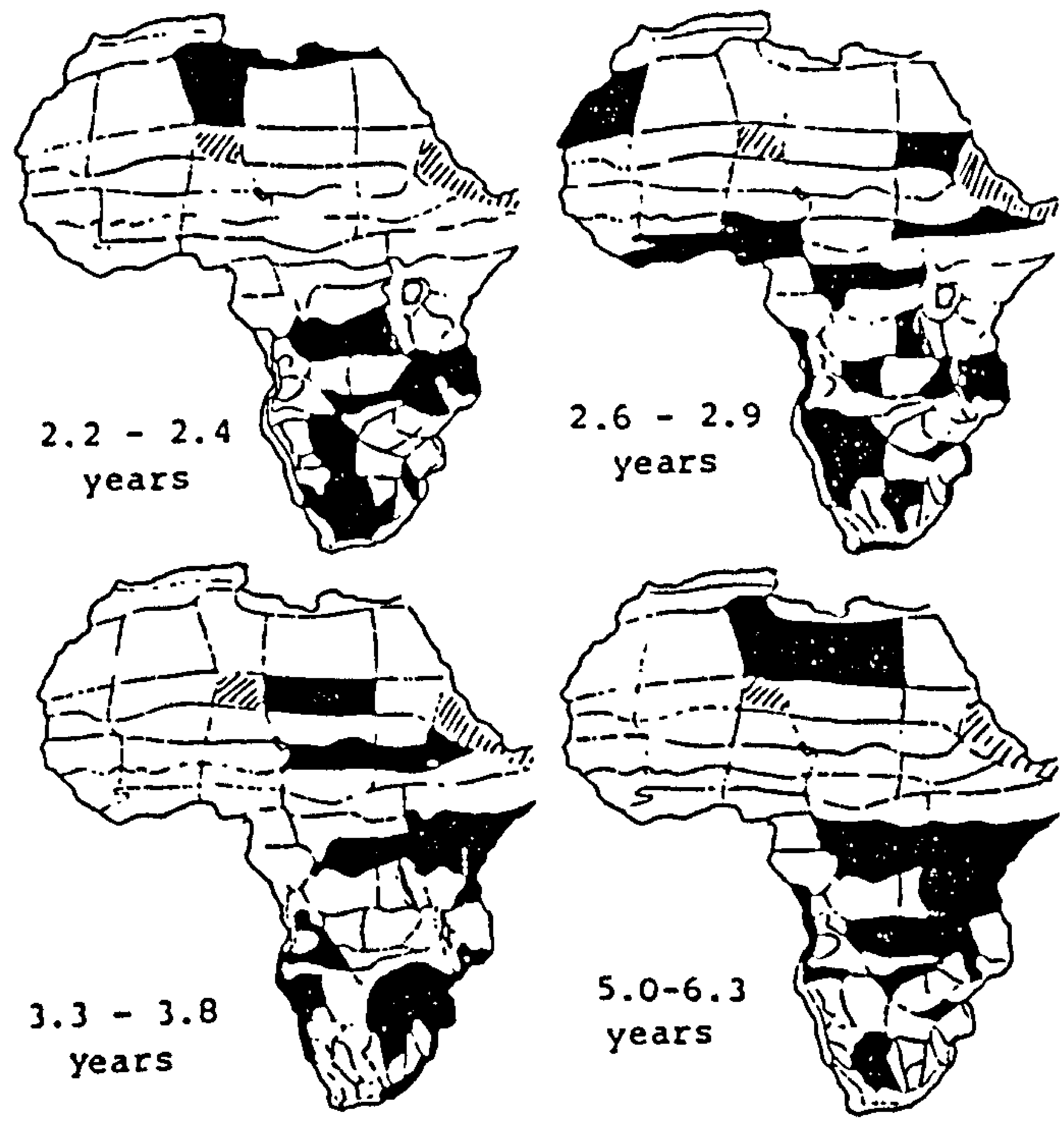


Figure 3.6 Distribution (solid shading) of significant spectral peaks in rainfall (after Nicholson, 1985). Hatched areas indicate no data. The overall significance of these results is difficult to judge because of the statistical problems of multiplicity and spatial correlation (Farmer and Wigley, 1985).

Persistence has however, been observed in the rainfall data in Africa. Nicolson (1983) found a strong persistence in Sahelo-saharan rainfall series but no persistence in the Kalahari rainfall series. Nicholson(1983) found autocorrelation coefficients for 1-, 2- and 3-year lags as 0.33, 0.28 and 0.26 respectively which are significant, partly confirming the existence of persistence in the data. Farmer and Wigley (1985) carried out analyses of persistence using area-averages and found similar results.

Tropical Africa had a 15-40% above average increase in rainfall in 1950s but deficits during the 1968-1973 droughts virtually matched the excesses of the 1950s and the downward trend continued from the 1970s into 1990s. For example, the past decades were extremely dry in Sudan (Figure 3.7). With the exception of 1978, western Sudan had one prolonged drought since the mid-1960s. The lowest recorded rainfall since the turn of the century occurred in 1984 in virtually the whole of the continent (Farmer and Wigley,1985, Olsson, 1983). Olsson (1993) analysed the rainfall distribution in the same stations (Figure 3.8) for the droughts of 1973 and 1974 and observed similar patterns which agreed with the findings in Nicholson (1978), Lamb (1982) and Hulme, (1993).

Olsson (1983, 1985) observed the daily variation of rainfall in the Sudan from 1950 to 1956 presented in Figure 3.9. The number of rainy days had more negative anomalies in the last decades than the amount of rainfall. The rainy days show a consistent pattern of unfavourable rainfall since the mid-1960s, and the trend is more consistent than in the case of the total rainfall. The significant decrease in rainfall since the mid-1960s is the main factor controlling vegetation growth in general and water yields in particular. Researchers have not established how much of the observed variation in vegetation conditions in land degradation, results from adverse human impact?. Once this relationship is established, it will help explain the variation of streamflows in the locality.

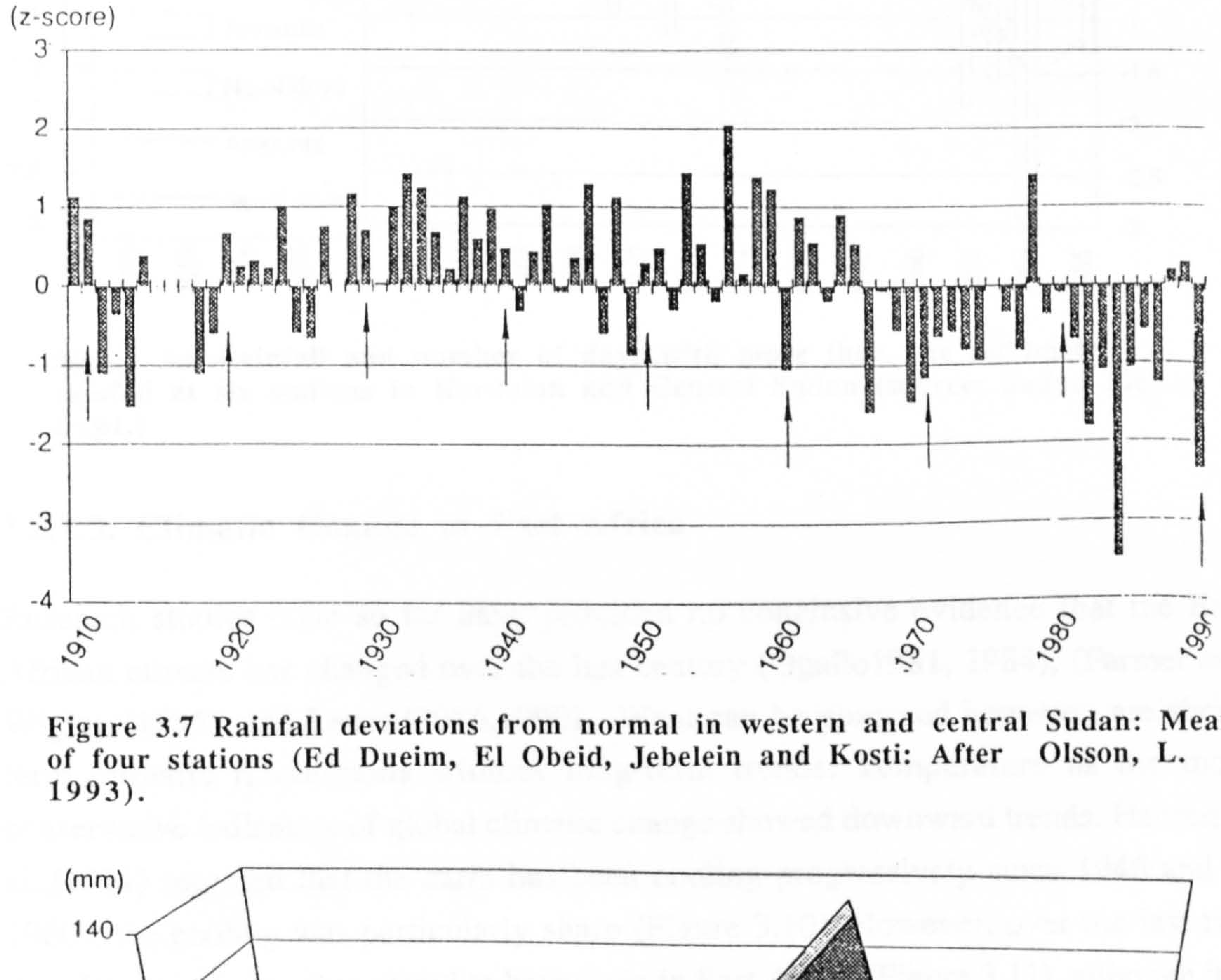


Figure 3.7 Rainfall deviations from normal in western and central Sudan: Mean of four stations (Ed Dueim, El Obeid, Jebelein and Kosti: After Olsson, L. 1993).

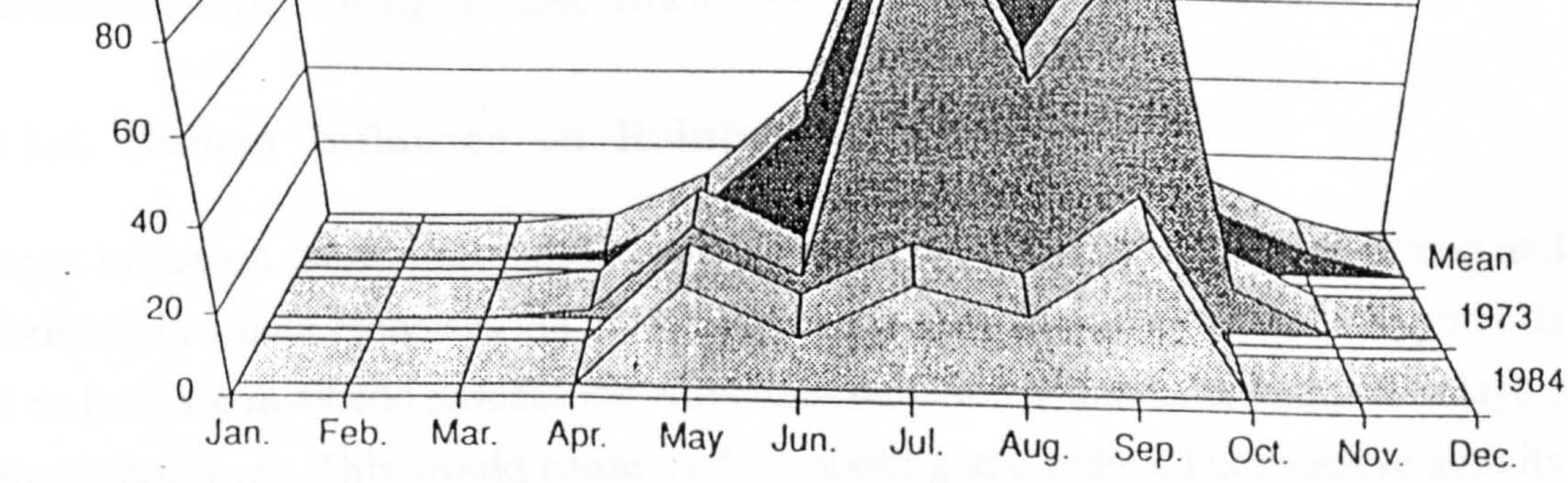


Figure 3.8. Rainfall distribution in 1973, 1984 and the mean for the period 1909-1990(After Olsson, 1993) in Sudan

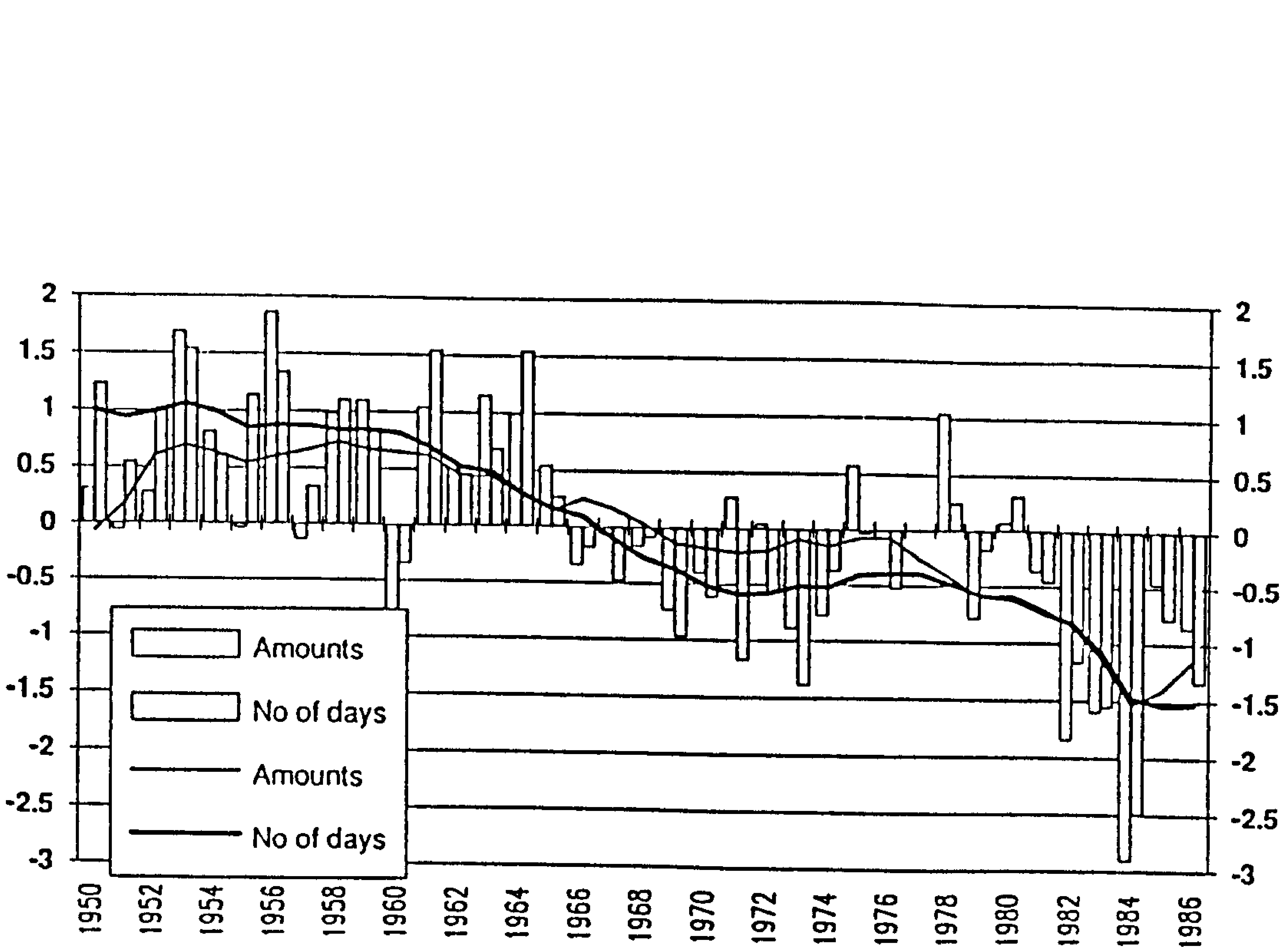


Figure 3.9 Rainfall and number of days with more than one millimetre rainfall at six stations in Kordofan and Central Sudan (source: Sudan Met. Dept.)

3.3.1.3. Climatic Change in East Africa

Research studies done so far have provided no conclusive evidence that the East African climate has changed over the last century (Ogallo1981, 1984), (Farmer and Wigley (1985), and Lema (1986,1990). What can be observed however, are short-term climatic fluctuations without long-term trends. Temperature as the most conservative indicators of global climatic change showed downward trends. Hansen et al.,(1981) reported that the earth has been cooling progressively since 1945 and in 1960s, the cooling was particularly sharp (Figure 3.10). However, over the last two decades, temperatures appeared to have risen in East Africa (Figure 3.11), although it is too early to describe the future fluctuation patterns (Lema, 1990). Coupled with the lack and inadequacy of data, it can be concluded that temperature change alone is not a good indicator of climatic change in East Africa.

3.3.1.4. Human Influence on Rainfall

Human influence on rainfall deficits in Africa is mainly linked with overgrazing and deforestation which increases the albedo. Otterman (1974) put forward the hypothesis that an increase in albedo reduces the amount of radiation available to heat lower layers of the atmosphere. This would cause surface cooling and reduced convective activity with less rainfall, thus completing a feedback loop of : less rain implies less vegetation implies greater albedo implies less rain. Charney (1975), modified Otterman's hypthesis to: higher albedo changes the heat balance of the surface-atmoshpere system and the large-scale changes in heat sources and sinks result in increased divergence in the lower atmosphere and reduced uplift over the higher albedo region. These changes in turn, should lead to less rainfall and a maintenance of drought conditions (Rasool, 1984).

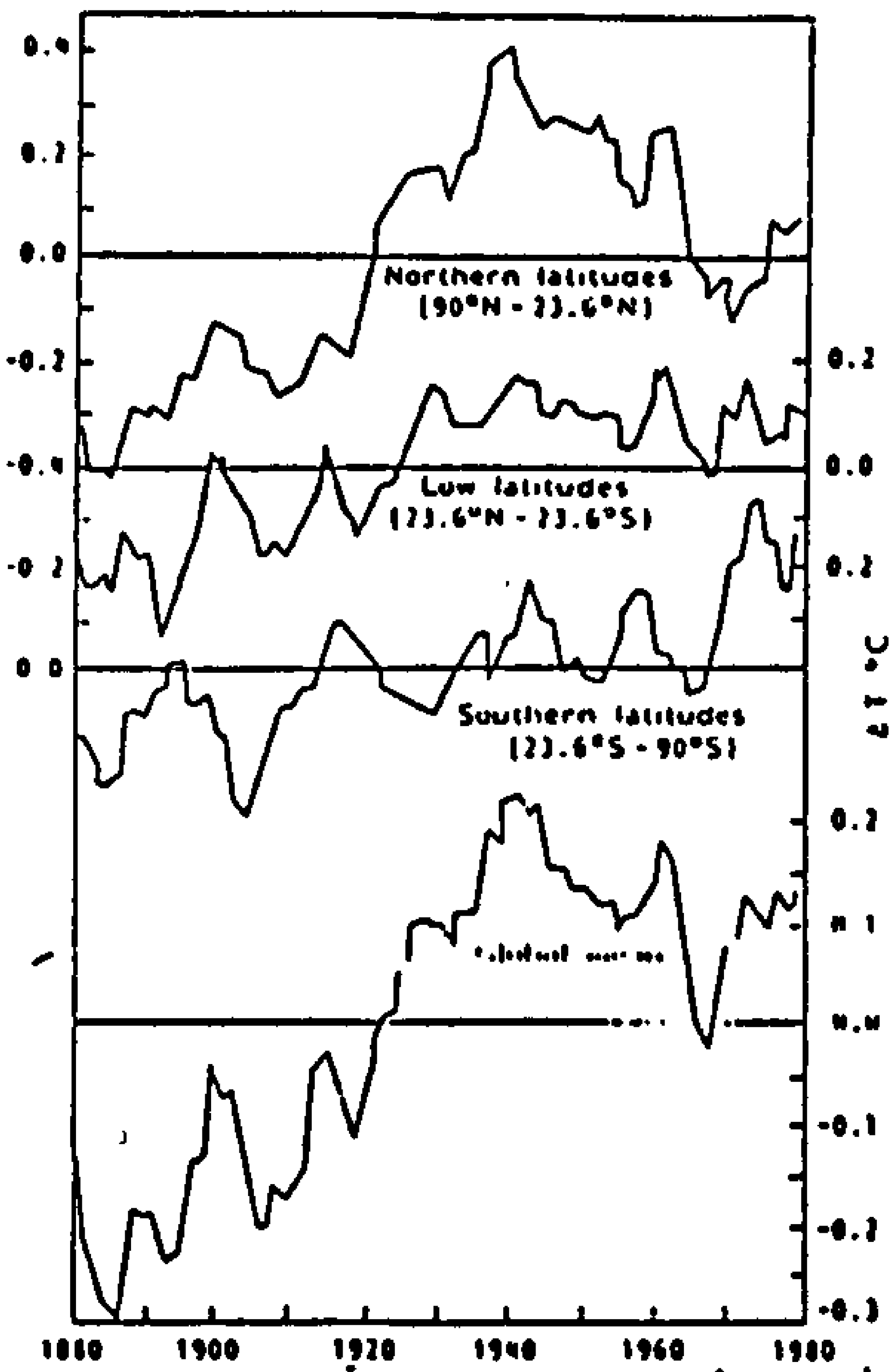


Figure 3.10. Observed surface temperature trends (after Hansen et al. 1981).

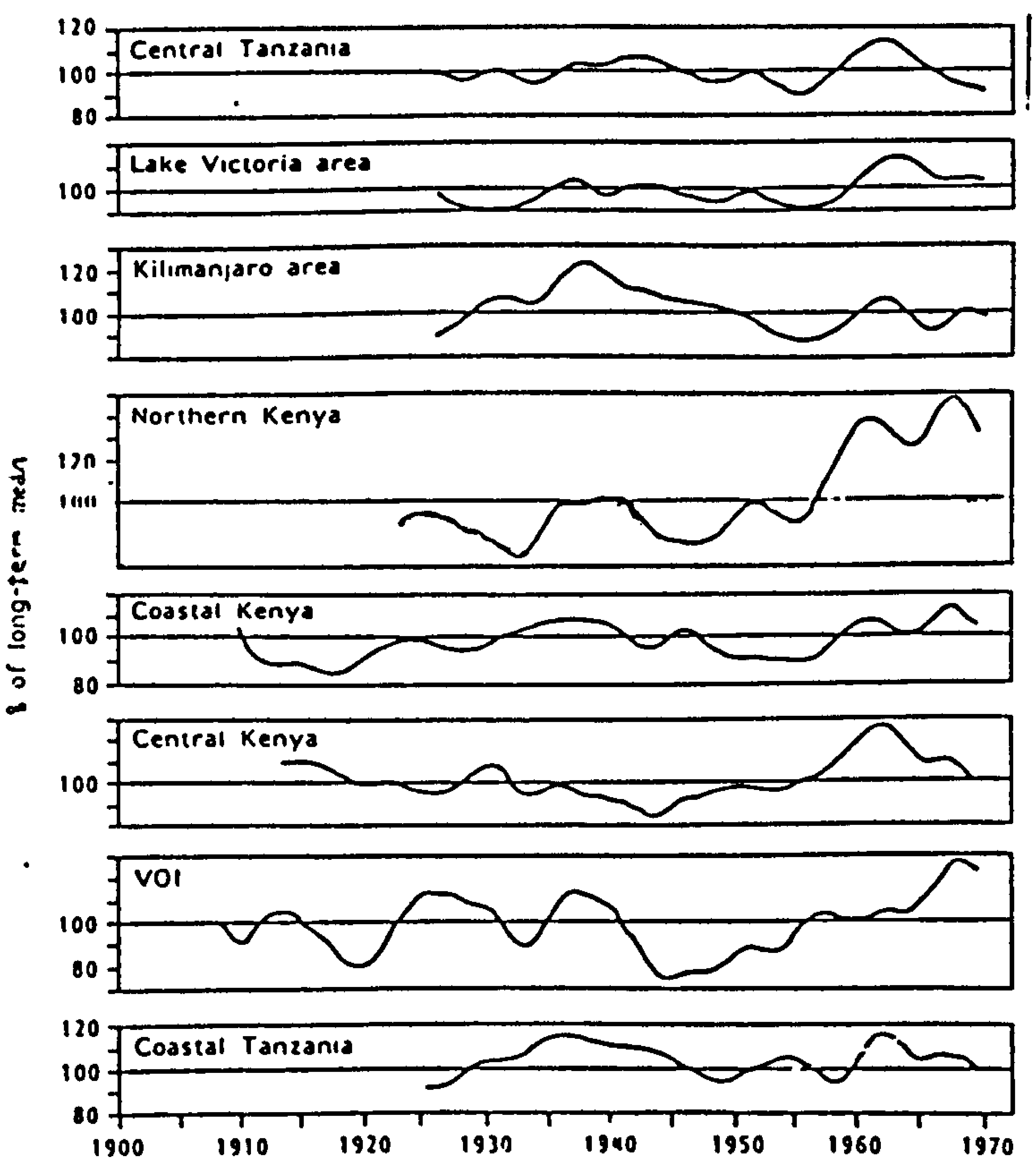


Figure 3.11 Smoothed rainfall series in East Africa (after Rodhe and Virji, 1976).

Computer modeling and experimental works were used to test these hypotheses (Nicholson, 1981). The results showed the atmosphere to be sensitive to land-surface evapotranspiration; so that changes in the available soil moisture or changes in albedo produce large changes in simulated climates. Research was then concentrated on establishing the relevance of the experimental results to a long-time scale changes in the African sahel zone. Satellite evidence for the sahel showed an increase in mean dry-season albedo from 0.29 in 1967 to 0.34 in 1973, and the mean wet-season albedo increased from 0.23 to 0.33 over the same period (Norton et. al., 1979), demonstrating a relatively weak negative relationship between the albedo and surface wetness. These results suggest that the changes are relatively small when averaged over large, continental-scale areas. However, although the physical realism of the albedo mechanism has been established, its relevance to drought remains controversial and not conclusive (Farmer and Wigley, 1985).

3.3.1.5 Summary

What emanates from the review is that the variations of rainfall in most regions of Africa over the last century and particularly, the last three decades resulted from factors which are still not fully understood and documented statistically. Hence it is not yet possible to make an objective forecast of the future regional climate in the continent.

There is however, some evidence for a very long time scale downward trend in rainfall. The data also suggest the existence of a large natural inter-annual rainfall variability in Africa, with the human intervention element playing a major role in the use and removal of surface vegetation cover in large magnitudes. This human activity is hypothesized to have increased the rainfall variability although this has not been proved. The variability of rainfall, is more pronounced at the small regions and watershed levels than at national level.

3.3.2. Temporal and Spatial Variation of Evapotranspiration

3.3.2.1. General Variations

This subsection examines the variations of evapotranspiration in a watershed, as a natural phenomenon and as a human-modified system in the form of land and water resource use, large scale vegetation cover changes, and urbanisation. The parameters are reviewed from the global to the watershed level.

Although precipitation falling on a watershed may be held and delayed in various locations within the system, it ultimately leaves the watershed as vapour or liquid water, the latter as surface or groundwater flow. With these exceptions, all water losses are therefore evaporative in nature. Through the processes of interception, direct evaporation and transpirations, the evapotranspiration process is accomplished.

Raudkivi (1979) asserts that approximately 70% of the annual precipiation on the land surface is returned to the atmosphere by evapotranspiration. Therefore evaporation and transpiration are very important elements of the hydrologic cycle, playing an important role in water balance analysis and in the design of hydroprojects. The rate of evaporation however, depends upon several hydrometeorologic and scale factors. The local rate for example, depends upon the availability of moisture, energy, and on the charateristics of the watershed. Solar radiation energy varies with latitude, season, weather, exposure of the surface and its albedo. In addition, moisture is transported away from the evaporating surface by wind and its turbulence.

In calm conditions, the air above the ground become saturated with water vapour and the rate of evaporation decreases rapidly. Wind then carries off the moist air and replaces it with warm air by convection or removes the moist air by turbulent dispersion and thus allows unhindered evaporation. In general, the higher the wind speed, the more effective the rate of moisture removal. Evaporation therefore varies from region to region. For example, evaporation from an open water surface in arid areas is about 4500 mm/yr and roughly 450 mm/yr in temperate climates. In all areas, solar radiation is the dominant source of energy and sets the broad limits of evaporation. The general physical principles of evaporation are well understood but still, there are some complexities which can cause errors in evaluating (Webb, 1975) and estimating it.

3.3.2.2. Evaporation

Evaporation is the transfer of moisture from an open or free water surface, a bare soil or water intercepted on a vegetal cover into the atmosphere. It is measured using atmometers and other containers of various shapes and dimensions. It can also be estimated from the water balance of a body of water or watershed area in question, provided sufficient accurate data covering the terms in the balance equation other than evaporation are available (Shahin, 1984). In the absence of actual data, empirical equations are used to estimate the evaporation.

The methods used to estimate the amount of evaporation range from the direct measurement using evaporation pans, to the use of empirical formulae, water balance, mass transfer and to energy budget methods. Most of the measurements are based on the assumption that evaporation is a function of the atmospheric conditions.

3.3.2.3. Transpiration

Transpiration is the process by which water vapour escapes from plant leaves, and enters the atmosphere (Lee, 1942). The transpiration process depends upon a supply of energy. Radiant and sensible heat energy supplied to plant leaves and stems causes evaporation within the plant leaf and when the resulting water vapour has diffused through the stoma and carried by turbulent mixing, the resulting water loss produces a water deficit within the leaf cells. This water deficit represents a suction force or potential which is transmitted from cell to cell through the plant system and is capable of drawing up moisture against the force of gravity (Hewlett and Nutter, 1969). Detailed discussion of transpiration is provided in Briggs (1967), Kozlowksi (1964), Slatyer (1967), Kramer (1969), McGowan et al., (1984), and Monteith (1985). Human activities therefore that alter the plant life, its characteristics and general extent of vegetation cover ultimately change the transpirational rate in the watershed.

3.3.2.4.Evapotranspiration

Evapotranspiration (ET) is the sum of water used per unit area by the vegetative growth in transpiration and that evaporated from the soil, snow or intercepted precipitation on a given area in a specified time. It is expressed in units of depth per unit of time. This implies that the factors affecting ET are essentially those affecting evaporation and transpiration. Therefore the rate of evapotranspiration depends upon the climate, plant development, density of vegetative cover, soil moisture supply, quality of water and length of growing season in the case of irrigated parts of the watershed. The climatic factors include; solar radiation, temperature, sunshine hours, humidity and wind speed.

When evapotranspiration takes place from a watershed completely covered by actively growing vegetation with unlimited supply of soil moisture, it is referred to as potential evapotranspiration (ETp). ETp can therefore be regarded as the upper limit of the actual evapotranspiration (ETa= ETm). The most common methods used for determining the ETp are the water-balance, the tank and lysimeter experiments, field soil moisture depletion studies, correlation with evaporation from a pan or from an open water body and estimation methods based on the physics of water vapour transfer and/or the heat energy-balance. Empirical methods are used where climatic data is available and direct measurements are impossible.

3.3.2.5. Estimation of Areal Evapotranspiration

The rate of evapotranspiration in a particular hydrologic system varies widely on a daily, monthly and on seasonal basis. The computation of the areal and basin wide ETa, on a rigorous basis requires the observations and measurements of all the required hydrometeorological variables which is difficult. In view of these associated difficulties, attempts have been made to estimate ETa indirectly from meteorological data using theoretical and empirical formulae (Rosenberg et. al., 1983).

Although ETp varies at each watershed point and continuously throughout the day, a spatially averaged ETp has been used successfully in small watersheds. The hydrometeorological variables at small watershed level is assumed not to change significantly and hence estimation of ETp is close to the actual value. Penman equation (1948, 1949, 1956) as modified by Businger (1956), Tanner and Pelton, (1960), Bavel, (1966) has been used successfully with climatic data to estimate the areal ETp of small watersheds. The latest modification of these equations is that recommended in FAO (1990) which is used predominantly in the estimation of reference evapotranspiration (ETo) in irrigated areas:

It is defined as the rate of evapotranspiration from a hypothetic crop with an assumed crop height of 120 mm, a fixed canopy resistance of $70sm^{-1}$ and an albedo of 0.23 closely resembling the evapotranspiration from an extensive surface of green grass of uniform height, actively growing, completely shading the ground and not short of water.

The estimation of the reference crop evapotranspiration is therefore determined using the combination formula based on Penman-Monteith approach (FAO, 1990). When combining the derivations for the aerodynamic and radiation terms, the combination formula becomes:

$$ET_o = \frac{0.408\Delta(R_n - G) + \gamma(900(T+273) * U_2(e_a - e_d)}{\Delta + \gamma(1+0.34U_2)} \qquad (3.1)$$

where:

ET_O is the reference crop evapotranspiration (mm/day)

R_n is the net radiation at crop surface (MJ m^{-2}/day)

G is the soil heat flux (MJm_{-2}/day)

T is the average temperature (deg.°C)

U_2 is the windspeed measured at 2m height (ms^{-1})

(e_a-e_d) is the vapour pressure deficit (kpa) in eqn 3.2

Δ is the slope of the vapour pressure curve (kpa, $deg.^{-1}$) in eqn: 3.3

g is the psychrometric constant (kpa $°C^{-1}$) in eqn: 3.4

$900 = (kJ^{-1}kgK)$ (3.2)

Vapour pressure deficit $VPD = e_a - e_d = \frac{e_a(T_{max}) + e_a(T_{min})}{2} - e_d$ (3.3)

where $e_a(T_{max})$ = saturation vapour pressure at $T_{max} = 0.611 \exp \frac{(17.27 * T)}{(T + 273.3)}$ (3.4)

and $e_d = \frac{RH_{mean}}{\frac{50}{e_a(T min)} + \frac{50}{e_a(T_{max})}}$ (3.5)

Vapour pressure curve: $\Delta = \frac{4098 * e_a}{(T + 273.3)^2}$ (Tetens.,1930; Murray 1967) (3.6)

where:

Δ is the slope vapour pressure curve (kpa $°C^{-1}$)

T is the air temperature (°C)

e_a is the saturation vapour pressure at temperature T (kpa)

e_d is the vapour pressure at dew point temperature (kpa)

and psychrometric constant (g), $\gamma = \frac{cpP}{\varpi} = \frac{0.00163P}{L}$ (3.7)

where,

c_p is the specific heat of moist air = 1.013 (KJ kg^{-1} $°C^{-1}$),

P is the atmospheric pressure at site (kpa)

w is the ratio molecul.weight water vapour/dry air = 0.622

L is the latent heat ($MJkg^{-1}$)

At watershed level however, the hydrologic system is composed of changing vegetation cover, urban structures, open grounds, forestry and bodies of water. Most of the methods used to estimate ETo, have limitations in handling these varied characteristics. For example, the Penman-Monteith method requires the knowledge of vegetation behaviour under different meteorologic conditions and times to convert ETo to ETm. Solar radiation and humidity data are also needed which are often unavailable. On the other hand, the energy balance and eddy correlation methods require sophisticated and expensive instrumentation. With the exception of the water balance approach (which ignores the dynamic characteristics of the regime), the above methods are difficult to adapt to small watershed studies. The water balance method is only accurate when multiyear periods are considered (Yin and Brook, 1992). To circumvent these difficulties, modelling approach is used in watershed hydrologic studies.

The conceptual modeling process starts with several assumptions. Actual evapotranspiration is estimated as ETo multiplied by a coefficient which is determined by antecedent soil moisture conditions (Saxton, and McGuiness, 1982). When there is sufficient moisture for vegetation water use, ETa equals ETm. ETo is estimated from

available climatological data. Since ETm determines the maximum possible value of ETa, it is a very important parameter in watershed hydrology. However, it is difficult to verify ETm estimates owing to the hypothetical nature of ETo (Thornthwaite, 1948)
In tropical Africa, Penman equation has been used successfully to estimate ETm. Dupriez (1959) measured and estimated PET in Congo, Rwanda-Burundi, Uganda, Tanzania and Kenya using lysimeters and Penman methods and found the daily potential PET in Kisozi and Musas in Rwanda-Burundi to be nearly 95% of the Penman free water surface evaporation. Hanna (1971) while studying the effects of water availability on tea in Uganda found PET from a full cover of tea to be 85% of that estimated with Penman's equation. These two field results enables the application of Penman equation in estimating PET in the sub-humid and humid areas in the tropics and specifically in East Africa (Shahin, 1985).

Dagg (1972) related the evapotranspiration rates to age of plants. The water use rose from 0.4 PET in young plants to 1.0PET when full cover is achieved. The evergreen forests had a ratio of ETa/PET= 0.9, average for the whole year. At Muguga in Kenya, the seasonal PET was 560mm, whereas the open water evaporation for the same season was 840mm, a ratio of, ETa/PET= 0.67. Evapotranspiration from grass grown lysimeter at the same location was observed by Glover and Forsgate (1964) for a period of 126 days in 1962. They found ETa= PET-1.17 for 5-day periods (R^2=0.94) and ET_a= 1.03PET for 10-day periods (R^2= 0.97), where ET_a is the actual evapotranspiration from the lysimeter (mm/day) and PET is the Penman evaporation using (n/N) for the relative duration of bright sunchine. The results from these experiments supports Dupriez (1959) view that Penman's formula is the best empirical method for estimating potential evapotranspiration in the tropics.

Results from the Penman's formulae in Nigeria, west Africa, gave reliable estimates on the annual, seasonal, monthly and daily basis and provided more reliable estimates of PET during the dry season and in the less humid areas. There is no doubt therefore that the Penman's formula also offers the best approach in estimating PET in West Africa (Ayoade, 1976).

In the absence of the basic climatic variables used in PET estimation, Christiansen (1968) and Christiansen and Hargreaves (1969) equation for estimation of PET from Class A pan evaporation is used. The equations for PET and coefficients take the form:

$$PET = 0.755_{pan} * C_{T2} * C_{W2} * C_{H2} * C_{S2} \quad (3.8)$$

where E_{pan} is measured Class A pan evaporation. The coefficients are dimensionless:

$$C_{T2} = 0.862 + 0.179(T_c/T_{co}) - 0.041(T_c/T_{co})^2 \qquad (3.9)$$

where T_c is the mean temperature in °C and T_{co}= 20°C

$$C_{W2} = 1.189 - 0.240(\frac{W}{W_o}) + 0.051(\frac{W}{W_o})^2 \qquad (3.10)$$

where W is the mean wind velocity 2 m above ground level in kph and W_o =6.7 kph.

$$C_{H2} = 0.499 + 0.620(\frac{H_m}{H_{mo}}) - 0.119(\frac{H_m}{H_{mo}})^2$$

where H_m is the mean relative humidity (RH) expressed decimally, and H_{mo} = 0.60

$$C_{S2} = 0.904 + 0.0080(\frac{S}{S_o}) + 0.088(\frac{S}{S_o})^2 \qquad (3.11)$$

where S is the percentage of possible sunshine hours, expressed decimally, and S_o=0.80. The original Christiansen reference referred to PET, but this may be regarded as grass reference ET, because the data from grass surfaces were used as calibration input to the method (Christiansen and Hargreaves, 1969).

3.3.2.6. Variation of Evapotranspiration in Tropical Africa.

The distribution patterns of PET over a small watershed area is relatively similar because the control factor is the availability of solar energy and moisture. For example, Ayoade (1988) estimated the spatial variations in rates of evapotranspiration in Africa (Figure 3.12) using the Thornthwaite's formula. The values varied from less than 850 mm in most parts of South Africa to over 1750 mm in the interiors of West Africa. Within the tropics the greatest seasonal variations occur in dry continental interiors while the lowest variations occur in humid coastal locations.

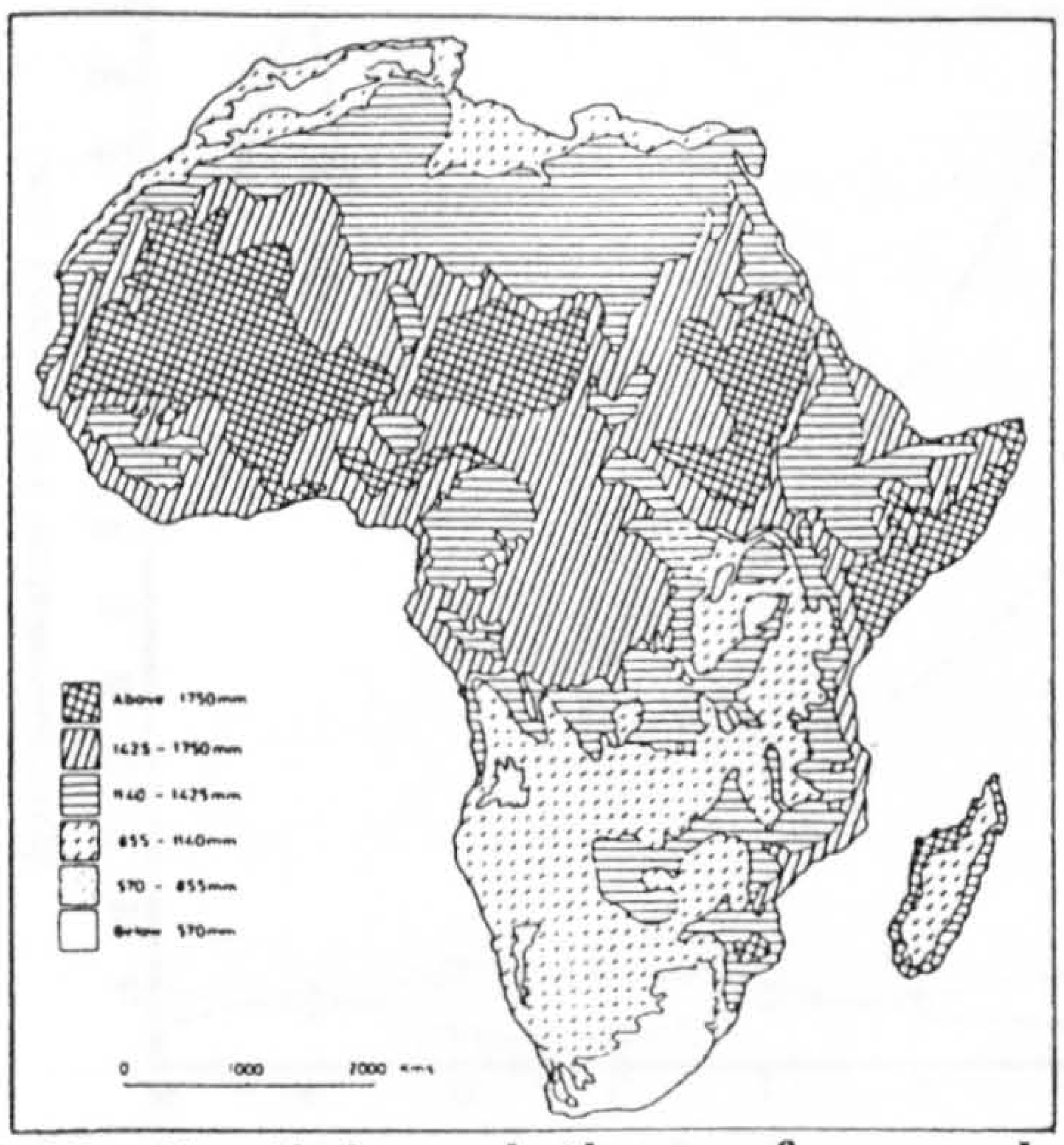


Figure 3.12. Spatial variations of annual potential evapotranspiration over Africa according to Thornthwaite's formula (after Ayoade, 1988).

Seasonal variations in PET are higher during dry periods when insolation is higher and relative humidity is lower, than during the rainy season when insolation is lower as a result of greater cloud cover and higher humidity values. The estimated values of PET tend to vary with the rainfall amounts in the tropics and with temperature in the extra-tropical areas. Thus in the tropics values of ETm are higher during the rainy season than during the dry season because of abundant solar energy. The seasonal variation of PET reflects variations in altitude of the sun, cloud cover, windspeed and humidity. Hence, high PET is meaningful only when related to available water. Therefore, seasonal variation of the amounts of PET in Africa is indirectly dependent upon the rainfall patterns.

Estimates of PET in forested watersheds in the humid tropics are highly variable and varies with the method used. Sengele (1981) estimated PET from a watershed in the Loweo Region of Zaire using Class A Pan, lysimeter and 15 non-recording and two recording rain gauges distributed uniformly over the watershed. The weighted average rainfall was computed using Thiessen's method. Water yield from the watershed was monitored by a Parshall flume, water level recorder and a current meter was used for instantaneous flow measurements. Modified Penman's formula was used to calculate the evaporation from a free water surface.

Penman estimated PET was 79% of precipitation, Pan evaporation over estimated the PET during humid periods and was equal or slightly lower than PET during the dry months. Water balance approach gave only approximation of PET because of the erratic distribution of rainfall and the desiccating winds in dry periods (Figure 3.13). These field findings in central Africa, further demonstrates the applicability of Penman's equation in the humid tropics.

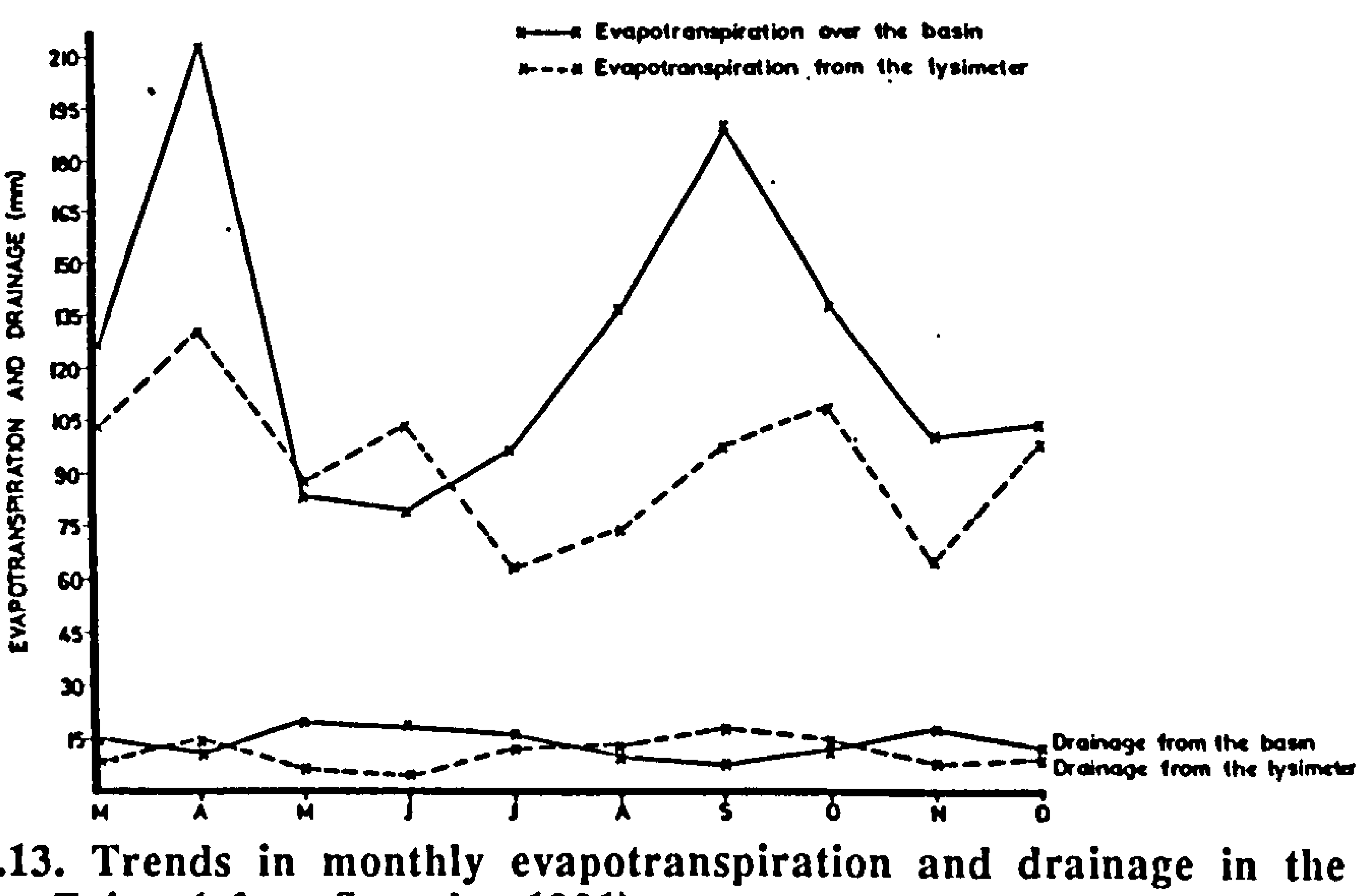


Figure 3.13. Trends in monthly evapotranspiration and drainage in the Loweo Region in Zaire (after Sengele, 1981).

3.3.2.7. Summary on Evapotranspiration

The review on evapotranspiration showed that a quantitative understanding of PET is important because of various reasons: First, over a long-term the difference between precipitation and PET is the amount of water available for human use and management. Hence a quantitative assessment of the water balance and the effects of climate and land use on these resources require understanding of the PET. Secondly most of the water lost through ET is used for vegetal growth that form the base of the earth's ecosystem, thus an understanding of the interactions between the ecosystem type and PET is essential. Thirdly, the portion of water from a given rainfall event contributing to the groundwater storage and streamflow depends upon the *insitu* moisture status of the watershed which can be established by determining the amount of ET that occurred since the previous rainfall events.

The review also demonstrated the high cost and difficulty of measuring PET directly. Hydrologists therefore have developed several approaches for estimating PET based on the physics of evaporation and the basic principles of the conservation of mass and energy.

The Penman-Monteith approach was shown to estimate the PET in both the tropics and temperate regions with a relatively good accuracy. While this method is applied mostly in irrigated areas, it provides good estimates of watershed evapotranspiration. There is however, still some work to be done in estimating actual evapotranspiration particularly at the arid and semi-arid tropical climates.

3.3.3. Infiltration and Groundwater Storage

3.3.3.1. Infiltration

Infiltration is the movement of water into the soil through the soil surface (Horton, 1933). Infiltration rate is the amount of water percolating into the ground in a given time. When rain falls after a dry period, the infiltration into the soil occurs at a relatively high rate but as the rainfall continues, the infiltration rate decreases with time and ultimately reaches a constant value which is approximately equal to the soil saturated hydraulic conductivity.

At the same time, water is drawn into the soil by forces and energies available in the soil environment. This environment, is comprised of soil particles, to which water molecules are attracted (Miller, 1977). The sizes of pores determine the relations of the soil surface areas to the mass of water. Infiltration is thus a hydrological process comparable to precipitation, evapotranspiration and streamflows. Not only are infiltration rates affected by soil properties, but also by the soil depth, and duration and intensity of rainfall as well as the original ground water table.

3.3.3.1. Factors Influencing Infiltration

When rain falls at intensities exceeding the soil infiltration rates, the following occurs: (1) A thin film of water forms on the surface and downflow starts; (2) the runoff accumulates in depressions; (3) when the depressions are full, it overflows; (4) this overflow enters microchannels, forms rills, and discharges into gullies and major channels and finally (5) lateral inflow from the land surface takes place. Before surface runoff starts, water is infiltrating into the soil and continues during the runoff period.

The rate of initial infiltration therefore depends upon the type and antecedent moisture state of the soil. It is higher in sandy soils than clays (Table.3.4). The ground cover also affects infiltration rates, the highest rates is obtained in forest cover and the lowest rates for bare, compacted soils. The infiltration rates for permanent pasture and for cropped land lie between these two extremes but the actual values depends on the nature of the pasture and on the cropping patterns (FAO, 1973). This implies that the better the land use, in terms of good cropping and vegetation cover, the higher the rate of infiltration, and hence more water percolation into the groundwater storage.

In a forested watershed precipitation is intercepted by the vegetation and held in storage on its foliage for some time. In the urbanized parts of the watershed precipitation falls on roofs, roadways, and other impervious surfaces. Infiltration in these urban areas is low and some times none. Surface storage builds up rapidly on the impervious surfaces and then flows into gutters before being collected either in open drains or in underground drainage systems. If these facilities are unavailable, surface runoff in the form of floods is experienced as shown in Figure 3.14. Effects of human activities on infiltration rates are clearly demonstrated in urban areas. Permeable natural surfaces are replaced by man-made surfaces, which reduce infiltration rates and increase overlandflow and hence flooding is common especially in cities with uncontrolled urban development (Ayoade, 1988).

Table 3.4. Typical infiltration rates on bare soils one hour after commencement of the rain (after Viessman et. al, 1988).

Soil Group	Infiltration Rates, mm/hr)
High (sandy, open-structured)	12.7- 25.4
Intermediate (loam)	2.5- 12.7
Low (clay, close-structured)	0.3- 2.5

Infiltration rates in different land uses, from some stations in Nigeria (Table 3.5), depict differences in infiltration rates over natural surfaces compared to the human-modified surfaces, confirming the observation that infiltration rates tend to decrease with intensity of human interventions. The data shows that infiltration rates decreased from 1000mm/hr in fully forested areas to barely 50mm/hr in bare grounds and farmed lands had average values of over 500 mm/hr.

This discussion has demonstrated that the capacity of soil to take in water varies with the activity of the vegetation, both in the root zone and above the soil surface. This directly affect water movement and indirectly conditions the soil media to pull water. Forest floor for example has a greater hydrologic benefit because it develops soils with high organic content, good aggregation, and a large population of microscopic and large organisms that keep soil structure open and its infiltration capacities high. The data in Table 3.6 show high rates of infiltration capacities at a forest floor, even after losing some surface layers of litter and removal of humus. This phenomenon suggests why many forested watersheds monitored for several years, have never been seen to produce surface runoff (Miller,1977), implying that rainfall rates do not exceed the high rates of infiltration capacities in the upper soil layers. For it to increase surface runoff and hence reduce minimum baseflows, human-accelerated deforestation, large-scale removal of vegetation cover, and intensive cultivation should take place.

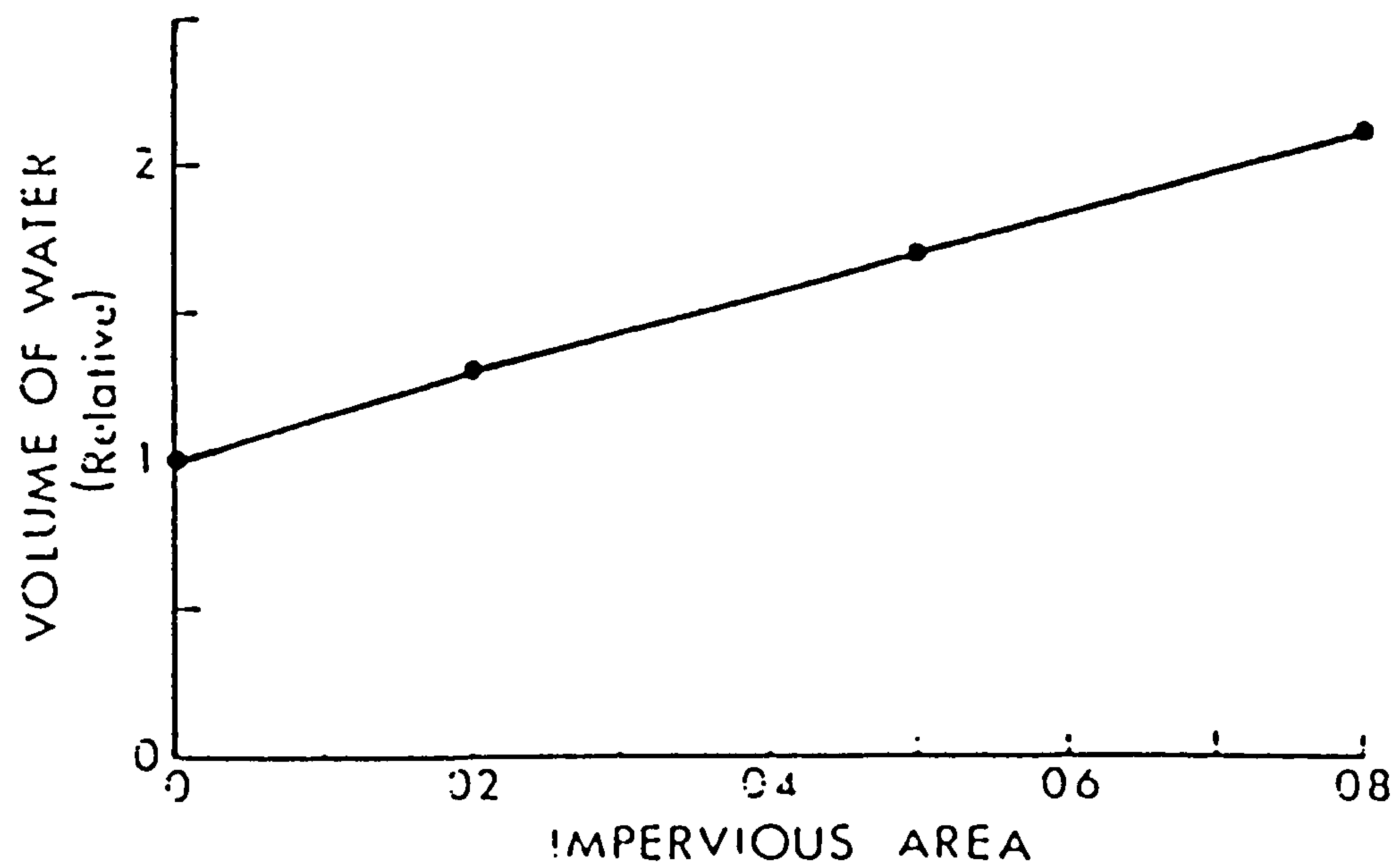


Figure 3.14. Infiltration characteristics in urbanized catchments.

Table 3.5. Mean infiltration capacities of different land use surfaces in Ibadan, Nigeria, during the 1973 dry season (after Akintola, 1974).

Type of Surface	Mean infiltration capacity (mm/hr)
Bare ground (crusted)	50
Bare ground (within family compound)	270
Highly disturbed grass	170
Disturbed grass	270
Undisturbed grass	490
Tall grass	620
Fallow land	610
Farm land	630
Forest	1000

Table 3.6. Infiltration capacities under different forest cover (Lull, 1964).

Ecosystem	Infiltration capacity (mm/hr)
Undisturbed forest floor	60
Forest without litter and humus layer	49
Forest floor burned annually	40
Pasture, unimproved	24
Succession	**Vegetation**
Old pasture	43
Pine forest, 30 years old	75
Pine forest, 60 years old	63
Oak -hickory forest	76

On the other hand, good agricultural practices improve infiltration capacities of soils. Infiltration behaviour and rates in Piedmont soils under different land uses presented in Figure 3.15 shows curves of accumulated water that infiltrated into the soils under different types of vegetation and agricultural management. The declining rate of infiltration is visible in the curvature of the lines with very good land husbandry.

Consequently, a soil structure is an important property which influences soil erodibility. On the contrary, grazing is often a human induced activity that damages soil structure through tramping and compaction. Heavily grazed lands tend to have lower infiltration capacities than ungrazed lands as depicted for example in Table 3.7. The removal of vegetation cover and associated litter, changes infiltration capacities, because the cover protects the soils from packing by raindrops and provides organic matter which binds soil particles. The infiltration rates were lowest at all the heavily grazed sites.

3.3.3.3. Temporal Variation of Infiltration Rates

Just as the characteristics of a soil-vegetation ecosystem vary throughout the year, so does infiltration. This can be shown by examining the mean rates of infiltration during individual rainstorms determined by comparison of rainfall and runoff intensities. A 29-year storm frequency distribution records in Iowa drainage basin in Table 3.8, each integrating the infiltration of water over a watershed area of 8 sq.km, and equally covered with annual and perennial vegetation. Infiltration values range from 2mm/hr in some storms to 45 mm/hr. Such a wide range shows the effects of variations in dryness of the soil at the particular seasonal time when a storm began, of porosity, and of other changeable soil conditions.

Table 3.7. Rates of infiltration on grazed and ungrazed lands in America (after Gofford and Hawkins, 1978) as quoted by Guidie, (1988)

Experimental site	Rate of infiltration in mm/hr Ungrazed	Heavily grazed
Montana	2.5-66.0	5.1-15.2
Oklahoma	134.6- 309.9	40.6- 83.8
Colorado	40.6-83.8	20.3- 30.5
Montana	109.2- 185.4	20.3- 96.5
Wyoming	30.5- 38.1	17.8- 30.5
Lousiana	45.7	17.8
Kansas	33.0	20.3
Arizona	40.6	30.5

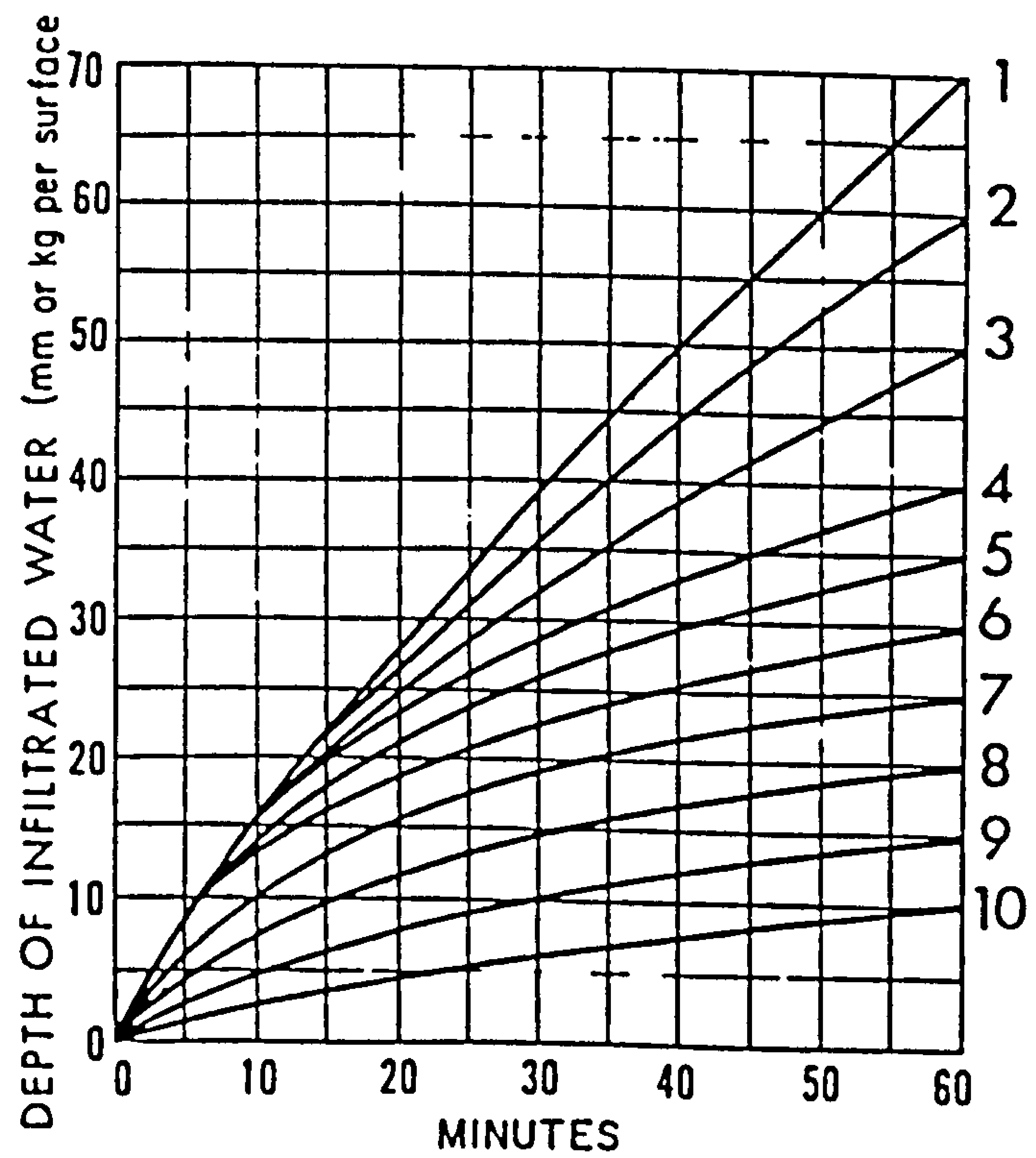


Figure 3.15. Curves of accumulated water infiltrated into Piedmont soils under different types of vegetation: 1-old pasture; 2- permanent pature 4-8 yr old; 3-lightly grazed permanent pasture 3-4 yr old; 4-moderately grazed permanent pasture; 5-hay; 6-heavily grazed permanent pasture; 7-strip-cropped and; 8-grain; 9-clean-tilled land; 10-crusted soils (after Holtan and Kirkpatrick, 1950).

Table 3.8. Cumulative frequency distribution of storm mean infiltration rates in the Ralston Creek basin, Iowa (after Johnson and Howe, 1956).

	Infiltration rate exceeded in a given percentage of storms (mm/h)	
Percentage of storms (%)	All storms	Spring storms
10	45	30
20	30	20
50	20	12
85	10	6
100	3	2

Temporal variations in infiltration rates and to an extent soil moisture storage are due to the varying amount of rainfall with time input and evapotranspiration. Spatial variations in soil moisture can also be ascribed to variations in climatic conditions of precipitation and evapotranspiration.

In addition, long-term changes in soil properties reduce its infiltration capacities. Assuming no secular change in precipitation, and taking into account long-term changes in the watershed, a slow recovery of infiltration was observed in experiments reported in USSR by Grin (1965) in Table 3.9. A complete recovery and restoration to pre-settlement conditions however, seem unlikely from these data. These results indicate no màjor change probably because of the assumption of no change in rainfall regime.

However, a different observation is seen in data analysed by Chemelil (1986) for changes in temporal land use in the Brazos Valley, Texas. The analysis of the data revealed decreased infiltration rates attributable to long-term continuous irrigation and compaction of the soil, which increased bulky density and reduced soil organic matter (Table 3.10). Low infiltration rates are more pronounced in fields with more than 40 years of continous irrigation and use of large machinery.

Table 3.9. Long-term infiltration rates in central Russia (after Grin, 1965) as quoted by Miller (1977).

Time Period	Infiltration capacity (mm/yr)
Before 10th century	460
End of 19th century	410
1925-1950	415
Early 1960s	425
Near future (1980s)	440
More distant future	450

Short term variation of infiltration capacities can be attributed to rainfall characteristics and soil antecedent moisture conditions in the watershed. For example infiltration rates may decrease with time for a given rainstorm for two reasons: First, the wetting of the soil in the beginning of the rain causes granules of clay or silt to expand thus closing some of the pores in the soil. Secondly, the film of water surrounding each soil grain forms a net of interconnected veins of water offering frictional resistance to the infiltrating water. This slows down the rate of downward movement of water, hence slowing the rate of entrance of new water from the surface (Leopold, 1974).

Table 3.10. Long-term infiltration capacities in cotton fields in the Brazos Valley, Texas (after Chemelil, 1986).

Field Number	Historical Land use Years under irrigation	Soil Type	Organic Matter(%)	Bulky Density(g/cm-3)	Infiltration Rates (mm/hr
11	more than 50 yrs	loam	0.07	1.5	0.067
20	more than 50yrs	loam	1.6	10.8	3.07
12	more than 44 yrs	clay/loam	0.7	1.6	0.087
17	more than 40 yrs	loam	1.64	1.4	3.07
3	more than 8 yrs	loam	1.75	1.3 5.	15
14	more than 5 yrs	loam	2.37	1.4	15.23

Similarly Ayoade (1988) reported relatively low infiltration rates in tropical Africa. This was attributed to high rainfall intensities which caused excessive surface runoff and soil erosion, thus reducing the amount available for plant transpiration. Secondly, high rainfall intensities encouraged formation of surface sealing and crusting of the soil surface which with time reduced infiltration rates.

3.3.3.4. Measurement and Estimation of Infiltration

Infiltration rates are measured using infiltrometers and estimated with hydrogaph analysis or computed using empirical equations. Two types of infiltrometers in use are the flooding and the ring infiltrometers. Detailed description and operation of these infiltrometers is provided in Gregory and Walling (1973). Indirectly, infiltration over a watershed is estimated by analysing the hydrographs of streamflow from natural rainfall. This ensures that the instant variation of precipitation is taken into account unlike in the use of infiltrometers.The effects of the watershed characteristics are also incorporated.

Methods used in estimating infiltration by hydrograph analysis are based on one or more of the basic principles described by Musgrave and Holton, (1964) namely: (1) the detention-flow relationship method to derive rainfall excess; (2) time-condensation methods to eliminate periods of inadequate rainfall and (3) the block method of dividing a rain storm into a series of blocks to obtain average rates of infiltration throughout the rainstorm.

3.3.3.5. Infiltration Indices

Hydrograph analyses in the form of infiltration indices relates more directly to the prevailing conditions of precipitation and watersheds than to the use of the infiltrometers. Two infiltration indices are frequently used. The ø-index (Linsley et al. 1958) is based on a constant infiltration rate. That part of the rainfall that exceeds the infiltration rate is considered to contribute to the surface runoff (Figure 3. 16). The procedure does not include the decrease in infiltration rate during the rainfall event.

Firstly,Ø-Index is a constant rate of infiltration such that the total rainfall in excess of the index is exactly equal to the direct runoff from the storm and is represented as:

$$Ø = \frac{\text{Watershed recharge}}{\text{duration of rainfall (t)}} \qquad \phi = \frac{F}{t} \tag{3.12}$$

where, F= watershed recharge (P-Q) and t= duration of rainfall (t). To obtain the index, an average discharge hydrograph for the watershed is analysed to obtain the direct runoff. The index is then derived using a trial and error method.

Secondly, W_i-index is used, which attempts to separate the initial water abstractions made up of interception and depression storage from the watershed recharge presented in the form:

$$W = \frac{\text{recharge} - S}{\text{duration of infiltration}} = \frac{P-Q-S}{T_f} \tag{3.13}$$

where, P-Q is the watershed recharge (mm), P is the total storm rainfall (mm), T_f is the duration of infiltration (hr), Q stream discharge (runoff, mm), S change in surface storage, which is difficult to estimate (mm). The minimum W_{min}, often approaches the soil hydraulic conductivity

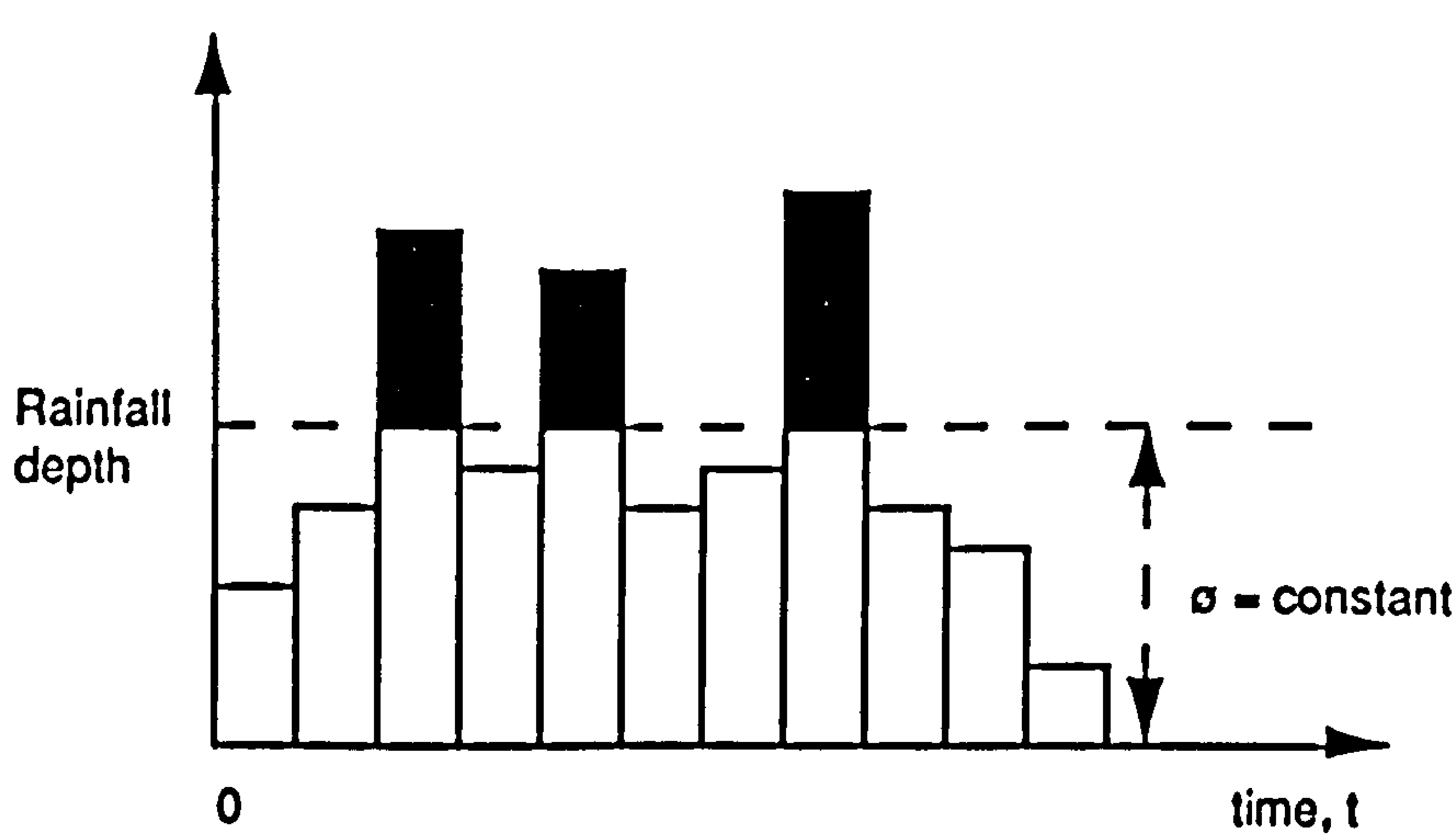


Figure 3.16. Determining effective rainfall hydrographs with the use of ø-index i.e for a constant infiltration rate (after Gutknecht, 1972).

3.3.3.6. Assessment of Infiltration Using Empirical Formulae

Several other approaches have been developed to quantify the elements of infiltration as a key step in the transformation of rainfall hydrographs into runoff hydrographs. Two basic empirical approaches are commonly used in the estimation of infiltration namely, the soil physics approach exemplified by the Green and Ampt model (1911) and the hydrological approach demonstrated by Horton (1933). The Green and Ampt model is the classical approach because of its reliability but the excessive amount of data required limits the extent of its use. Horton's hydrologic approach derived from the energy-work principle as a function of time is used in the following format:

$$f_p = f_c + (f_o - f_c)e^{-kt} \qquad (3.14)$$

where, f_p is the infiltration rate at capacity rates at time t_i (mm/hr),
f_o and f_c are the initial and final infiltration rates (mm/hr), k is a constant and a function of soil and vegetation cover. Integrating equation 3.14 gives the accumulated volume, F:

$$F = f_{ctp} + \frac{fo - fc(1 - e^{-kt_p})}{k_{fc}} \qquad (3.15)$$

Horton's infiltration rates are capacity rates and based on the assumption that the rainfall rate is always greater than infiltration capacity rates so that some ponding will always result. Little experimental data exist on the parameters f_o and k (Oyebande, 1993). This approach accounts for the diminishing infiltration rate presented in Figure (3.17). The fluctuations of rainfall intensity or antecedent soil moisture, for which runoff can commence even with small depth or low-intensity rainfall are often not considered (Tauer and Humborg, 1992).

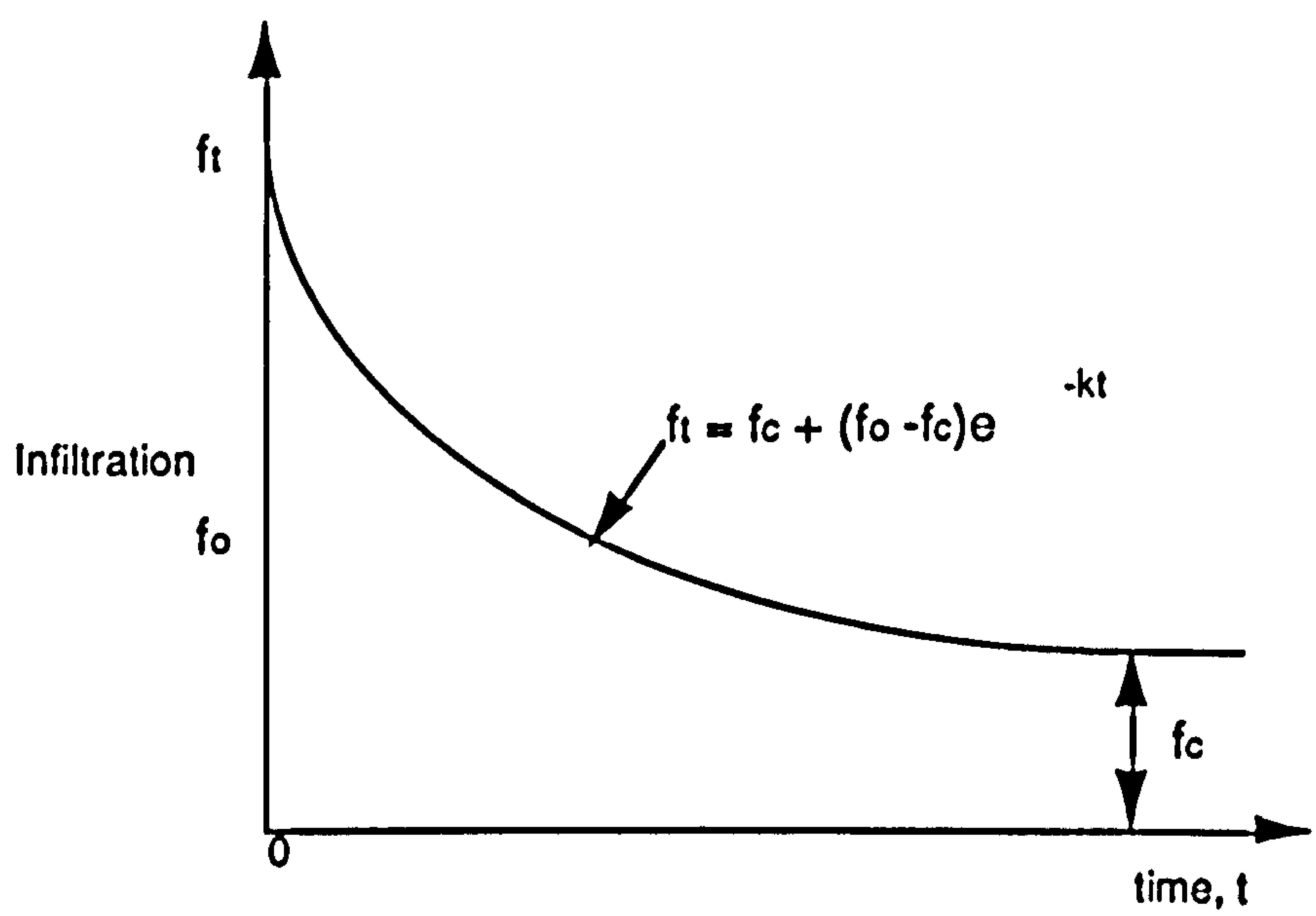


Figure 3.17. Effective rainfall hydrographs based on variable infiltration rates according to Horton(1933) and Shultz (1967) in (Gutkecht, 1972).

3.3.3.7. Groundwater Storage

Ground water is derived mainly from precipitation. It is water that has infiltrated and percolated downward into the earth directly from the atmosphere, from streams, or other natural bodies of water or from artificial water channels (recharge systems, leakage from reservoirs, irrigation).

It is that water which, during the process of the hydrologic cycle, passes beneath the ground surface and reaches a zone of saturation where it is stored for a period of time, and then passed on to the next stage. Groundwater storage applies to that stage of the hydrologic cycle during which water occurs as groundwater in the zone of saturation, including that part when water is entering and leaving storage (Figure 3.18).

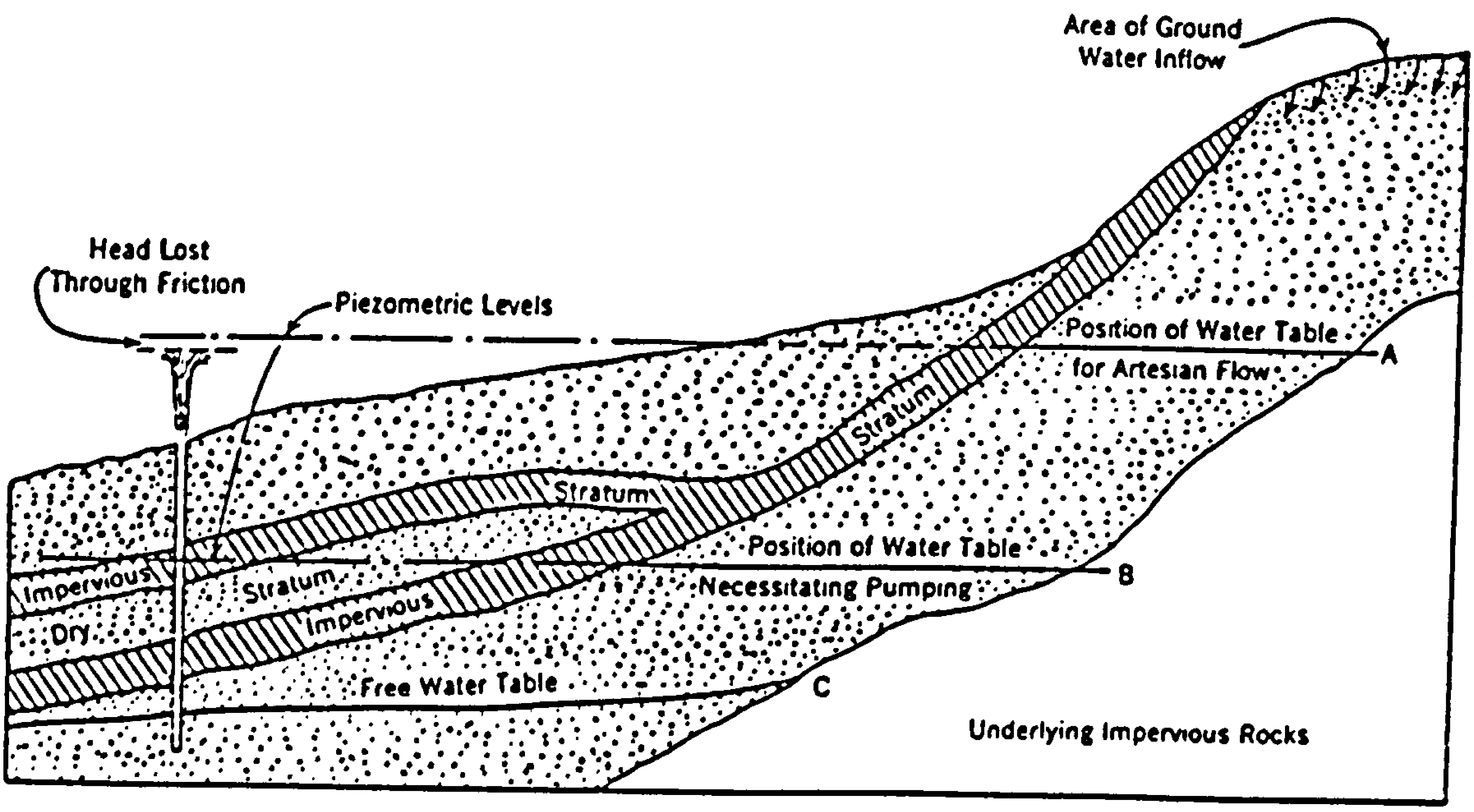


Figure 3.18. Occurrence of free perched and artesian ground water

The ground water storage serves as a regulator in the hydrologic cycle, holding briefly water that would otherwise have passed rapidly through succeeding phases of the cycle. It serves as a regulator of streamflows, maintaining it during dry periods. It acts as a source of water supply in localities where surface water is inadequate, and as a source from which water is abstracted in a regulated manner. Like other storages, these bodies can be analyzed in terms of inputs, outputs, and resulting changes in the volume of water stored.

The environment of groundwater varies spatially depending on the local geology. and on environmental factors.Ground water aquifers differ greatly in the amount of water they store because of different porosities that range from 0.2 to 0.4 in gravel, 0.1 in sandy and less than 0.1 in sandstone. The spatial differentiation in groundwater environments does not necessarily display the same pattern like at the ground surface, except where seepage and recharge areas in shallow groundwater are prominent or where such intrusive systems as landfills have been established (Miller, 1977). With increasing depth these spatial differences tend to diminish and more uniform aquifer conditions underlie the mosaic of surface watersheds.

Groundwater storage is recognized as a resource having value in itself, because it provides a secure, clean and potable water. Groundwater storages therefore can be managed just like surface reserviors, water being drawn in dry seasons and then replaced during the rainy seasons in order to sustain the reservoirs year-to-year without depletion. In most cases however, man removes the water from year-to-year without replacement in terms of an accelerated infiltration rates or artificial recharge. Artificial recharge is achieved naturally only through recharge from deep percolation of irrigation water, but sometimes this water allows contaminants to enter into the reservoirs and render them unsafe for human consumption. In effect, the continuous use of ground water without an equally continuous recharge ultimately will change the streamflow patterns.

3.3.3.8. Human-induced Changes in Groundwater Storage

Annual variation in groundwater storages, indicates the season when most percolations occur. Short-term and seasonal changes in pressure are not usually as marked as in a free water surface, because, a confined aquifer does not receive brief pulses of percolating water, rather a slow steady inflow, perhaps by leakage from other aquifers. Changes in the volume of water may be buffered by changes in the aquifer volume itself. It will recover spatial volume when the water volume increases.

Long-term changes in groundwater storage take place slowly. For example the amount of percolation from the higher layers may decrease with increasing urbanisation. Drilling and water abstraction of the groundwater stored in the past in this case inevitably reduces their volume and continuous abstractions results in cones of depressions around areas of concentrated demands. These are manifested by: (i) the decrease in water table, (ii) artesian wells cease to flow, a (iii) decrease of surface bodies of water which these aquifers used to supply, and (iv) sea water intrusion into the aquifer to occupy space vacated by the shrinking groundwater storage.

Extreme changes in groundwater tables due to human actions are widely reported. For example, the high groundwater table in the Salt River valley in Arizona (USA), because of over irrigation in the 14th century (Visher and Mink, 1964). When irrigation resumed in late 19th century, and accelerated in the 20th century, during the construction of Roosevelt Dam (Skibnitzke et al. 1961), swamps promptly developed again. Pumping was done to lower the water table. Since 1923 shrinkage in volume is indicated by a 45m drop in water table (Anderson, 1948). A proper management of water became a matter of primary interest to the general public but despite this knowledge, shrinkages continued unabated at an average of 3m/yr water table drop (USWRC, 1968).

A prediction of continued decline in the volume of this water body was carried out. Over the period 1964/74, Anderson, (1948) predicted an average decline of 16m in water table. Over the period, 1964/84, his model predicted a 32-m average drop, with a shift in its distribution. Mann, (1963) perhaps puts it in a better perspective thus:

" while some of the recent developers of groundwater were interested in a permanent livelihood on the desert, many were of the suitcase variety, willing to make an investment for short-term profits with full knowledge that the resource eventually would play out"

Another form of human management of groundwater storage involves pumping of water using wells. A well works by creating a cone of depression in the water table that will enable water to move by gravity. The flow into this borehole depends, on the features of the local geological environment. This pumpage interupts water that otherwise would flow into channels, thereby reducing off-site yield in rivers.

Over irrigation on the other hand raises groundwater tables. The history of the Salt River Valley event is reflected in the elevation of water table shown in Figure 3.19 (Harshbarger et al. 1966). It portrays the water table rise produced by percolation from surface-water irrigation beginning about the turn of the century, the slow decrease during the 1920s and 1930s, and the sudden fall after the installation of large pumps after 1945. Comparing with the curve of pumping rate indicates that leveling off of the rate of pumping about 1953 did not stop the shrinkage of the groundwater body, since withdrawals continued in very large amounts.

Groundwater reservoirs extend over greater areas than typical river basins or watersheds, and and as a commonly shared resource by several ecosystems. As water moves within the groundwater body, it traverses distances greater than the diameter of the watersheds from which it originally percolated; the characteristics of the watershed hydrologic regime are carried (inter-basin transfer) beneath the watersheds and vice-versa. This therefore makes it difficult to exactly quantify the groundwater potential of a watershed in terms of its quantity and quality, because, other human intervention such as well development outside the watershed indirectly affects its river flow regime.

Changes in vegetation cover in expansive areas also affect local groundwater storages. Increases in percolation resulting from lower evapotranspiration after woody, deep-rooted vegetation are converted to shallow-rooted, short-lived cereal crops, upsets the local water balances. After the removal of trees, for example, the soil profiles become wetter, and salt water moved down slopes, eventually forming a water-table at the foot of the slope within the capillary reach of the surface (Leeper,1970). Conversion of vegetation in the Victoria area in USA, from forestry to grazed perennial grass, resulted in decreased evapotranspiration brought by corresponding increase in percolation and movement of groundwater (Downess, 1961).

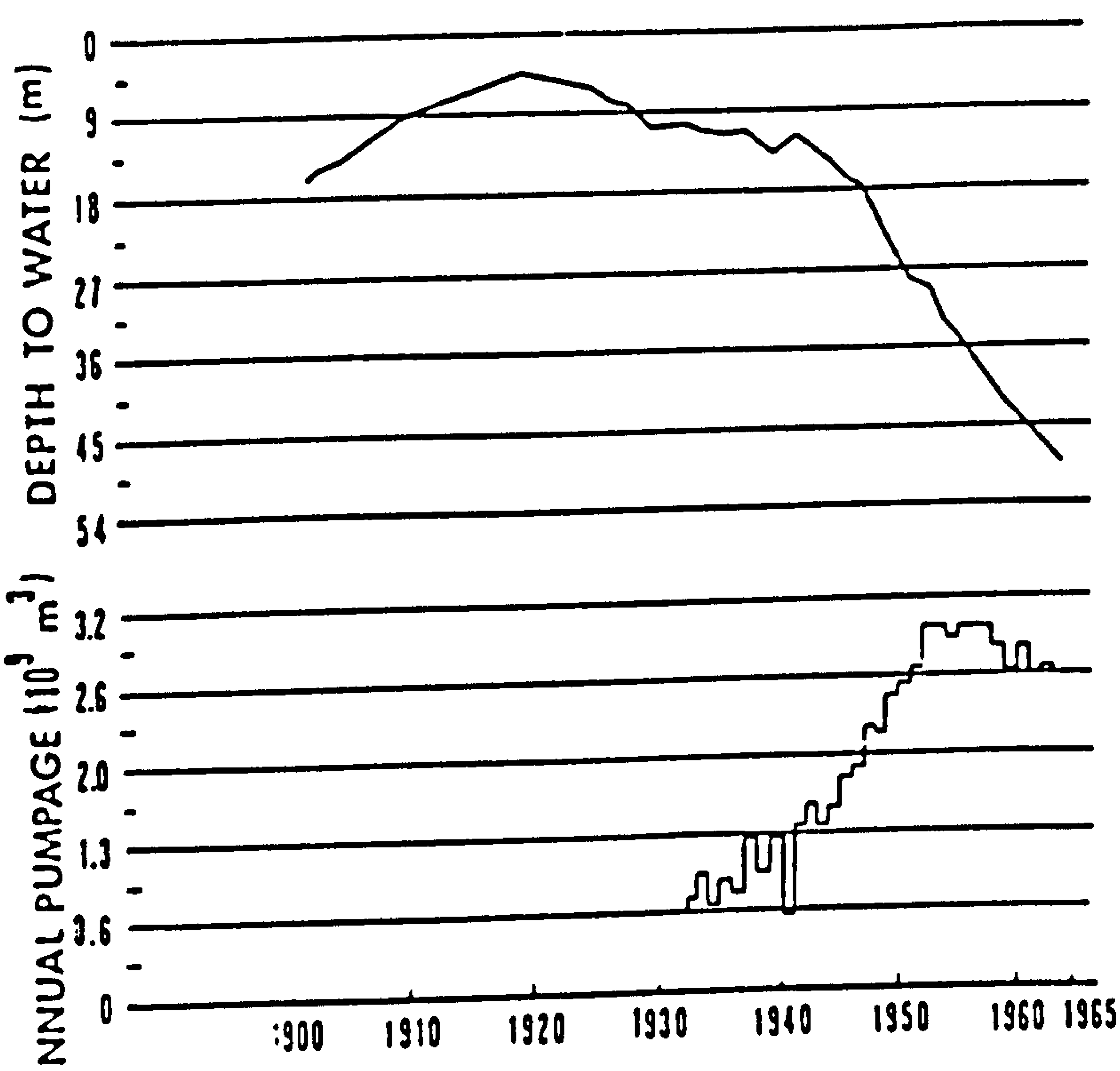


Figure 3.19. History of pumpage of groundwater from the Salt River Valley and changes in depth to the water table (after Harshbarger et al., 1966)

Increased water demands by increased population means increased groundwater pumpage (Drennan, 1979). This reduces water tables and replacement especially in coastal areas, of fresh water by salt water. Other environmental consequences of these two phenomena include ground subsidence and soil salination. Ground water abstraction in London for example shows a rapid increase in the number of wells tapping groundwater in the area since 1850 (Porter, 1978) as presented in Figure 3.20. Figure 3.20 also illustrates the widespread and substantial changes in ground-water conditions that resulted. The piezometric surface in the aquifer fell by more than 60 m over hundreds of square kilometres (Goudie, 1986) implying huge volumes of water in a period of 115 years.

Small-scale stream recharge from groundwater bodies sometimes occurs as a consequence of man's efforts to modify the soil environment through drainage works. The drainage systems used ensure that water tables remain below the ground surface. This alters the hydrogeophysical nature of the area in the long run. Warmerdam (1982) studied the effects of drainage improvement on hydrological regimes of a small representative watershed in the Netherlands.Since poor drainage conditions involve relatively shallow water table depths in wet seasons, the depth of the water table and its range of fluctuations are important parameters to characteristize the drainage conditions of the area. The effects of these human interventions were evaluated by comparing the frequency distributions of depths before and after the 1967 rehabilitation works. To establish whether any change in the depth had occurred as a result of the works, double mass curves of water-table were plotted. To indicate any trend of depths, cumulatives of these values for 22 wells in the Hupselse Beek and Leerinkeek watershed were plotted against the corresponding water levels in four bench mark wells. From this the average lowering of the water-table ranged from 10 to 35 cm per year in 14 wells which had significant breaks in the double mass curve slopes. For example Figure 3.21 presents the double mass curve for well 56 in relation to control well number 13.

Depths of water tables in forested and grassland sites in Sweden's Velen basin, were monitored by Anderson (1989). By comparing the groundwater levels at the clear cut sites with average levels in the basin, it was concluded that the depth of water table decreased after the clear cut (Figure 3.22). When parameter values, with good fit before deforestation (1968-1970) were used for the period after the deforestation (1972-1973), significant underestimations of the soil moisture content were observed. When changes in the relationship between moisture depletion and the actual to potential evapotranspiration ratio were considered, the fit was significantly improved.

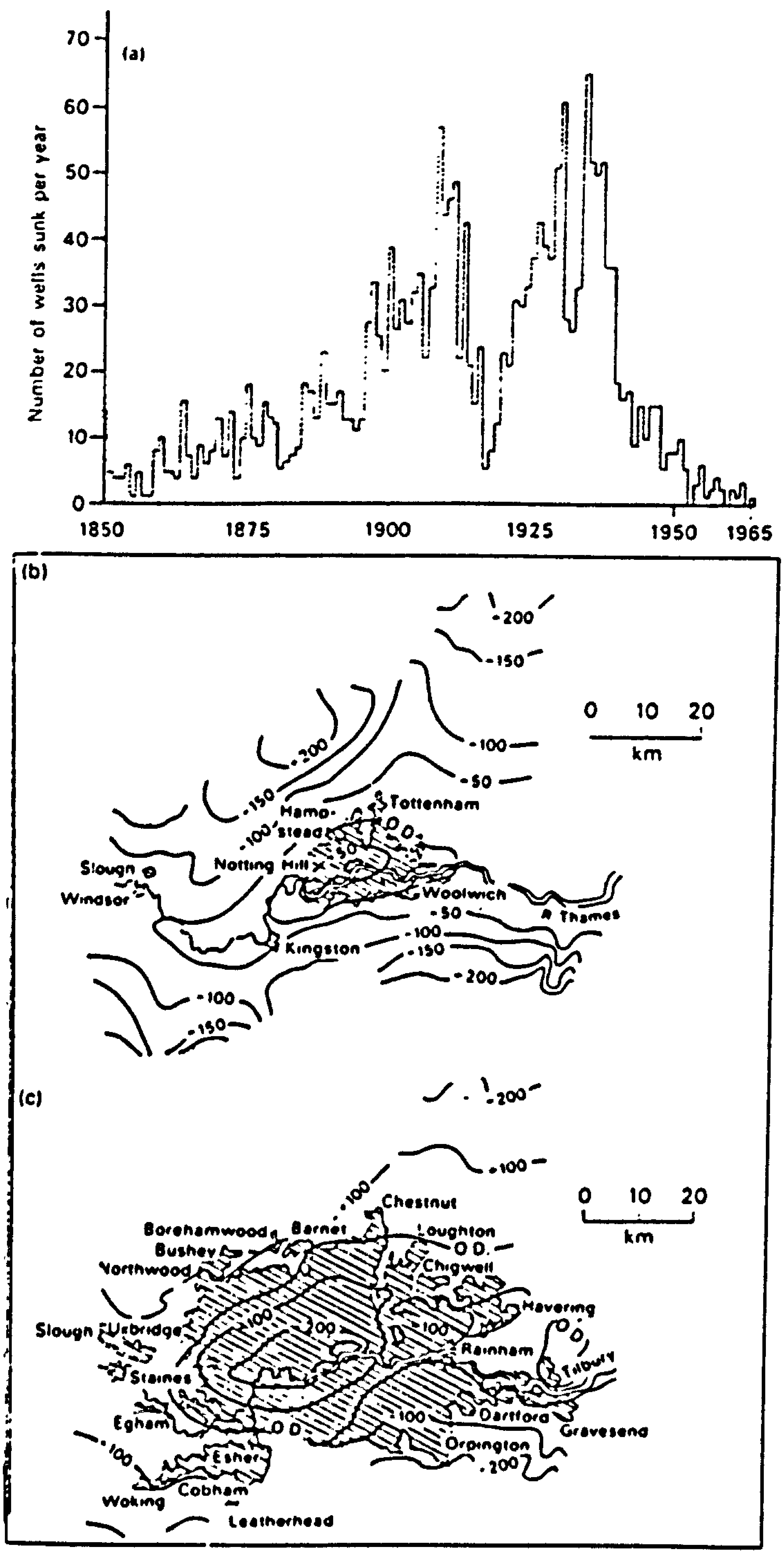


Figure 3.20. Changing groundwater conditions in London area: (a) construction of wells tapping the confined aquifer below London (1850-1965), (b) ground-water contours in 1875, (c) ground-water contours in 1965 (after Porter, 1978 and modified by Goudie, 1986)

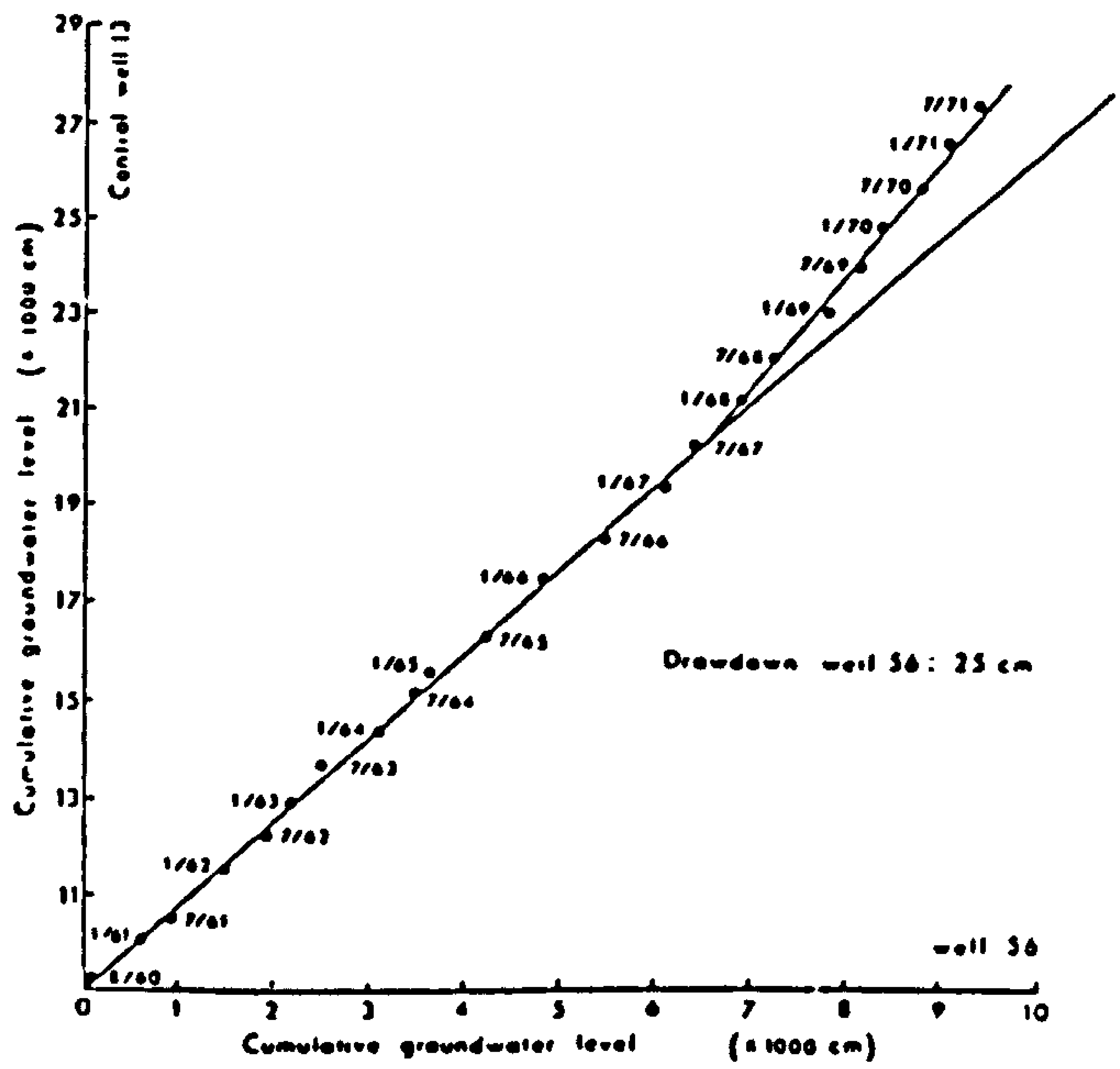


Figure 3.21. Double mass curve of bimonthly groundwater levels above the reference level for Well and control Well 13 in Hupselse Beek Leerinkee watersheds in The Netherlands

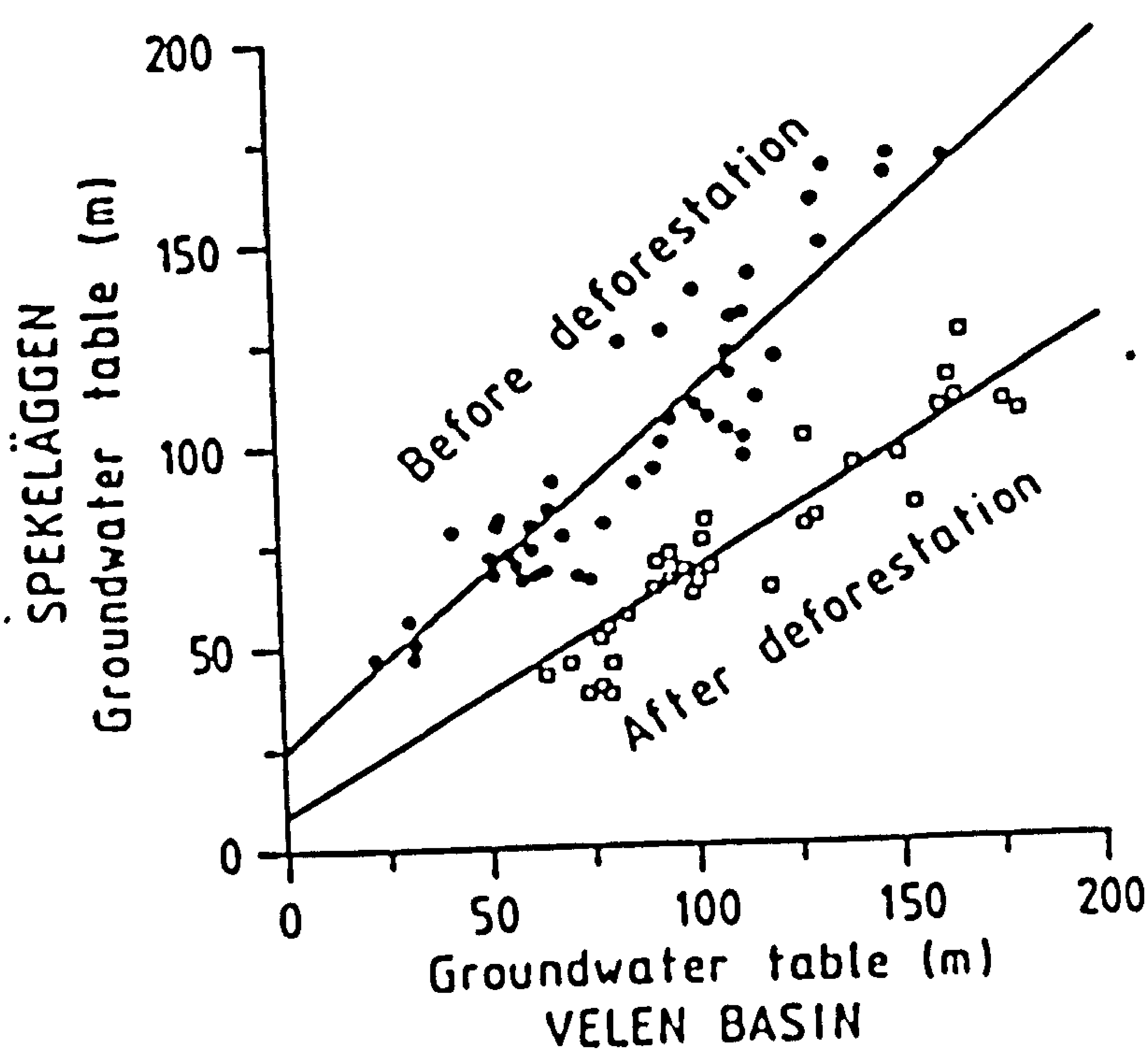


Figure 3.22. Relation between groundwater levels at Spekelaggen before (1968/71) and after (1972/74) deforestation achieved by regression against average levels in the Velen basin as a whole (after Anderson, 1988).

3.3.3.9. Summary

Infiltration and soil-water movement are perhaps the most important hydrologic processes, since they determine the rate and amounts of water available for surface and subsurface runoff, the amounts of water available for evapotranspiration and the rates and amounts of recharge to groundwater storage. Human activities however, have a direct effect on the ground surface, soil properties and the antecedent soil moisture condition. An alteration of any of these changes the natural water balance in the basin.

As infiltration rates and capacities are affected by the ground surface changing conditions, the ground water storage is directly and indirectly affected. First, the amount of water infiltrating and percolating into the ground water storage is varied and ultimately reduced. Secondly, the horizontal and vertical soil water movement is changed as a result of the subsurface interventions. This reduces the three dimensional soil water movement into the groundwater storage which in turn reduces the amount of water released as streamflow.

3.3.4. Human-induced Variation on Streamflows

3.3.4.1. General Variations

Streamflows from undisturbed forested watersheds are the net result of the physiography of the watershed and its climate. They provide a measure of the response of a watershed to the variable inputs and internal hydrologic processes. As well as runoff, streamflow is also referred to as stream or river discharge. The cumecs (1 m^3s^{-1}) and cumecs per sq.km are the commonly used units. For a clearer comparison with precipitation, streamflow is expressed as a depth equivalent over a watershed area.

In natural conditions, the watershed is regarded as a system receiving inputs of precipitation and transforming them into outputs of evapotranspiration and streamflow. Allowing for the temporal and spatial changes of storage within the system, inputs must be equalled by output. Because of the temporal variability illustrated in Figure 3.23 the annual streamflow hydrograph comprises short periods of sudden increased flows associated with precipitation and intervening, much longer, periods when streamflow represents the outflow from stored water in the watershed and when the hydrograph therefore takes the exponential form of of the typical exhaustion curve (Ward and Robinson, 1988).

Since water resources development represent a human intervention in the hydrologic system and streamflows being an integrated watershed output, it should provide a sensitive indicator of variations in rainfall. An examination of runoff variation in terms of mean annual flows during the period 1968-73 for nine gauging stations in Africa, reveal progressively higher deficiences as the drought progressed (Table 3.11).

Table 3.11. Variations of the annual deficiency of mean annual discharge during the 1968-1973 drought in Africa (from UNESCO-WMO (1985)

Station and country	Drainage area, km^2	Years of Observation	Mean Flow (m^3s^{-1})	Deficiency Excess (%) 1968	69	70	71	72	73	1974
Kori of Badegiucheri in Niger	825	35	1.34	-87	-57	+15	-77	-66	-47	-30
Ba Tha at Ati, Chad	45290	20	19.1	-74	-34	+87	-72	-75	-27	-24
Mejerdan at Jendouba in Tunisia	2414	76	7.21	-16	+5	-38	-3	+9	+131	-60
Shebelle at Malca Wacana, Ehthiopia	5290	16	27.5	-11	-1	+6	+5	-20		
Nile at Aswan, Egypt	-	107	2950	-13	-12	-2	-3	-31	-4	-3
Ubangui at Bangui Cameroon	203500	45	4678	-6	+10	-5	-13	-11	-14	-3
Niger at Niamey, Niger	700000	42	989	-3	+22	-18	-19	-25	-39	-9
Maevaranga at Amboassary, Madagascar	22430	24	600	-78	-3	+64	+61	-59	-45	-27
Zambezi at Maramba Zambezi	300000	67	1100	+102	+55	+14	-11	-30		

3.3.4.2.Streamflow-Rainfall Response to changes in vegetation cover

The relationships between vegetation and hydrology is a well established phenomenon. The vegetation intercepts some of the rainfall and much of this evaporates, reducing the amount reaching the ground interface. Vegetation also reduces water available for groundwater recharge through transpiration and yet it also increases groundwater recharge by reducing the overland flow. The quantity of water required for transpiration and maintenance of biomasses is large, hence the role of vegetation is important in hydrology. Figure 3.24 illustrates this observation in a relatively low runoff in unmodified vegetation in an African rainforest river basin.

The illustration above shows how small is precipitation contributing to runoff in unmodified rainforest. When annual precipitation is 1200 mm, less than 10% (100 mm) becomes runoff; with an annual precipitation of 1800 mm, runoff only increases to about14% (250 mm), hence land clearing and removal of vegetation would most likely increase the runoff values. Thus even with a high rainfall, runoff is still relatively small compared to total precipitation because of large interceptions and good groundwater recharge. Studies by Davy et al. (1976) provides even a clearer picture of the role of vegetation cover on annual runoff presented in Figure 3.25. In this representation, the rate of runoff in response to annual precipitation increases with increased degree of ground cover, with the bare rock (curve a) having the least runoff.

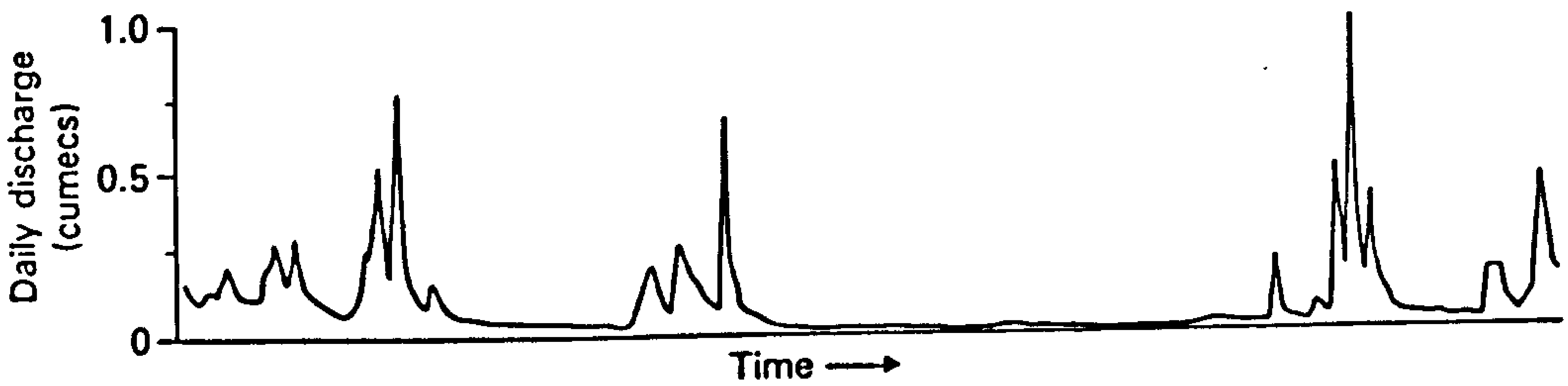


Figure 3.23. Annual hydrograph for the Catchwater Drain, North Humberside, 1967 (after Ward and Robinson, 1989)

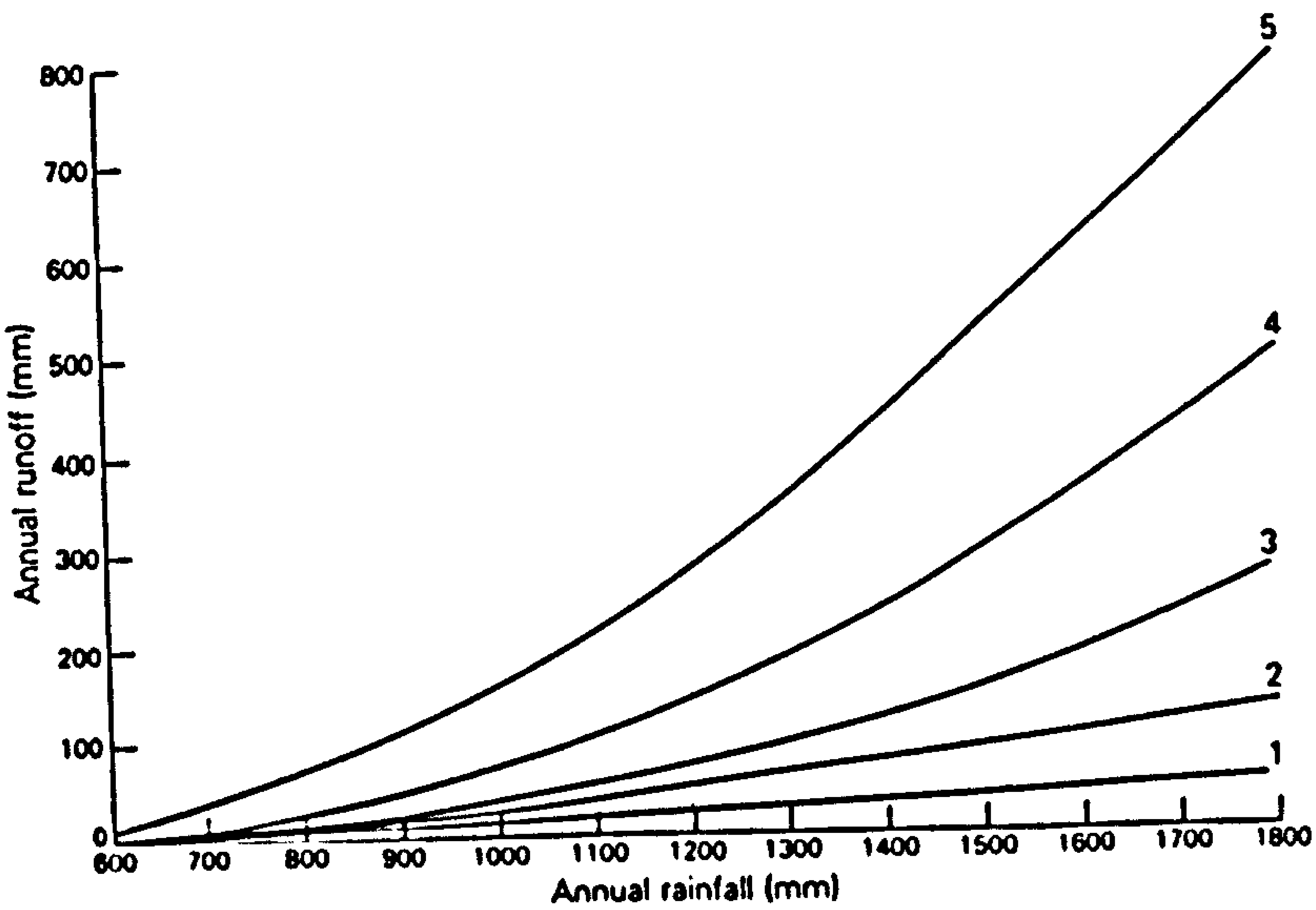


Figure 3.24. Annual rainfall-runoff relationships in some African rainforest environments (after Lewis and Berry, 1988)

Annual fluctuations or short-term changes in natural vegetation can also be related to runoff using vegetation indices. The best approach is using normalized difference vegetation index (NDVI) from remote sensing imagery. The different reflectance behaviour of soil and vegetation enable the determination of the indices as a way of measuring the proportion of vegetation. The principle is that, vegetation possesses a reflectance with a minimum in the infrared range (IR), and a maximum in the red range (R); whereas for uncovered soils, the reverse is true (Figure 3.26)

The simplest vegetation index (SVI) is obtained from simple linkages of infrared and red range. Perry and Lautenschlager (1984) provide detailed review of vegetation indices used with data from Landsat MSS and SPOT satellite imageries. This index is generated daily from the National Oceanic Atmospheric Administration (NOAA), Advanced Very High Frequency Resolution Radiometer (AVHRR) satellite data. A 3-day composite is produced on monthly basis. The SVI is represented in the ratio:

$$SVI = \frac{IR}{R} \quad (3.16)$$

where,

SVI is the simple vegetation index (dimensionless),

IR is the brightness values in the infrared channel (dimensionless),

R is the brightness value in the red channel (dimensionless)

and the Normalized Difference Vegetation Index as:

$$NDVI = \frac{IR - R}{IR + R} \quad (3.17)$$

where, NDVI is the Normalized Difference Vegetation Index (dimensionless).

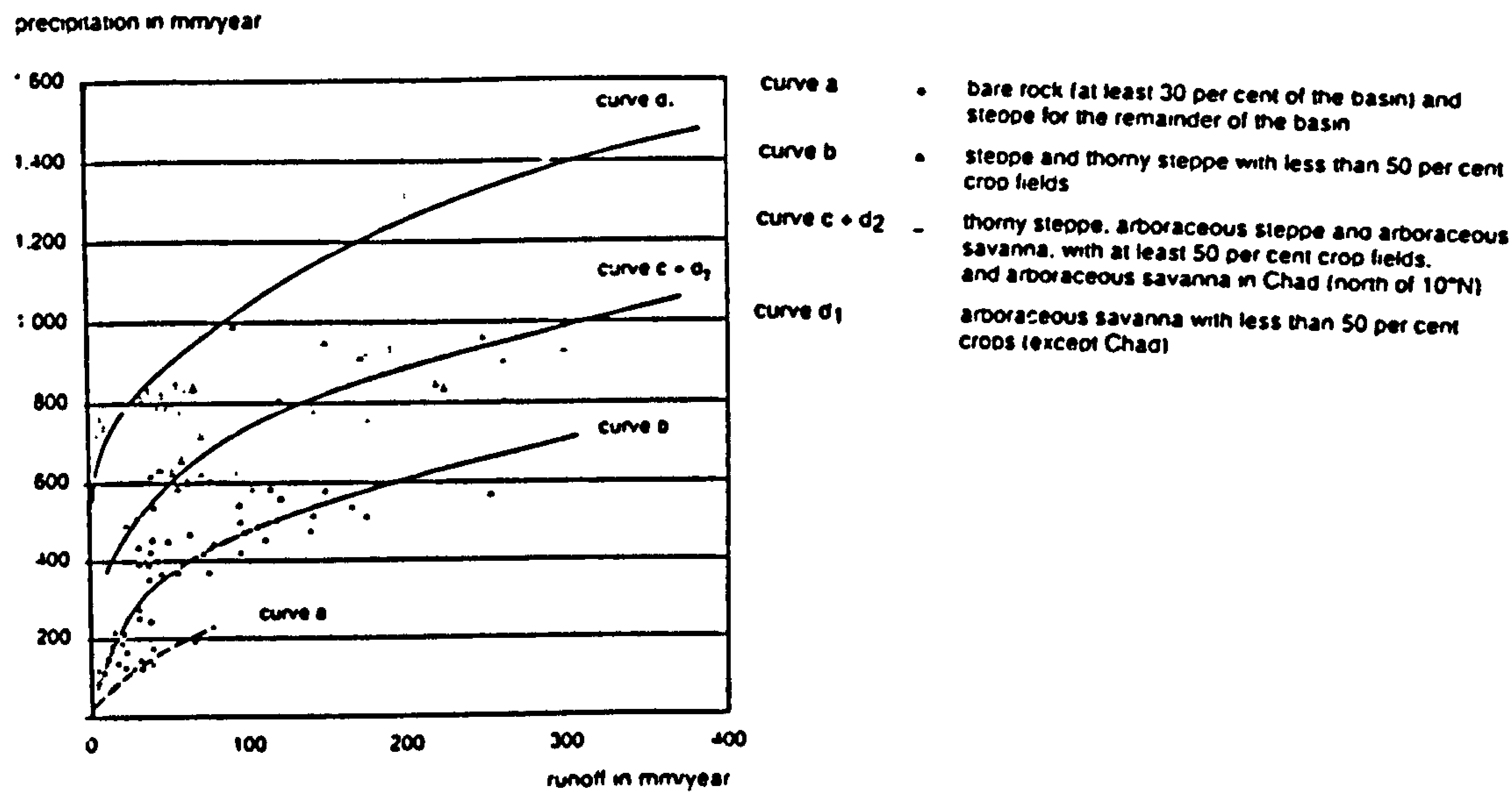


Figure 3.25. Relationship between rainfall, vegetation and runoff in West Africa (after Davy et al.,1976)

This approach of representation do not require ground truth readings (radiation behaviour, soil cover, leaf area index) hence it is the easier to establish the rate of vegetation cover change, which is related to the variation of streamflows given known rainfall amounts. The NDVI therefore was found suitable for global vegetation monitoring because it partially compensates for changing illumination conditions, surface slope and receiving angles. Clouds, water and snow have larger reflectance in the near infra-red, so for these features, the NDVI is negative. Rock and bare soils have similar reflectances in these two bands and results in vegetation indices are close to zero. In scenes with vegetation, the index ranges from 0.1 to 0.6, the highest value indicating greater density and greeness of plant canopy.

Atmospheric effects such as scatterring by dust and aerosols, all act to increase (R) with respect to (IR) and reduce the indices. The NDVI equal to zero is taken as an indication of a total absence of vegetation. Further discussion and results from field application of NDVI is provided in Scheneider et al. (1981). A relationship between the rainfall, runoff and changes in the values of the vegetation indices should therefore provide the influence of vegetation change on streamflows.

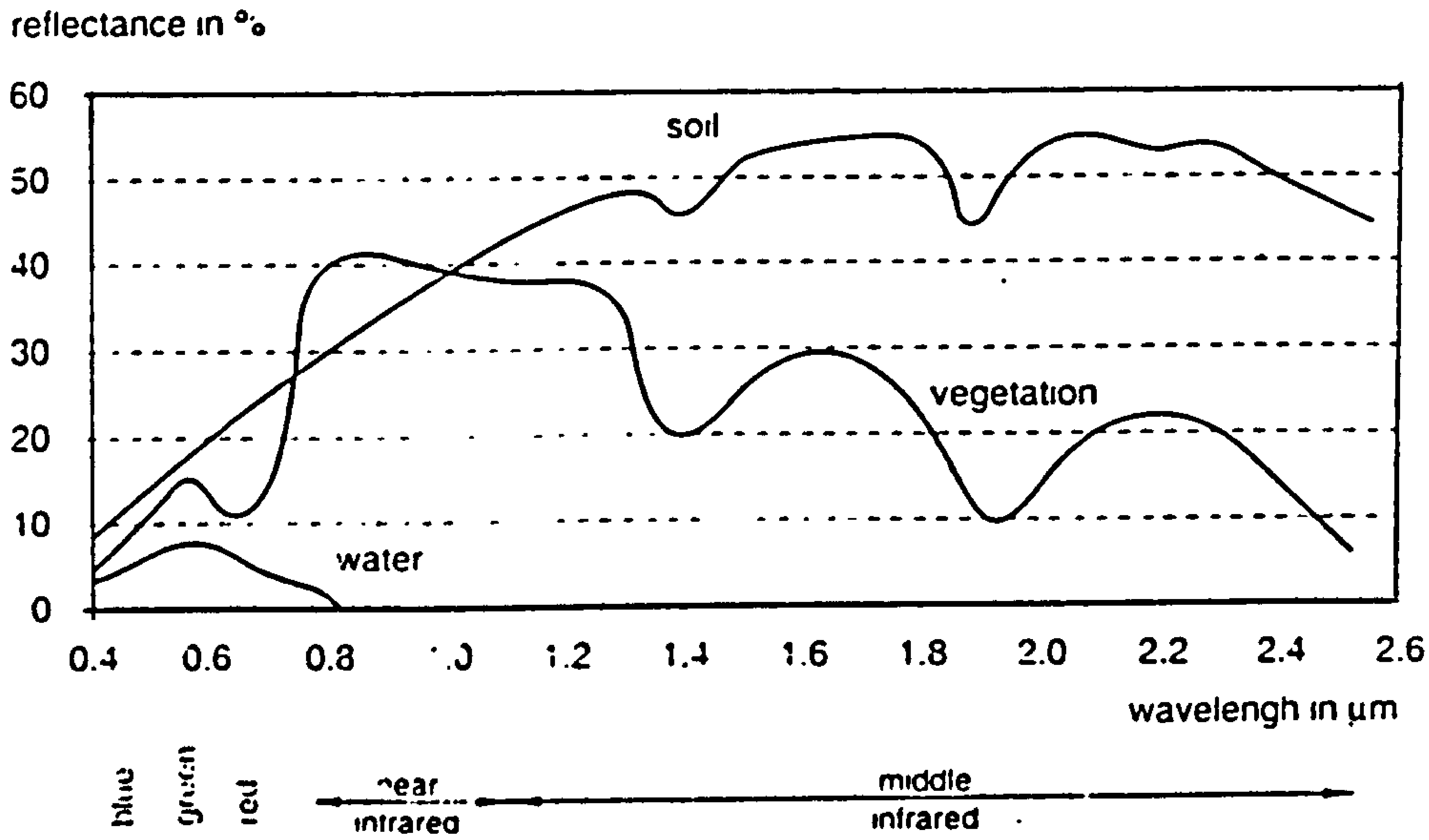


Figure 3.26. Spectral reflectance characteristics for vegetation, soil and water (after Richards, 1986)

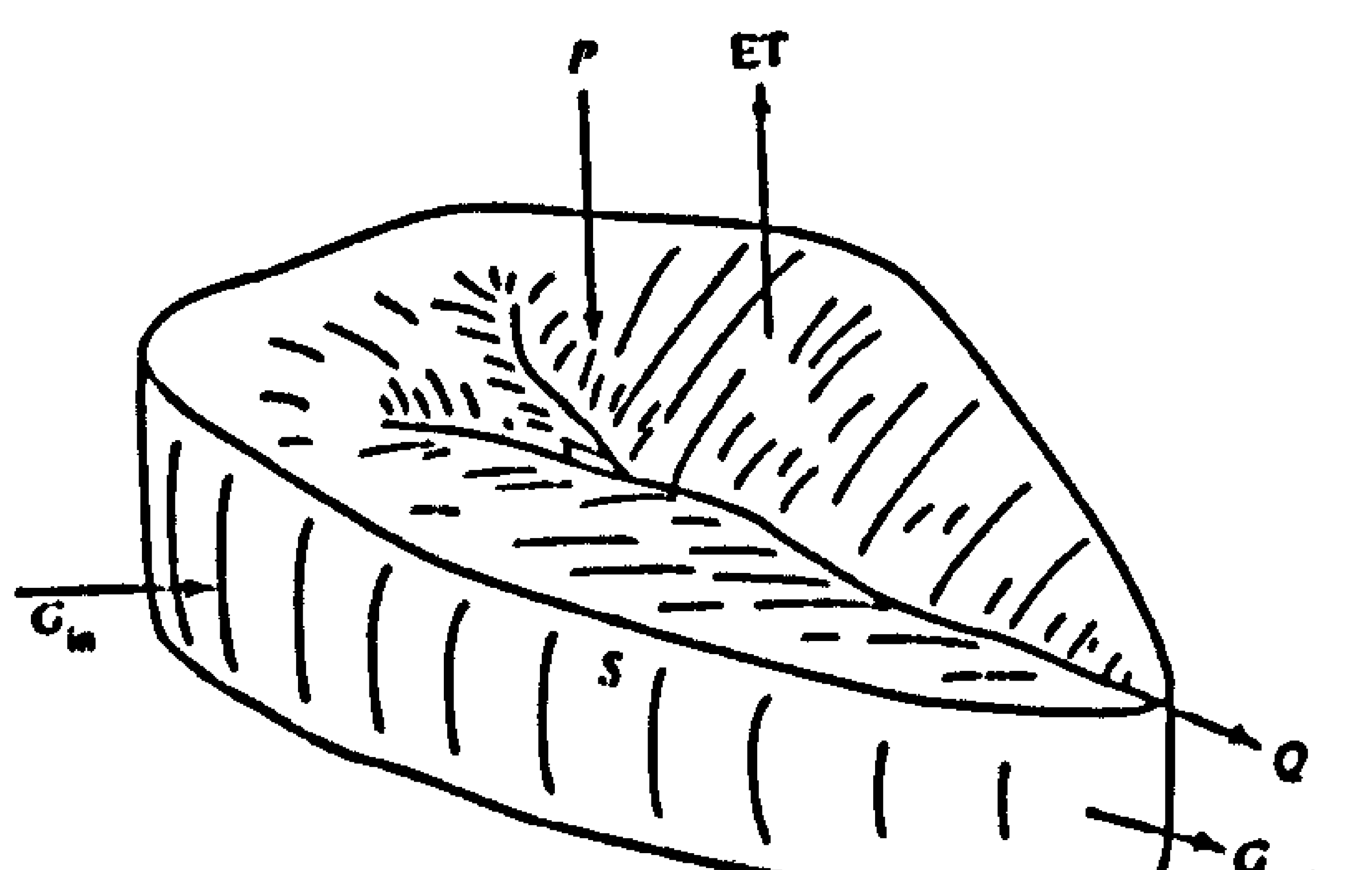
3.3.4.3. Summary on Streamflows

This section has summarized specific impacts of human activities on streamflows. Since the long-term average streamflow represents the upper limit on the available water for human use and management, it is rational to examine its historical variability. The search for this kind of information becomes a continuous process. Land and water development projects for example were established to influence the storage and distribution of water in the basin, resulting in rapid variations of streamflows. In the end these activities reduce and vary the total streamflow output from the watershed. There is thus a need to understand the watershed hydrologic conditions especially in areas experiencing extreme human activities so as to plan for a sustainable environment. Remote sensing technology provides this information both in the spatial and temporal dimension.

3.3.5. Variations in Watershed Water Balance

The water balance approach is useful in studying the behaviour of a watershed and the impact of physical changes on the system. The method identifies areas in which information on water dynamics in the hydrological system is inadequate. Determinations of the discharges and storages of water in the system thus becomes a quantitative means of differentiating advertent from the inadvertent changes. The discharges and storages change over time and the overall water balance provides a means of following sequences of these modification in the hydrologic system. The modification is then interpreted in terms of an altered opportunity for water to infiltrate, a change in its interception, an increased drainage of groundwater, a lengthened period of evapotranspiration, or an abrupt increase in water loss. The altered balances of these parameters then measures the direction and intensity of temporal and spatial human-induced modifications.

A conceptual water balance representation shown in Figure 3.27 can be used to define useful terms and show the importance of natural factors in determining the regional water balance. Over a long time periods (decades) in which there are no significant climatic trends or geologic changes and no anthropogenic inputs, outputs, or storage modifications, the idealized period is assumed to have zero net changes in storage (Dingman, 1994). Thus over a long period, the major controlling factors influencing the runoff is the residual between the precipitation and evapotranspiration.



Schematic diagram of a watershed, showing the components of the regional water balance: P = precipitation, ET = evapotranspiration, O = stream outflow, G_{in} = ground-water inflow, G_{out} = ground-water outflow.

Figure 3.27. Schematic diagram of an idealized watershed showing the elements of a regional water balance.

3.3.5.1. Water Balance Equations

Water balance equations can be used to estimate the evaporation from an open body of water where direct measurements are not possible. On the basis of this approach, it is possible to make a quantitative evaluation of water resources and their change under the influence of human activites. The water balance equation for any natural area or water body indicates the relative values of inflow, outflow and change in water storage for the basin. This can be represented as:

Water Input = surface runoff + Water loss + Δgroundwater storage
For a typical small watershed, the water balance can be characterized by an equation of the form:

$$P = ET + Q + O + \Delta G + \Delta S + I + M \qquad (3.18)$$

where, P is the mean watershed precipitation (mm), ET is long-term regional evapotranspiration (mm), Q is the runoff (streamflow) (mm), O is the groundwater outflow (mm), ΔG is the groundwater storage increment (mm), ΔS is the soil moisture increment (mm), I is the amount of water intercepted (mm), M is the water recharge or depletion due to human activity (mm)

For application in a variety of situations, the equation can be simplified or complicated depending upon the quantity and quality of the available data, the purpose of the estimation, the type of the watershed, its characteristics and the phase of the hydrological regime in which the water balance is required.

Because the components of the equation are naturally difficult to measure, these components are grouped together to leave only those which are easily measurable. A time interval of a year is often taken for water balance of a watershed because the annual values of the components tend not to vary as much as the daily and monthly values. Consequently, the water balance of a small watershed may be represented as:

$$P = ET + Q + SM + GW \qquad (3.19)$$

where, SM is increase in the soil moisture stored in the watershed (mm),
GW is increase in the groundwater storage in the watershed (mm),
and letting SUM be the sum of water stored in the watershed in the form of soil moisture (SM) and groundwater storage (GW), the equation now becomes:

$$P = Q + ET + \Delta SUM \qquad (3.20)$$

To evaluate changes in water stored in the watershed at certain periods of time, the equation is $\Delta SUM = P - Q - ET$. The evapotranspiration is estimated from hydrometeorological variables using the Penman equation (1948), the Morton (1983) conceptual equation or the Christiansen and Hargeaves (1969) where climatic data is inadequate. The runoff is obtained from the annual discharges and the rainfall data in most cases are available. For continuous quality data the equation has been used successfully in small watersheds with minimal changes in vegetation cover.

From the human view point, the long-term mean runoff is fundamental to its survival, because it represents the maximum rate at which water is potentially available for human management and use, and hence a measure of the ultimate water resources in a region (Dingman, 1994). Human attempts to modify this amount have consisted of the construction of storage reservoirs, modification of precipitation through "rainmaking" and modifying ET by modifying vegetation cover. This review has clearly shown that, climatic change whether natural or human-induced, will affect the runoff regime, hence, the availability of water resources. Most of the human-induced changes have feedback mechanisms such as the changes in hydrologic regimes which in turn affect climatic variables.

3.4. ANALYSIS OF HUMAN-INDUCED HYDROLOGICAL CHANGES

3.4.1. Introduction

Methods for evaluating potential impacts of human activities are still at the development stage. The analysis, interpretation, and presentation of the data is complex especially, when it must be evaluated for numerous and simultaneous impacts. It is even more intricate in analyzing historical data in order to infer the causal mechanism. Indices have been used to provide comparative information which aids the evaluation process. An index relates an observed value of selected components to the standard established for that component. It expresses the desirability or undesirability of the observed component in relation to man and his environment (UNESCO., 1984). Therefore indices of various forms are used to identify potential changes which have occurred due to human activities.

3.4.2. Residual Analysis

Detection of homogeneities in hydrological time series is used to infer cause and effect relations such as climatic and land use changes in watersheds. When using the double mass curve (DMC) analysis for example, inhomogeneities are identified with confidence if there is a discrete break in the time series plot. This however, is uncommon because hydrological changes are gradual in character and take several years for the effect to be clearly identifiable. Further, DMC is difficult to be statistically tested so that a significance of change is quantified and to enable inference of the causal mechanism.

A method that would distinguish between natural and systematic hydrological changes (Berneir, 1977 and Gottschalk, 1982) is what is required. A more practical approach is the residual analyses described in Lowing et al. (1987). The approach denotes the time series as X_i, $i= 1,2........, N$; with N years of observations and mean of the series ($\bar{x}$)

and standard deviation (S_x), the standardized deviation of X from its mean is represented as:

$$X' = \frac{X'_i - \bar{x}}{S_x} \quad \text{for} \sum_{i=1}^{N} X_i = 0 \tag{3.21}$$

A presumed homogeneous series denoted, (Y_i, i= 1, 2...., N) is introduced with its mean ($\bar{y}$) and standard deviation (s_y). The correlation between the series X_i and Y_i; for i= 1, 2....N is r. In terms of the standardized variables X'_i and Y'_i, the regression of X_i and Y_i becomes:

$$X' = r * Y'_i + \sqrt{(1-r^2)}e'_i \tag{3.22}$$

where e'_i is the residual (random) and on re-writing, it takes the form,

$$e'_i = \frac{(X'_i - r * Y'_i)}{\sqrt{(1-r^2)}} \tag{3.23}$$

Assuming a normally distributed residuals, Lowing et al.(1987) transformed the cumulative sum of residuals to a discrete approximations of ß(t) of Brownian motion by dividing it by ($\sqrt{N}$.

$$\beta(t) = \frac{\sum_{i=1}^{Nt} e'_i}{\sqrt{N}} \tag{3.24}$$

where, t= i/N for 0 < t < 1, which has a zero mean and standard deviation equal to $\{\sqrt{(t(1-t))}\}$ (Karlin, 1968). The probability distribution of the maximum absolute value of ß(t) can now be determined as recommeded in Darling (1967):

$$P = \{|\max \beta(t) \le l\mu|\} \beta(1) = 0) = 1 - \mu = K(l\mu) \text{ for } 0 \le t \le 1 \tag{3.25}$$

where $K(\mu)$ is the Kolmogorov function and ($l\mu$) is the critical value of a test quantity with significance level , tabulated in statistical tables. The hypothesis of homogeneity in the data series are rejected at the 95% significance level.

3.4.3. Homogeneity Analysis

Spatial homogeneity tests of data series are carried out to establish the change of hydrological condition of controlled watersheds and compare with other areas to establish the extent and magnitude of human-induced hydrological changes. A stable rainfall-runoff relationship is used to search for evidence of nonhomogeneity of hydrological data series. The identification of the nohomogeneity can then be used to examine the watershed and its changing performance in generating streamflow with time: ' a cause-effect relation'.

For example UNESCO (1987) used this approach to examine the effects of Yuechen reservoir in Zhangie, China on stream flow changes. The study assumed that human activities had a negligible effect on the annual precipitation pattern in the reservoir basin. Thus, if any shift is observed in the rainfall-runoff relation, it could be interpreted as a nonhomogeneity existing in the streamflow data. The cause is then inferred from the historical land and water use rates in the basin.

The methodology also enables the analysis of the rainfall-runoff relations at different time periods, in order to establish the cause of the non-homogeneity of the rainfall-runoff series prior to and after the event. Figure 3.28 shows the curves obtained in the Yuechen reservoir effect on the rainfall-runoff regime. Three curves are distinguishable and are annotated with the year of observation.

From these curves, curve (A) reflects the normal characteristics of the basin during the period 1951-1957, before the construction of the hydroproject. Curve (B) moves to the left, indicating less runoff for given amount of rainfall with respect to control curve (A). During this period (1958-1965), a large number of reservoirs were constructed in the basin, hence this behaviour of the rainfall-runoff relation may be assumed to have been caused by the construction of the hydroprojects. The third curve (C) for 1966-1972 period, moves even further to the left, indicating, further reductions in runoff, possibly because of excessive development of irrigation systems which reduced runoff feeding the reservoir.

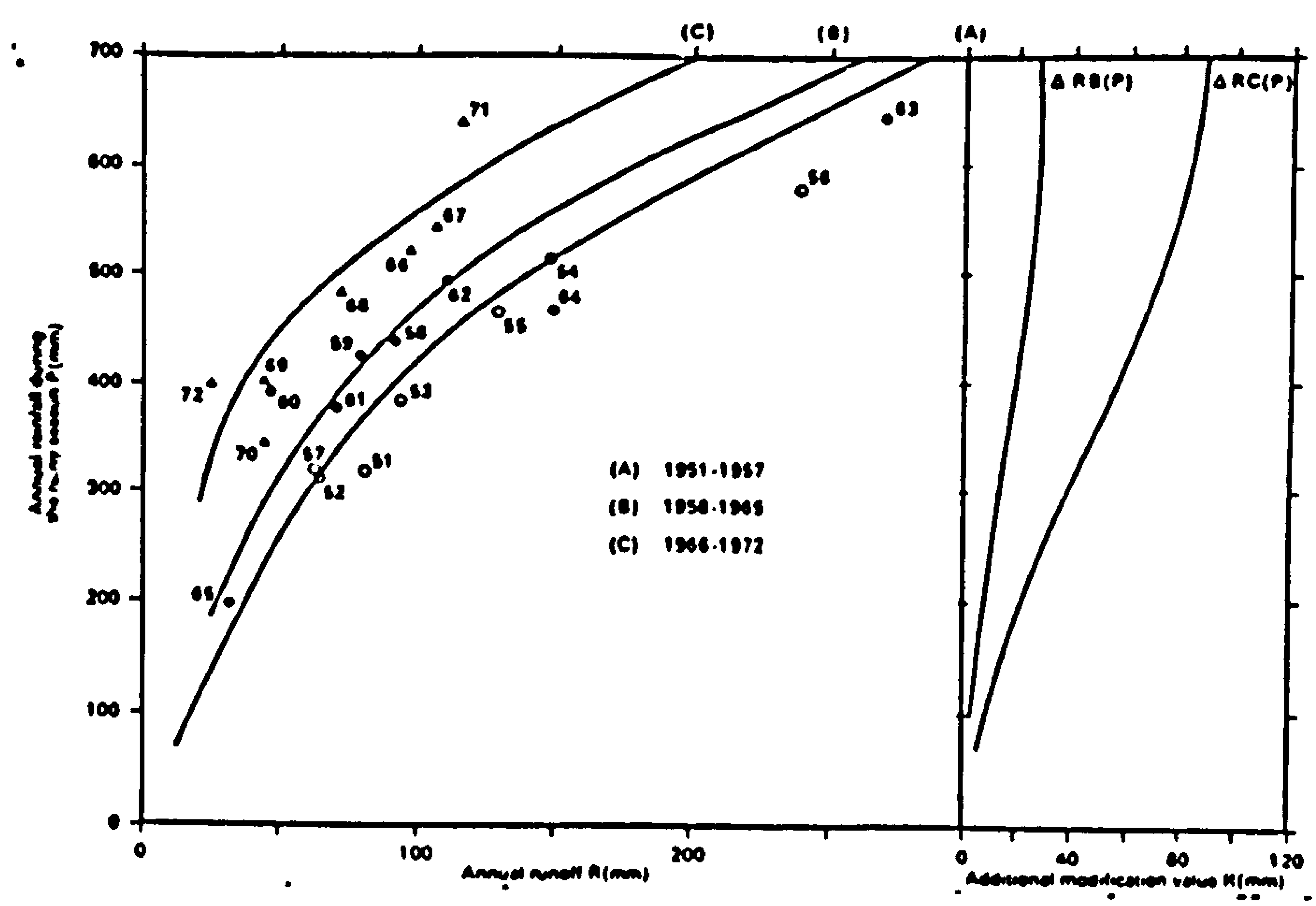


Figure 3.28. Relation between rainy season rainfall and annual runoff, Yuechen reservoir in China (UNESCO No. 42, 1987).

Thus, the curves have indicated that the non-homogeneity existing in the annual runoff series, in which the cause and effect relation can be established. If an estimation is required, the differences between the assumed control curve (A) and curves (B) and (C) provides correction factors for the reconstruction of the historical and natural flow series, of the form:

$$
\begin{aligned}
CRA(P) &= RA(P) - RA(P) = 0 && \text{for } 1951\text{-}1957 \\
CRB(P) &= RA(P) - RB(P) && \text{for } 1958\text{-}1965 \\
CRC(P) &= RA(P) - RC(P) && \text{for } 1966\text{-}1972
\end{aligned}
\tag{3.26}
$$

where,

RA(P), RB(P), and RC(P) are the runoff values corresponding in each rainfall-runoff relation to rainfall, P and CRA(P), CRB(P), and CRC(P) are the additive correction factors (UNESCO, 1987). Using this reconstruction procedure, it is possible to obtain flow series of the river which would have occurred if the upstream reservoir and irrigation schemes had not been developed. The procedure is applicable in areas were rainfall variablity is considered insignificant and other developments such as groundwater exploration is not excessive.

3.4.4. Homogeneity Tests on Means and Variances

Homogeneity tests on the means and variances are used to test whether two hydrological series or partial periodic series should be regarded as having originated from the same parent population in order to detect a trend or periodicity of the data series. For a certain aspect (eg. trend, variance, periodicity etc), this test statistic can be used. Objective homogeneity tests of a series can also be carried out more effectively if the historical processes in the watershed are fully understood and known. Tests of the means and variances of the data at identified partial periodic series enables the isolation of the cause-effect relation.

The homogeneity tests begin with the estimation of the sample variance using the Fisher's ratio (F). The hypothesis about the homogeneity of the sampling variance is either rejected if F greater $F\mu$ or accepted if less (where $F\mu$ is the critical value). If accepted, the analysis assumes that the observed data do not contradict the suggested hypothesis. Using this method to estimate the degree of the impact of land use changes, upon the long-term streamflow in the Naiba river in Soviet Union UNESCO (1987) used the Fisher's ratio test computed from the equation:

$$F = \frac{S_x^2}{S_y^2} \tag{3.27}$$

where, S^2x and S^2y are the sample variance, the larger variance being taken as the numerator. X and Y are two different hydrologic series. The critical value of F_μ is taken from statistical tables, as a function of the sample size nx = ny; the adopted significance level, μ; the lag one autocorrelation, r1; and the correlation between the series, (rho). Since F-test is only valid if the data are normally distributed, the data series are first normalized.

Secondly, homogeneity tests of the means of unequal hydrological data series are carried out using the students' t-test, computed from the following representation (Chebotarev, 1974).

$$t = (\bar{x} - \bar{y}) \frac{(n_x - n_y(n_x + n_{y-2}))}{(n_x + n_y) + (n_x - 1)S_x^2 + (n_{y-1})(S_y^2} \qquad (3.28)$$

where, (x) and (y are the sample means, the critical values of the test in statistical tables. Detailed procedures for homogeneity testing of hydrological data is provided in Rozhdestvensky and Chebotarev (1974., UNESCO, 1987).

3.4.4.1. The Barlett's Test (1937)

Other tests in common use are those developed by Barlett (1937), Hartley (1950), and Spearman rank correlation test (Siegel, 1956) These are specifically used to examine whether hydrological data series could be considered as homogenous with regard to the variances. The data series are divided into several or few annual partial series, whose variances, var1, var_2,....var_n are estimated. The null hypothesis is that, the variances of k samples are equal to the general value of var2. The unbiased estimate s_2 of var2 is calculated according to:

$$s^2 = \sum_{i=1}^{k} (n_{i-1}) * \frac{s_i^2}{\sum_{i=1}^{k} (n_{i-1})} \qquad (3.29)$$

where,

n_i is the number of elements in the ^{i}th sample,
s_i is the variance of the ^{i}th sample.

Subsequently, the functions, M and C are calculated thus:

$$M = \sum_{i=1}^{k-1} (n_{i-1}) * \ln(s^2) - \sum_{i=1}^{k} (n_{i-1}) * \ln(s_i^2) \qquad (3.30)$$

and

$$C = 1 + \frac{\left\{\sum_{i=1}^{K}\left(\frac{1}{n-1}\right)\right\}}{3(k-1)} - \left\{\frac{1}{\sum_{i=1}^{k} n_{i-1}}\right\} \tag{3.31}$$

if $n_i = n$ for all samples, then,

$$s^2 = \frac{\sum_{i=1}^{k} s_i^2}{k} \tag{3.32}$$

$$M = (n-1)(k * \ln(s^2) - \sum_{i=1}^{k} \ln(s_i^2) \tag{3.33}$$

and

$$C = 1 + \frac{(k+1)}{3k(n-1)} \tag{3.34}$$

The test statistic then is: $B = \frac{M}{C}$ (3.35)

For n greater than 6, this value follows approximately, the c^2 distribution with k-1 degrees of freedom if the null hypothesis is true. The c^2-distribution with v degrees of freedom is defined by its probability density function:

$$F(c^2) = \frac{(c^2)^{n/2-1 \exp(-c^2/2)}}{2^{n/2} E'(n/2)} \tag{3.36}$$

for $\quad 0 < c^2 < 1$

The distribution function $F(c^2)$ is tabulated in the International Standards Organisation, (ISO 1979). The c^2- value is compared with the calculated value of B. If the c^2 value is smaller than B then the null hypothesis is rejected and if larger, then the hypothesis that the variances do not differ significantly is not rejected.

3.4.4.2. The Hartley's Test (1950)

The Hartley's (1950) test is also used to test the homogeneity of the variance of hydrologic data series. It tests whether the variances of a number of samples can be considered to be equal. It is however, only applicable to data series of equal time periods. The test statistic is:

$$\phi = \frac{s_{max}^2}{s_{min}^2} \tag{3.37}$$

where;

s^2_{max} is the maximum sample variance, s^2_{min} is the minimum sample variance. Hartley (1950) tabulated the values of the distribution function F(Ø) =0.95 and F(Ø) =0.99 for a number of samples k= 2,.....12 and for a number of sample sizes, *m*. For example, for k= 5 and *m* = 40 can be found:

$$F(Ø) = 0.95;\ Ø = 2.53 \qquad (3.38)$$
$$F(Ø) = 0.99;\ Ø = 3.10$$

Comparing the calculated and tabulated values at 95% and 99% level of significance, and if the calculated values are larger, then the null hypothesis that all variances are equal is rejected and vice virsa. Different values of k and *m*, are presented in Hartley's Tables (Hartley, 1950).

3.4.4.3. The Spearman's Rank Correlation Test (1956)

Ranks are also employed to determine the degree of association between two random hydrologic series, X and Y. Instead of the actual values, the rank of the variable magnitudes are used. This approach is known as the Spearman's rank correlation test (Siegel,1956). The associated correlation coefficient is calculated as follows:

$$R = 1 - \frac{6R}{n1(n2\text{-}1)} \qquad (3.39)$$

where;

n is the number of elements in each series,

R is the number of the squares of the differences between paired rank numbers R_x and R_y of both the hydrologic series.

hence,

$$R = \sum (Rx - Ry)^2 \qquad (3.40)$$

The sampling distribution given for rR only holds under quite restrictive assumption that the joint distribution for X and Y is normal. Rank correlation methods overcomes this limitation and also demonstrates the stronger attribute of measuring certain relationship that are not linear.

The coefficient rR is used as the test statistic. The test hypothesis is " no correlation between the hydrologic series". For n >9, the distribution function of the coefficient rR when the null hypothesis is true is approximated by Student's t-distribution. For n>40, it is approximated by the normal distribution with mean of e(rR)= 0 and a variance, var(rR)= 1/(n-1). Using this test, it is possible to detect trends of discontinuities in the series under examination.

To each year of the hydrologic series, a rank is assigned, the sum of squares of the difference between the ranked numbers of successive years are computed. The critical values for a significance level of 0.10 for example is given by Siegel (1950) as:

$$F(°R) = 0.05,\ rR = -1.645*\sqrt{Var(rR)} \qquad (3.41)$$

$$F(°R) = 0.95,\ rR = +1.645*\sqrt{var(rR)} \qquad (3.42)$$

If the calculated correlation coefficient (rR) falls within these limits, it means that, between the two series, no statistically significant correlation is present; thus within the series itself, there is no serial correlation at the lag under consideration. Values of Spearman correlation coefficients are given in most statistical tables.

3.4.4.4. Summary on homogeneity tests

The homogeneity amongst data series is difficult to achieve because of the inherent errors resulting from the data collection procedures, data processing and the noise naturally masked in the data. However, the mean values should group the data into a population sample which can represent its actuality. Consequently, it is possible to compare the homogeneity of these means at different time periods to establish existence of temporal changes. From this, it is possible to infer the causes of the observed changes between selected subperiodic means of the data series.

Secondly, this approach enables the detection of trends exhibited by hydrologic series and hence isolate the cause-effect relation. The review has provided the possibility of its use to detect inhomogeneities in hydrological series due to changes in climate and land use. Spatial homogeneity tests are also used to establish the differecence in hydrologic conditions of controlled and disturbed watersheds.

3.5. TIME SERIES ANALYSIS

3.5.1. Introduction

Time series analysis of hydrological data is an analysis in which time is an independent variable. The purpose is to formulate and calibrate models that can describe the time-dependent characteristics of the hydrological variable and also to predict future values of the time-dependent variable (McCuen and Snyder, 1985). Methodologies used in the time series are also used to analyze spatial hydrologic data series of a watershed, in which case, space is referred as the independent variable.

Time and space are not causal properties rather, they are parameters by which true cause and effect is established. Time series are also analysed to separate the systematic and non-systematic variation, and characterize the time dependence with the systematic component. The first step in examining a time series is to plot the data on a time scale, which then allows the general properties like stationarity, to emerge and show up the presence of outlying values which need to be evaluated.

A time series may be considered as the sum of three components: a trend component, a periodic or cyclic component, and a random component. Any one or combination of these components may exist in a particular time series, hence it is necessary to examine and consider the properties of each component in order to understand its analysis.

3.5.2. Persistence

Hydrologic data series is bound to exhibit some degree of persistence because of the large inertia of some processes within the hydrologic system. Since local human activities distort this phenomenon, by modifying smoothness, and sometimes the persistence is lost in random fluctuations, the natural persistence of the data series is examined to establish points of change in persistence which can be related to human interventions.

Attempts to quantify and test for persistence fall into two main groups. Firstly, sequences of annual hydrologic data series are used to compute its first order serial correlation coefficient (r_{k1}). This is then tested against the null hypothesis that the sample derives from a population whose true rk1 is zero (i.e statistically independently (Yevjevich, 1964), WMO (1966a), Brunet-Moret (1975) and Sonnga (1979). The most robust procedure is that adopted by Brunet-Moret(1975) as the coefficient, A of the first order Markov autoregressive equation:

$$X_{i+1} = AX_i + Z_{i+1} \tag{3.43}$$

where, X_i, X_{i+1} are the hydrologic totals in years (i) and (i+1) years, and Z_{i+1} is a random term. In investigating the causes of variation in A, a trend line is drawn to identify the series with least persistence.

Secondly, the statistics of runs are used to analyze persistence, i.e, the length of successive years of below or above average conditions. Detailed description of the tests are provided in WMO (1966b) and Clarke (1973). These tests in most cases conform directly to the subjective impression of droughts. The conclusions drawn, are used to establish the cause-effect relation in the hydrologic series by examining the changed trends in long-term persistence. The serial correlation coefficient, r_k, corresponding to any lag, k, (k= 1,2,......n/4) is computed the formula:

$$r_k = \frac{\sum(X_i - X)(X_{i+k} - \overline{X})}{\sum(X_i - X)^2} \tag{3.44}$$

where, X_i is the hydrologic variable in the ith year,

$\bar{x}$ is long-term mean value of the variable,

r_k is the serial correlation for any lag (k)

The 95 % confidence intervals for r_k is computed from the equation (Yevjevich, 1972) of the form:

$$\frac{-1 + 1.96(n-k-1)}{(n-k-1)} \tag{3.45}$$

3.5.3. Trend Component

A steady and regular movement in a time series through which the values are either increasing or decreasing is termed as a trend. Trends appearing in an hydrologic series are part of a low frequency oscillatory movement induced by climatic factors or through changes in land use and catchment characteristics. Trends of this nature are only discernible when the data series is of a long duration. It is possible to use the ordinary least squares (OLS) procedures to represent linear as well as higher order polynomial trends as:

$$xT(t) = x_o + a_1 + a_2t^2 + \dots a_nt^n + x \tag{3.46}$$

where,

x is a residual term,

t denotes time,

$xT_{(t)}$ is the observed value of the variable when time is discrete,

a_1, a_2.. a_n are the coefficients chosen to minimize the equation.

By studying one or a few characteristics at a time whilst eliminating or reducing the effect of the others, a better picture of the process is gained. The filtering methods often used are the moving average and variate difference.

3.5.4. The periodic component

The periodic type of time series are common in hydrological data. Rainfall, runoff, and evaporation rates for example show periodic trends. When a periodic trend is expected, the period of the trend is identified using a moving average analysis, a correlation analysis, or spectral analysis. Once a periodic trend has been identified in the data series, a functional form can be used to represent the trend. The trend consists of one or more harmonics which combine to form a time-varying function whose wave repeats itself at regular intervals such that:

$$xF(t) = x(t \pm nT) \tag{3.47}$$

where, T is the estimate of the frequency component. The harmonics is described by the Fourier representation,

$$xF(t) = \sum_{k=1}^{m} (A_k \sin 2\pi f_{kt} + B_k \cos 2\pi f_{kt}) \tag{3.48}$$

where, xF(t) is the estimate of the frequency component,

A_k, B_k are the Fourier coefficients of the kth harmonic,

f_k is the frequency (cycles per time unit)

k is an integer identifying the harmonic,

m is the total number of harmonics considered non-negligible.

The frequency is related to the period of the first harmonic $f_1 = 1/T$, and for other harmonics by $f_k = 1(T/k)$. The frequencies observed in environmental phenomena may have periods anywhere from a fraction of a day to a year (James, 1993).

3.5.5. Random or stochastic component

The stochastic component does not have a deterministic pattern of behaviour and hence it can be predicted only in its statistical properties:

$$xR(t) = Z_n \sigma R \tag{3.49}$$

where, Z_n is the standardised normal variate with zero mean and standard deviation unity, όR is the standard deviation of the random component of the time series.

A combined time series is represented by a relationship that combines the three components (Raudkivi, 1977; McCuen and Snyder, 1985 and James, (1993) in the form:

Total time series = trend (T) + periodic(F) + random (R)

$$X(t) = xT(t) + xF(t) + xR(t) \tag{3.50}$$

3.5.6. Filtering Techniques

Filtering techniques are used to extract information inherent in hydrologic time series in order to establish the cause-effect relationship influencing the parameter under investigation. Moving average filtering for example is used to separate the systematic and non-systematic variations. The technique is based on the premise that, the systematic components exhibit some autocorrelation while the random fluctuations are not autocorrelated (McCuen and Snyder, 1985). Hence, the averaging of adjacent observations will eliminate the random fluctuations, with the result converging to a close description of the systematic trend. In general, the moving average computation uses the weighted average of adjacent observations to produce a new time series that consists of the systematic trend. Consider for example, a time series Y_i, whose filtered series , Y'_i is derived by:

$$Y'_i = \sum w_j Y_{i-k+j} \quad \text{for } i = (k+1), (k+2), \ldots (n-k) \qquad (3.51)$$

where,

n is the length (period) of measured time series (years, months,days),

m the length of smoothing interval,

w_j the weight given to the jth value in the smoothing interval,

$$\text{k is given by: } k = \frac{(m-1)}{2} \qquad (3.52)$$

The smoothing interval should be an integer, and a total of 2k observations are lost. The simplest weighting scheme would be the arithmetic mean (wj=1/m). Most filters assume equally spaced measurements in the time series, thus the periodic variations are eliminated when the smoothing interval equals the period or periodic component.

To demonstrate the use of the moving average smoothing to detect secular trends in streamflow data due to urbanisation, McCuen and Snyder (1985) used annual flood series of the Pond Creek watershed in Kentucky, USA. The example presented in Figure 3.29. shows that between 1946 and 1966, the percentage (%) of urbanization increased from 2.3 to 13.3% and the degree of channelization increased from 18.6 to 56.7% with most of the changes occuring after 1954.

The smoothed series have a trend and relatively little variation in the smoothed series prior to 1954. After a significant development in the mid 1950s, flood peaks appear larger, as evidenced from the nearly linear trend in the smooth series. The filtering , does not give an assurrance that the trend evident in the smooth series is the result of urban development alone because, several activities, occur simultaneously. Once a trend is established, a reason for it is identified or the assumed cause is evaluated. In the absence of any other cause mechanism, then the presumed cause is held true with statistical tests supporting the findings.

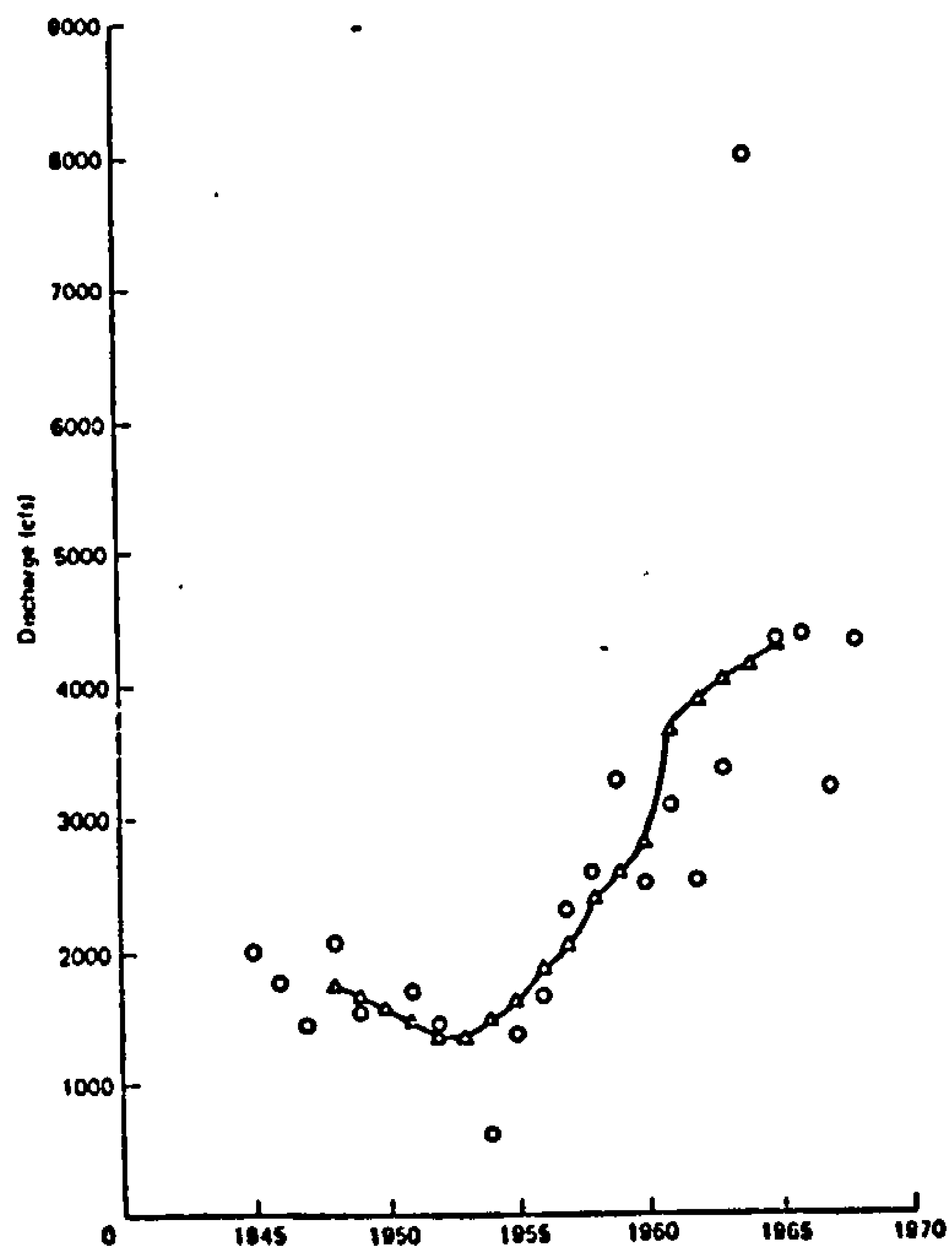


Figure 3.29. Annual flood for the Pond Creek watershed, 1945-1968 (1) and annual flood smoothed series (2), Kentucky, USA (after McCuen and Snyder, 1985)

The best application of this approach is on the decomposition of the time series (Hirsch et al., 1990). The technique makes it possible to decompose the overall variations in a time series of the flows into three components; the seasonal, the trend, and the irregular (random) components. The decomposition is based on judgemental factors relating to the choice of the filters used to smooth out the trend and seasonal components. Decomposition of monthly flows is accomplished in a mathematical model of the form:

Monthly flow = trend*seasonal*irregular(random)

Hirsch et al. (1990) studying the effects of human actions on hydrologic systems in the USA used the decomposition techniques on flows of the Colorado river immediately below the Hoover Dam in Arizona and Nevada. The methodology revealed several important features (Figure 3.30) which can be attributed to human development and use of natural resources in the river basin. There was for example no regular trend prior to 1935, after that, Lake Mead was completed and storage of water started. In 1963, Glen Canyon dam was completed which further regulated flows after 1963. The trend component shows variation in the order of 1 to 2 upto 1935, and a low flow period in 1935/43 (period of filling dam). Between 1944 and1963, large variations existed but of longer durations than prior to 1935. With the commissioning of the Glen Canyon dam, the trend shows an unvarying record of low flows.

These examples have shown that time series analyses of hydrological data can be used to reveal the presence or non-presence of trends, periodicities and random behaviour. The causal mechanism can then be related to the human intervention on the natural flow regime. Similar approaches have been shown to work elsewhere. Petts (1984) reviewed these extensively and used the approaches to analyse the effects of human-impounded rivers on hydrological regimes and ecosystems.

Current knowledge enables us to manipulate some components of the river systems to minimize impacts resulting from upstream river impounding, although such 'environmental engineering' approaches are as yet of limited success (Petts, 1984).

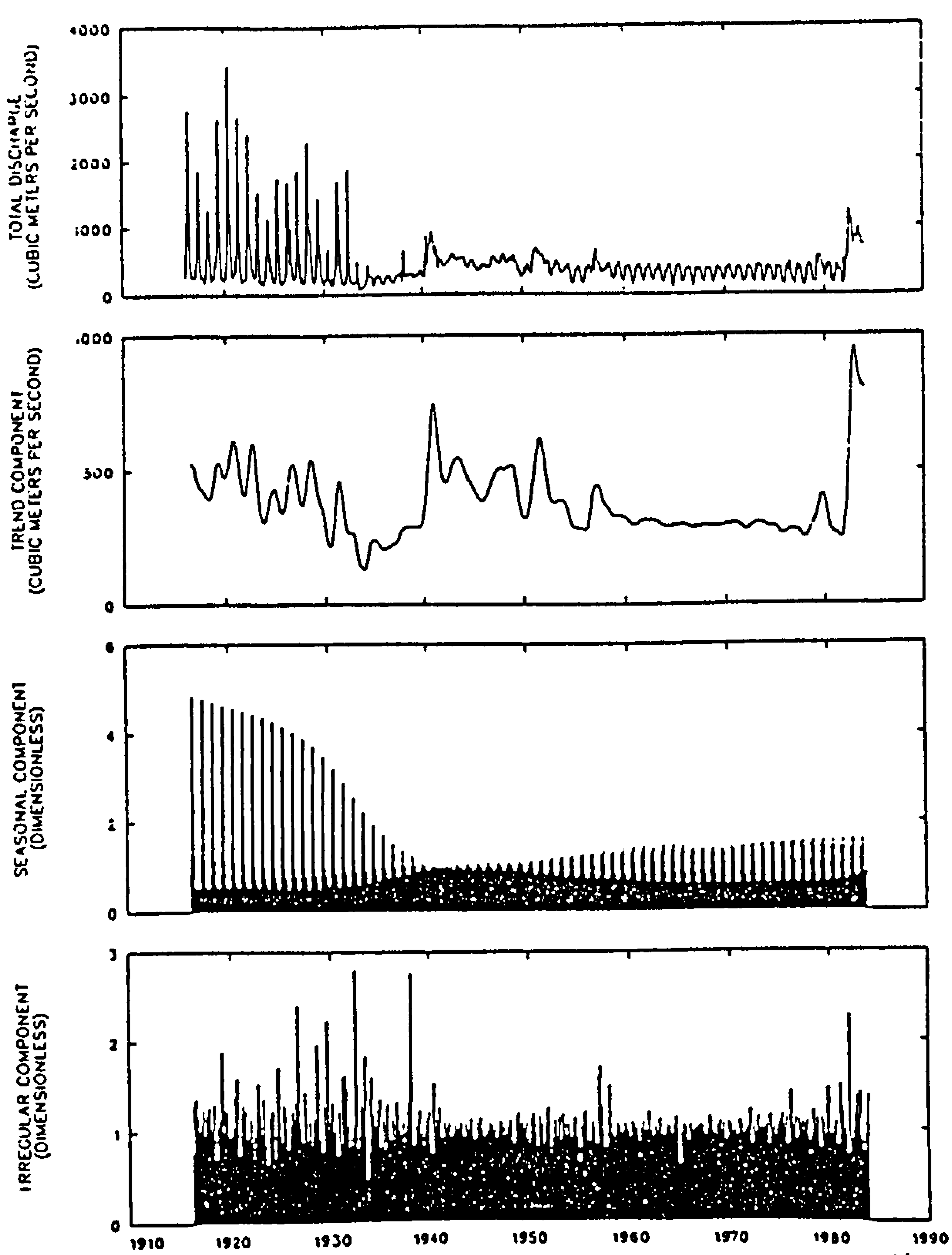


Figure 3.30. Time series decomposition of monthly flows for the Colorado River at Topack, Arizona, below Hoover Dam (constructed in 1935) and Glen Canyon (constructed in 1963)...(after Hirsch et al. 1990).

3.6. CONCLUDING REMARKS

1 Anthropogenic changes on streamflows may be estimated quantitatively from a knowledge of the long-term variations of runoff in combination with the analysis of the natural fluctuations of meteorological factors and of the economic development in the watershed. Such an approach allows for an estimation of the integrated influence of major anthropogenic factors within the basin and may serve as a basis for objective evaluation of water resources. The approach however is limited in scope because it does not reveal the physical nature of the processes and there is no way to distinguish the influence of each factor independently. No possibility of estimating for certain, the role of those anthropogenic factors which are not sufficiently marked

•2 Quantitative estimation of human influence upon streamflow, even if reliable long-term observational series on the hydrometeorological regime are available, is rather difficult because human modifications are superimposed on natural runoff variations which exceed the magnitude of the artificial changes. This is however solved by developing a long-term correlation between runoff and the natural runoff formation factors and the indices which characterize quantitatively the degree of the economic development of the watershed. Taking into account the regional physiography, the economic development and the extent of the utilisation of water resources, various indices and approaches may be chosen to determine the role of the anthropogenic factors.

•3 The value of runoff at a given place and within a given period of time depends on the volume of rainfall which has fallen during that period of time and in preceding periods. It also depends upon the amount of water transferred from the given basin to another, and transferred from each period of time to a later period, and on the losses of precipitation due to evapotranspiration and human use. This behaviour is influenced by the precipitation regime, by the physical watershed characteristics, geographic and geologic features, and by climatic conditions. Given a function representing these factors and the hydrological conditions of the watersheds in the present and preceding periods, it should be possible hypothetically to calculate the runoff of a watershed at a chosen site and within a prescribed period of time.

•4 The significance of the results and limits of applying models to assess changes in the hydrological regime due to human activity is acknowledged. There are however, often difficulties concerning the availability of-and-reliability-of hydrological records prior to and after the watershed disturbance as well as the manner in which the operation was conducted. If the data are available, their accuracy is likely to range between limits of ± 20%. If therefore the changes induced by human activities fall within these limits, the results of applying such models should be treated with caution. The results obtained and quoted by several authors concerning the effect of human impact on the hydrologic regime therefore, are sometimes different or often contradicting.

•5 The quantitative estimate of the hydrological changes due to human activities is based on methods ranging from simple graphs and double mass curves to deterministic and stochastic models. Statistical correlation and regression still provides useful means of analysing past behaviour and for the prediction of future consequence. These methods have several advantages over graphical analysis because of their ability to consider multi-variate problems and the availability of mathematical meaures for the reliability of estimates. Analysis of time series before and after the change in regime provides an understanding of the corresponding changes in the various statistical elements of the system. Parameter optimisation was recognized as an important and powerful tool to understand the cause-effect relations.

CHAPTER IV

DATA COLLECTION AND VALIDATION PROCEDURES

4.1. DESCRIPTION OF THE WATERSHED

The main physical characteristics of a watershed are its size, elevation, orientation, shape, slope, soil type, drainage, water storage capability, land use, and vegetation cover. The combination of these factors hydrologically classify the watershed. For example, hydrologically large watersheds are those in which storage effects dominate, making the response of rainfall to be sluggish, and insensitive to variation of rainfall intensity and land use. On the other hand, small watersheds are controlled by overland flow and land use, slopes etc, which have a strong influence on the magnitude of peak discharges. Therefore, the type and nature of a watershed influence rainfall conversion and contribution to streamflows.

Consequently a reconnaissance survey was conducted on the case study watershed to examine and describe its physiography, geology, land use, and water resources and initiate detailed study of streamflows and land use. To facilitate a rational comparison and assessment of hydrological parameters, the watershed was divided into three subwatersheds, 1, 2, and 3 (Figure 4.1). Subwatershed 1 (SWSI) is the upper forested area draining through the river gauging station (RGS), 2FC05, subwatershed 2 (SWSII) through 2FC09 and 3 (SWSIII) through 2FC16, situated at the river mouth in the shore of Lake Nakuru. The area of each subwatershed increases downstream, with that of SWSIII equal to the total watershed area. Historical hydrological and climatic data were collected and analysed after a thorough quality checks and control.

4.1.1. Physiography of the watershed

Enjoro river watershed lies in between longitudes 35°50'E and 36°21'E and between latitudes 0° 15'S and 0° 25'S in the Rift Valley Province of Kenya. The river drains into lake Nakuru through the peri-urban areas of Egerton University, Njoro Township and the urban areas around Lake Nakuru National Park. Its elevation at the top ridge is roughly 2698 m above sea level (a.s.l) falling to around 1722 m a.s.l (a drop of 1000m) in less than 60 km horizontal distance. The approximate watershed area is 229 km2. The most representative feature of the behaviour of the watershed system is its response function. Therefore the comparison of streamflow patterns from the subwatersheds is based on the physiographic parameters. This enables the identification of hydrological responses which could be attributed to factors other than land use.

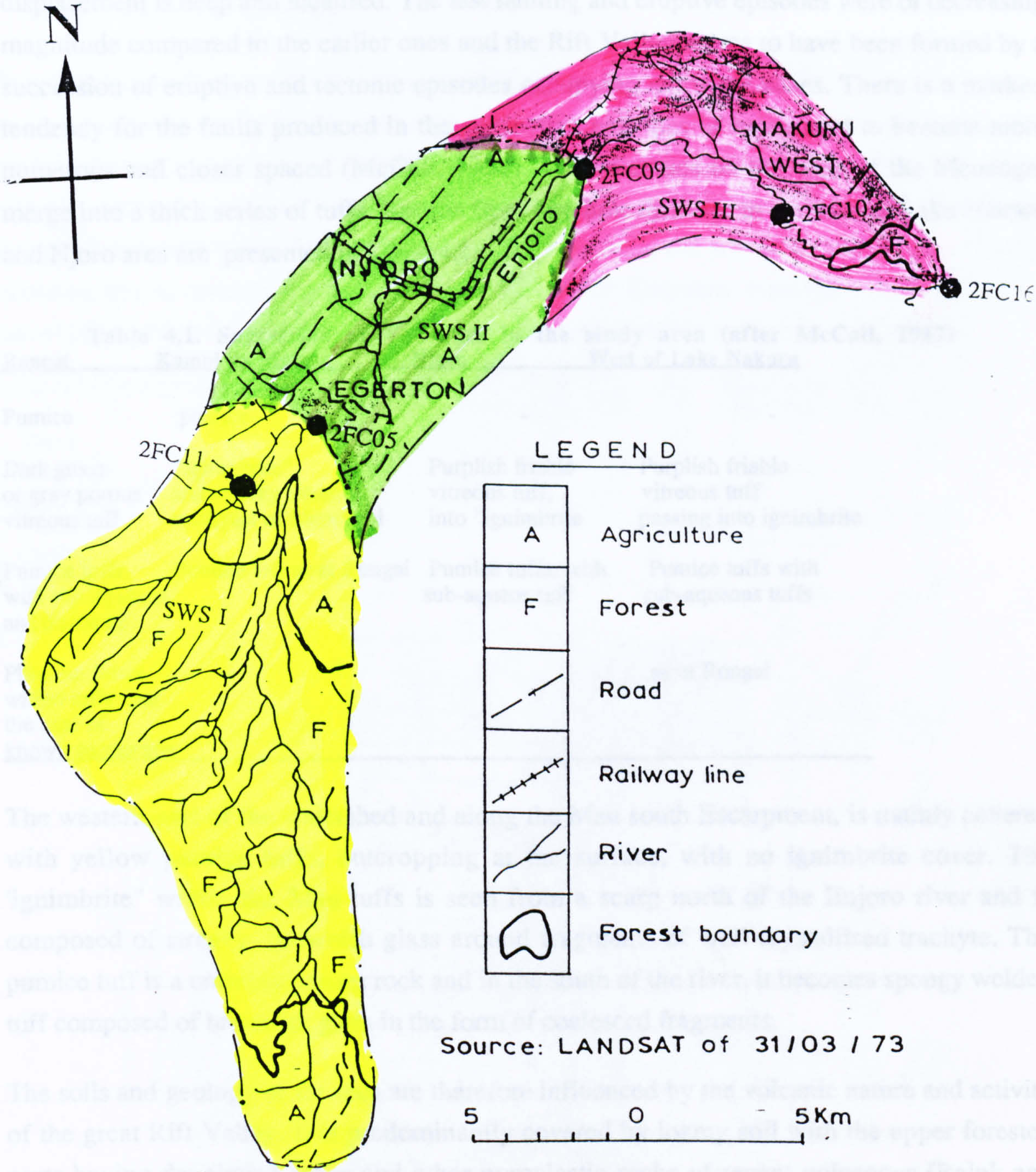


Figure 4.1. The location and boundaries of the Enjoro River watershed, showing streamflow gauging stations (schematic diagram).

The geological history of the Njoro area is extremely complex, due entirely to the tectonic and volcanic disturbances of the Rift Valley which dislocated the peneplained surfaces of the African shield, forming separated ridges and troughs, tending for the most part north-south and piling up great masses of volcanic rock on these structures (McCall, 1967).

The history of the Rift Valley is remarkably complex. Within the Lake Nakuru basin, the displacement is deep and localised. The last faulting and eruptive episodes were of decreasing magnitude compared to the earlier ones and the Rift Valley seems to have been formed by a succession of eruptive and tectonic episodes occupying narrower zones. There is a marked tendency for the faults produced in the successive episodes of movement to become more numerous and closer spaced (McCall, 1967). The volcanics on the west of the Menengai merge into a thick series of tuffs. Results from borehole drillings in the west of Lake Nakuru and Njoro area are presented in Table 4.1.

Table 4.1. Succession of Volcanics in the study area (after McCall, 1967)

Rongai	Kampi ya Moto	Njoro	West of Lake Nakuru
Pumice	pumice	-	-
Dark green or grey porous vitreous tuff	"Ignimbrite" and streaky vitreous trachytes of Menengai	Purplish friable vitreous tuff, into "ignimbrite	Purplish friable vitreous tuff passing into ignimbrite
Pumice tuffs, with sub-aqueous and diatomite	Pumice tuff as at Rongai	Pumice tuffs, with sub-aqueos tuff	Pumice tuffs with sub-aqueous tuffs
Phonlite flows within tuffs near the base of known succession			as in Rongai

The western part of the watershed and along the Mau south Escarpment, is mainly covered with yellow pumice tuffs, outcropping at the surface, with no ignimbrite cover. The 'ignimbrite" within the Mau tuffs is seen from a scarp north of the Enjoro river and is composed of streaky, brownish glass around fragments of well crystallised trachyte. The pumice tuff is a crumbly porous rock and in the south of the river, it becomes spongy welded tuff composed of brownish glass in the form of coalesced fragments.

The soils and geology of the area are therefore influenced by the volcanic nature and activity of the great Rift Valley. It is predominantly covered by loamy soil with the upper forested parts having developed ashes and other pyroclastic rocks of recent volcanoes (Ralph and Helmidt, 1984) and deep to deep well drained to moderately deep loamy sandy clays (vitric Andosols).

The lower reaches around the Ronda area, are generally covered with erosive lacustrine soils. The area is generally underlain by undifferentiated pyroclastic materials consisting of mainly poorly consolidated volcanic tuffs and ashes (Thompson and Dodson, 1963). Tuffs are widespread in the watershed and are frequently altered into clays. The geological formation in the respective reaches depict a major spatial variability which obviously influenced the hydrogeological and moisture storage of the watershed.

4.2. THE CLIMATE OF THE WATERSHED

4.2.1. Rainfall Regime

The climate is influenced by its wide range of relief and topography. Heavy thunderstorms occur mostly in the afternoons with very high erosive intensities. Long-term mean annual rainfall from a 54-year-record is 1200 mm at the upper reaches distributed trimodally with peaks in April, August and November. The lower reaches including lake Nakuru receives about 800 mm with temperatures and rainfall influenced by the watershed relief. The rainfall recording station network is shown in Figure 4.2 and the long-term seasonal values of rainfall data obtained from the Meteorological Department are presented in Figures 4.3 (a-g).

RAIN GAUGE STATIONS

1. Teret
2. Nessuiet
3. Egerton
4. NPBRS
5. Ogilgei
6. Tecfarm
7. Nakuru
8. Likia

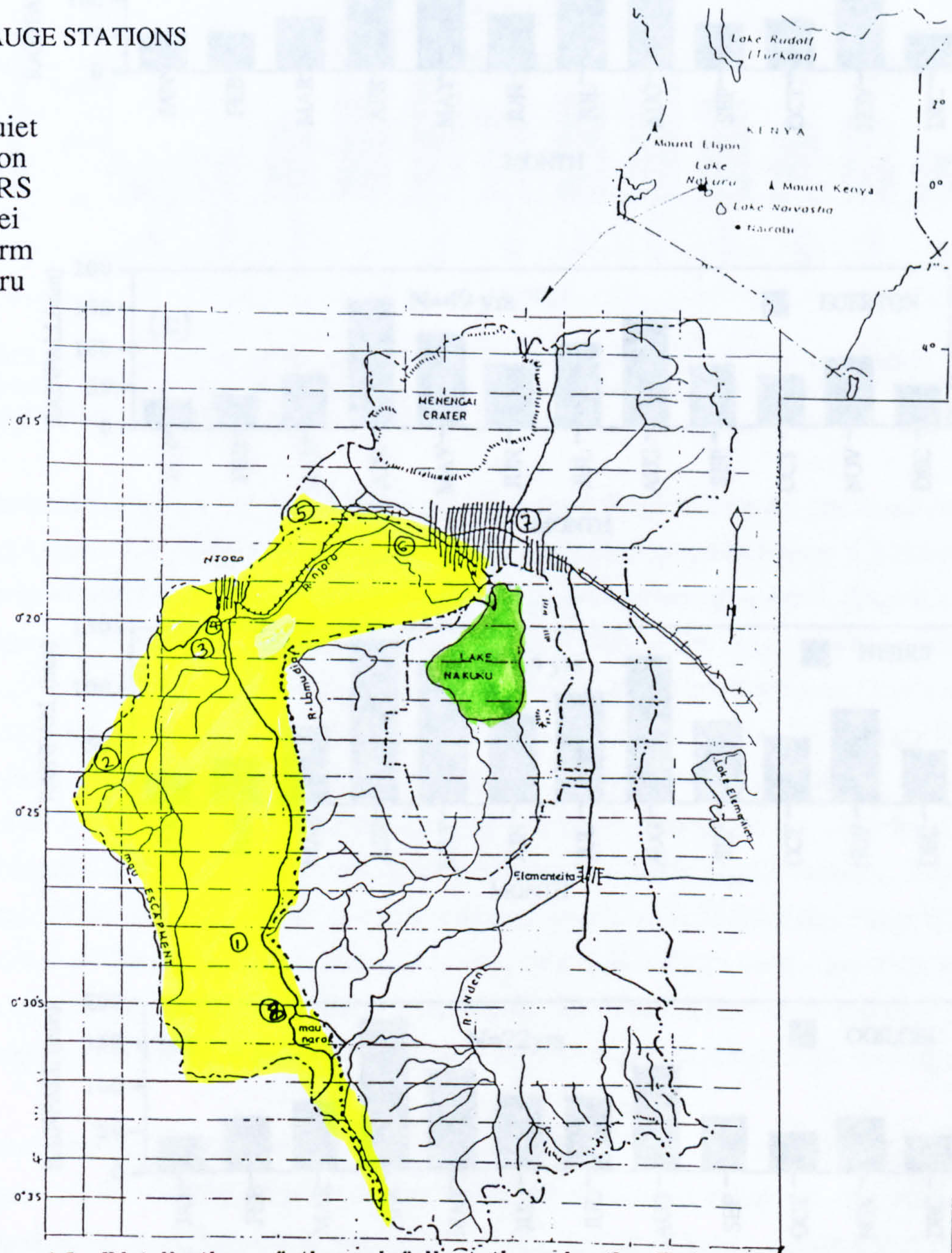
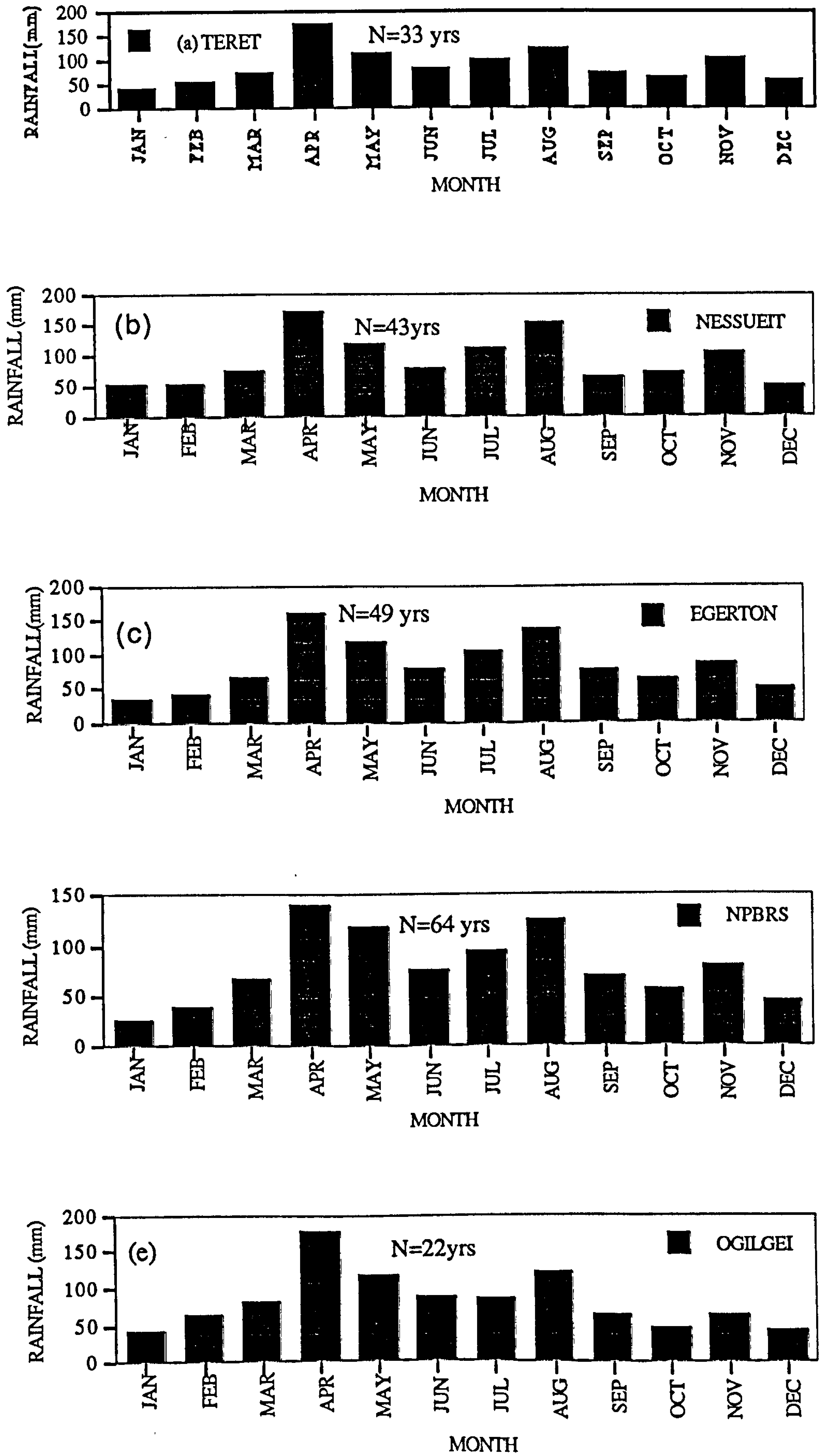
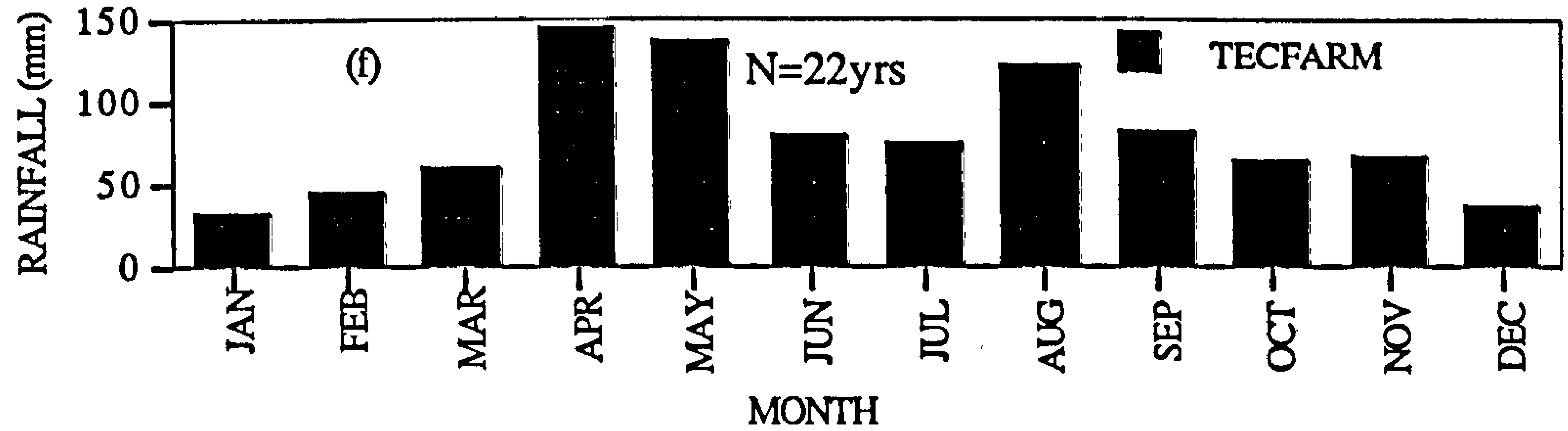


Figure 4.2. Distribution of the rainfall stations in the Enjoro river watershed in Kenya


RAINFALL (mm)
(a) TERET
N=33 yrs
(b)
N=43yrs
NESSUEIT
(c)
N=49 yrs
EGERTON
N=64 yrs
NPBRS
(e)
N=22yrs
OGILGEI
JAN FEB MAR APR MAY JUN JUL AUG SEP OCT NOV DEC
MONTH



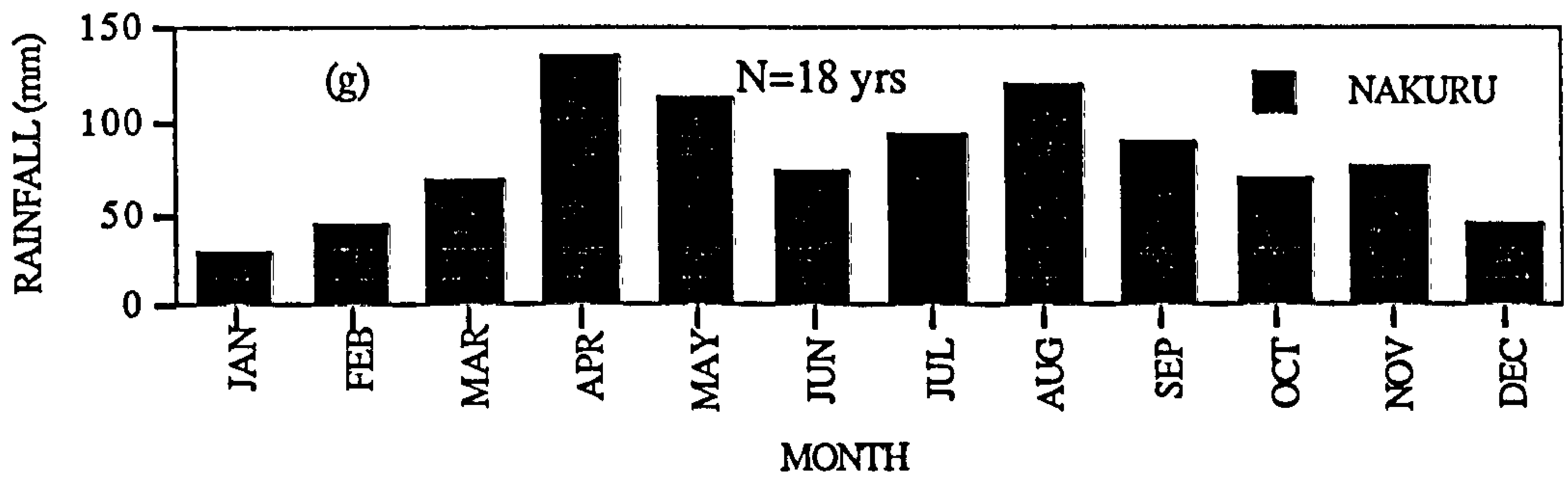


Figure 4.3(a-g)). Long-term monthly variation of rainfall at Teret, Nessuit, Egerton, NPBRS, Ogilgei, TecFarm, and Nakuru rainfall recording stations(Source: Meteorological Dept., Nairobi)

The statistical descriptors of the rainfall from these recording stations are presented in Table 5.1. An examination of the historical patterns of the rainfall given in Figure 4.3 show long-term mean monthly values to exhibit three distinct peaks occurring in April, August and to a small extent in November. This distribution is more evident in the upper elevation rainfall stations of Teret, Egerton, NPBRS and Nessuiet. As the elevation decreases (Table 4.5), the rainfall amounts and peaks differ. For example, the Ogilgei station and Technology farm (Techfarm) which are about 4 km apart exhibit the main peak in April, although of unequal magnitudes (a maximum of 190 for Ogilgei and 148 for Tecfarm). On the other hand, Tecfarm's peak of 130 mm occur in August, compared to 120 in Ogilgei station. The November peaks are more or less of the same magnitudes, with Technology Farm being a little bit higher. The further south east and lowest elevation Nakuru station have the same trimodal pattern, similar to the stations in the higher elevations. The rainfall regime commences in the mid of March and continues to November, although with varying amounts.

The wettest months occur between April and September and the driest months, between December and March. The variations observed amongst stations values are caused mainly by the watershed relief, and probably the land use in the area may have a minor effect. As stated earlier, the elevation of the entire watershed drops from roughly 2700 m a.s.l to 1722 ma.s.l in only a horizontal distance less than 60 km. This drastic drop may have contributed to the observed extreme rainfall variation. Secondly, the land use changed from mainly agriculture and forestry in the higher elevation, traversing the peri-urban areas and into the high density urban developments in Nakuru Municipality. Whether the land use change from forestry to higher percentage of urban area caused changes in rainfall amounts and its characteristics is still a debatable issue. This study cannot conclusively attribute any of the observed rainfall patterns to changes to urbanisation because the changes are repetitive in other stations around the area (section 3.2.4)

The seasonal variations of air temperature average (used as an indicator of any climatic change) from a 1960-1990 record is assessed by plotting the long-term mean, maximum and minimum temperature values at Egerton, NPBRS and Nakuru stations in Figure 4.4 and 5(a&b). The plots depict a pattern of higher temperatures from December to March which correspond to the driest periods of the year. The coldest months fall between June and September, where the minimum temperatures is about 7 °C in Egerton, increasing downstream to 16 °C at NPBRS and around 18 °C at Nakuru station. Again, air temperatures are influenced by the elevation of the stations and the minimum values occurring during the periods of higher rainfall. The seasonal air temperature regime is the same as that of the rainfall pattern, and the highest evaporation occur at periods of highest radiation (Kimani et. al., 1992)

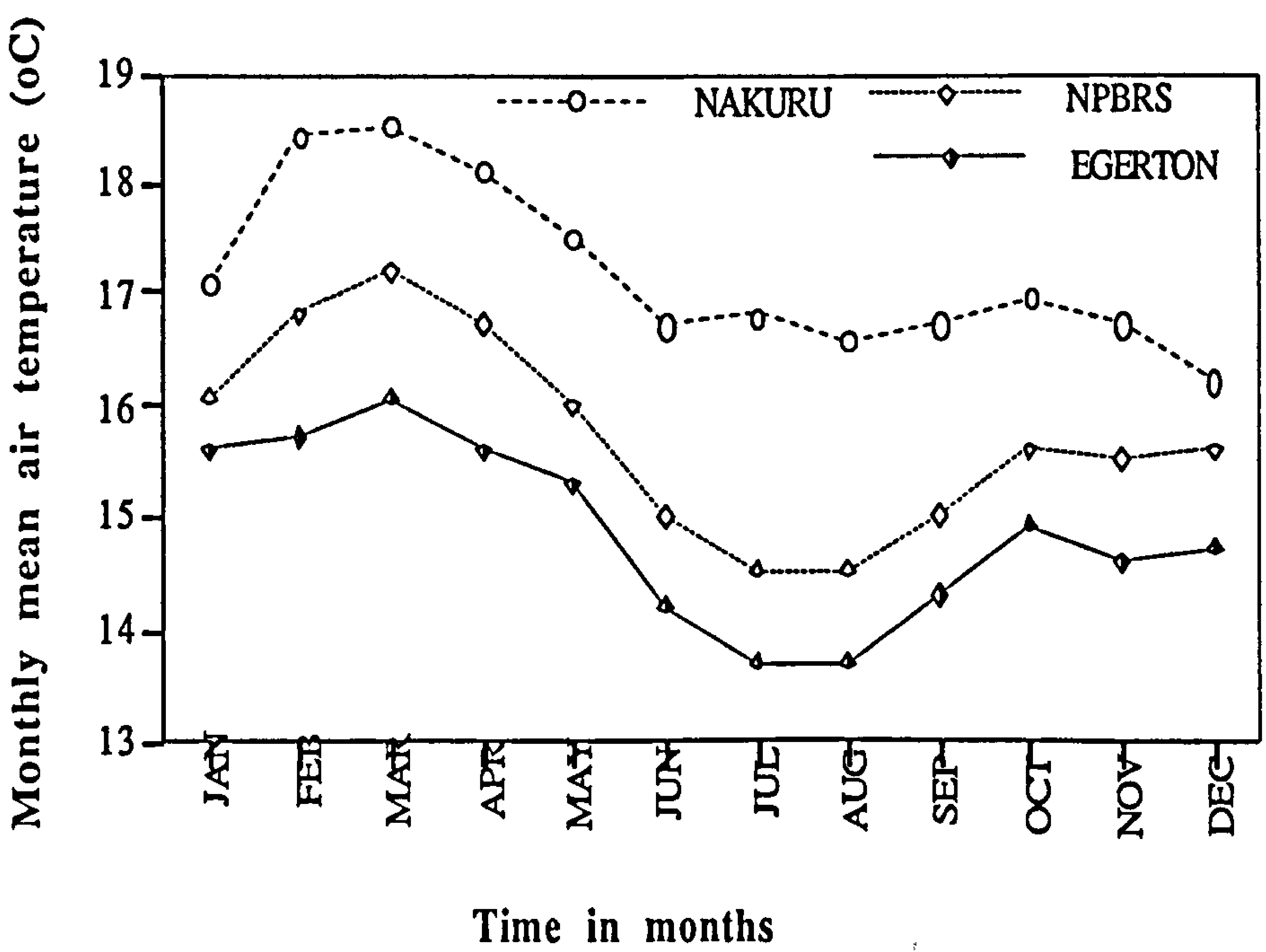


Fig4.4. Long-term mean monthly air temperature (oC) at selected weather stations

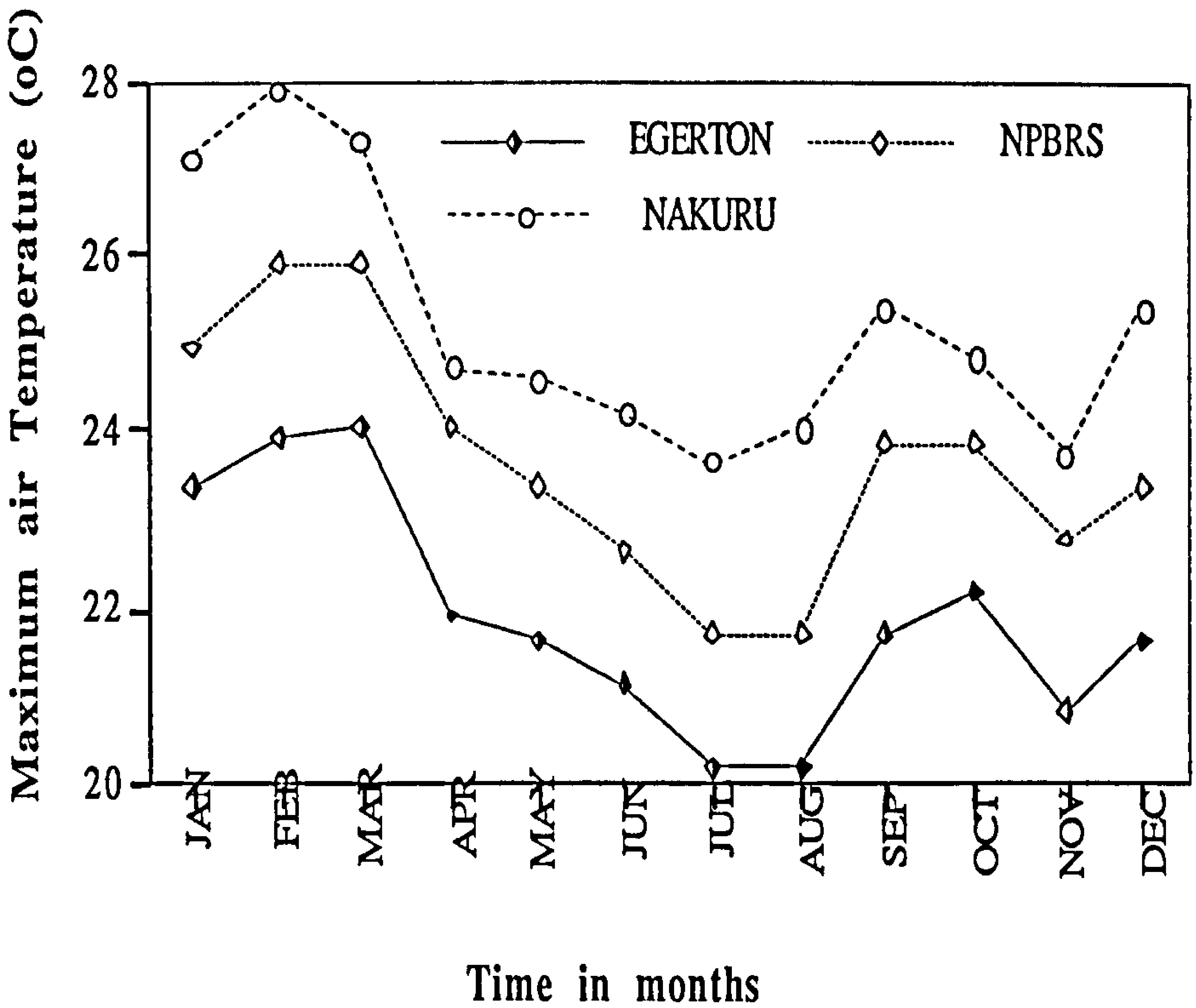
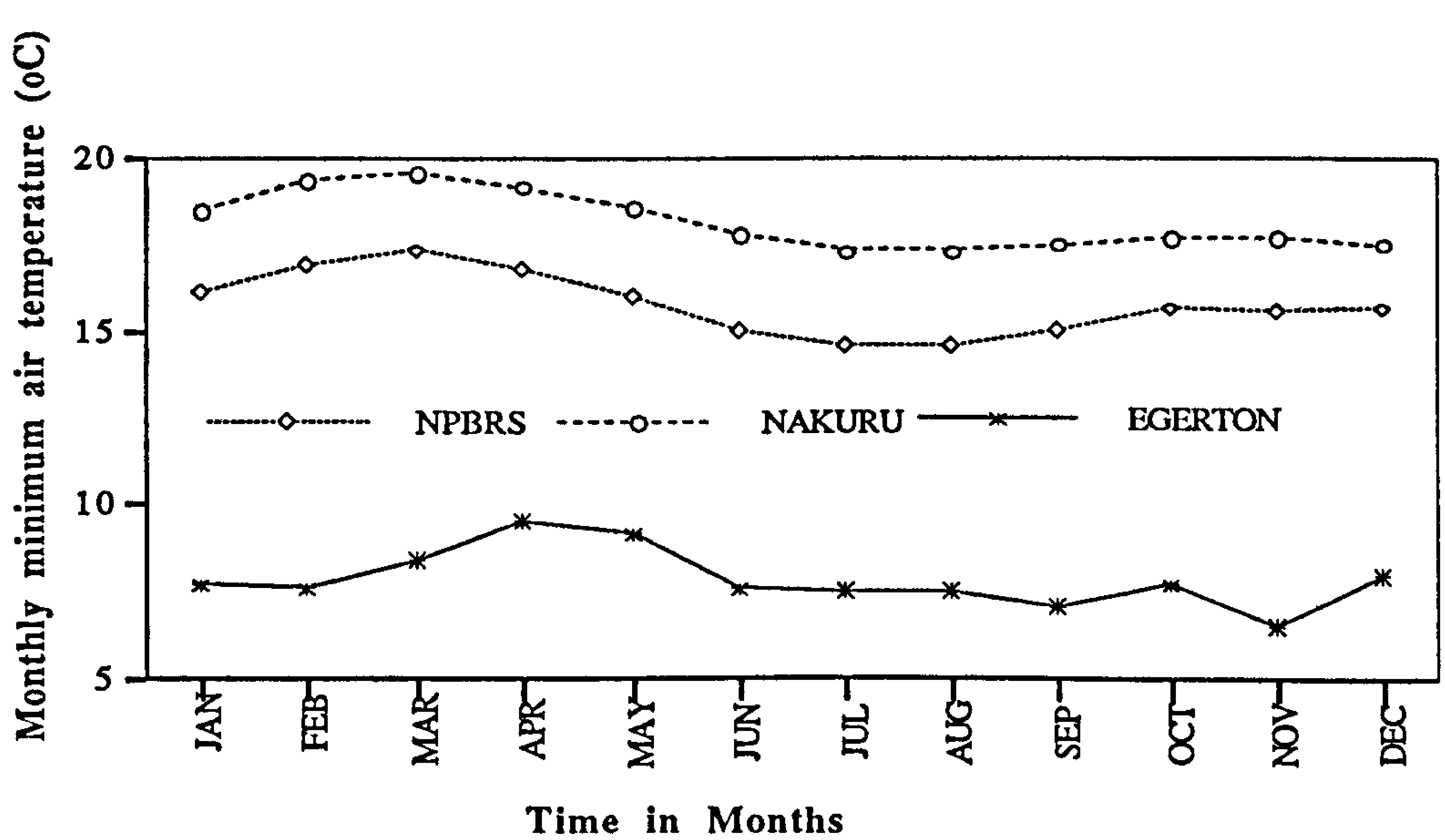


Figure 4.5a. Long-term mean monthly maximum air temperature in Enjoro river watershed.



(Source: Meteorological Department, Nairobi).

Figure 4.5b. Long-term minimum monthly air temperature (oC) at Enjoro river watershed

4.3. NETWORK OF METEOROLOGICAL STATIONS

4.3.1. World Meteorological Organisation Registered Stations.

Three World Meteorological Organisation (WMO) registered weather stations exist in the watershed. Data from the stations on evaporation, rainfall, radiation, temperature, wind, humidity were used after quality control checks. The stations are located at Egerton University, Njoro Plant Breeding Research Station (NPBRS), and Nakuru Meteorological Station. All these however did not have complete and continuous data. Two are located at the central part of the watershed and one at the south east end of the watershed. The locations and history of the stations are presented in Table 4.2.

Table 4.2. World Meteorological Organisation Stations in the Watershed.

Name of Station	Station No.	Latitude	Longitude	Elevation	Yrs of record
Nakuru Met. Station	09036261	00° 16' S	36° 04' E	1880m	26 yrs
Njoro Plant Breeding	09035021	00° 20' S	35° 57' E	2160m	18 yrs
Egerton University	09035092	00° 23' S	35° 55' E	2238m	27 yrs

4.3.2. Rainfall Recording Stations

There are six (5) standard rain gauge stations with an average record of 30 years. These are distributed rather uniformly within the watershed. The stations are the Technology Farm station, Ogilgei Farm station, Likia Forest Station, Teret, and Nessuit Forest Stations. With this number of stations within a catchment area of roughly 229 km^2, the quantity of data collected were enough to describe the climatic actuality of the watershed. The historical information of the rain gauges is given in Table 4.3. Data collection and data quality control by the Meteorological Department followed recommended procedures (WMO, 1974a, 1975b and 1976).

Table 4.3. Rainfall Recording Stations in Enjoro River Watershed.

Rain gauge Station	Station Number	Elevation (m.a.s.l)	Years of Record
Teret Forest	09035233	2435m	33
Nessuit Forest	09035119	2420m	43
Technology Farm	0903676	2015m*	22
Ogilgei Farm	09035000	2163m	22
Likia Forest	09035232	2540m	32
Nakuru Met.	09036261	1880 m	18
Njoro NPBRS	09035021	2138 m	64
Egerton University	09035092	2238 m	49

* approximate elevation

4.4. RIVER GAUGING STATIONS

4.4.1. Historical Patterns

In general, the floor around Nakuru is characterised with a poor runoff regime, mainly due to numerous pumiceous formations which mantle the older rock surfaces (McCall, 1957). The river system drains into Lake Nakuru from the Mau Escarpment and loses much of its flow in porous or fissured zones. This influent nature of the river means less surface water reaches Lake Nakuru especially during the dry seasons. The bulk of the flow is assumed to accrue in water tables around the lake.

There are four operational staff river gauging stations (RGS) in the watershed, 2FC11 at the Little Shuru tributary, 2FC05, 2FC09 and 2FC16 along the main Enjoro river as shown in Figure 4.1. The stations are operated and maintained by the Kenyas' Ministry of Water Development. Historical evidence and field checks revealed that the design and calibration of the stations followed the International Standardisation Organisation procedures (ISO:1100, 1975). The streamflow data from these stations however, are inconsistent, with gaps in the daily and monthly values. Re-calibration of the rating curves was carried out for this study to establish any changes and undertake quality checks on the uncertainties of the stations. The establishment history and data collection in the stations is presented in Table 4.4.

Table 4.4. Distribution of river gauging stations in Enjoro river watershed.

Gauging Site	RGS	Record	Latitude	Longitude	Area (km^2)
Little Shuru Tributary	2FC11	1959-1993	0.3694°S	35.9027° E	44
Enjoro river at Egerton	2FC05	1941-1993	0.3550°S	35.9236° E	125
Enjoro river at Kirobon	2FC09	1959-1993	0.3167°S	35.9500° E	168
Enjoro river at Ronda	2FC10	1954-1975	0.3083°S	36.0011° E	
Enjoro river at L.Shore	2FC16	1990-1993	0.3172°S	36.0796° E	229*

* approximate area

The distribution of these gauging stations start from the upper reaches of the watershed to the lower reaches at the river's entry into lake Nakuru. Discharge data in RGS station 2FC10 was excluded in the analysis because, river Enjoro became influent around this location resulting in zero gauge discharge records for a long time. The influent nature of the river was attributed to the unstable geological formation on the floor of the Rift Valley and enhanced by an intensive human harvesting of sand in the river bed. Recording on the station was subsequently abandoned in 1975. Records from RGS 2FC16 established in 1989 was equally omitted in the analysis because of its short and inconsistent data.

4.5. DATA COLLECTION AND VALIDATION

The methods of moving averages were used to establish an existence of trend (Davis, 1964). Sudden changes in the response functions or inconsistencies in the data were identified using mass plots and double mass curve analyses (WMO, 1974 and UNESCO/WMO, 1977, WMO, 1980). The apparent non homogeneity of the data were substantiated with ground-truth transect survey and folklore historical evidence during the 1992 watershed survey.

Further, human-induced changes were quantitatively estimated using graphical illustrations, models and residual analysis. The use of statistical correlation, regression and the analysis of time series before and after the perceived changes were emphasised. The data were then checked for temporal and spatial homogeneity (see section 4.8). The temporal homogeneity analyses were used to detect trends of sudden changes in the data. The variables which exhibited decreasing or increasing trends with time because of the long cycles, climatic, land use or other changes were examined in more detail.

The spatial homogeneity of the data were checked to determine if the observed variation in the parameters was significant and can be attributed to differing physiographic and climatic features or human activities in the watershed. The spatial homogeneity procedures developed by Yevjevich (1969a) were applied to the meteorological data. In general, the data analysis format depended upon the availability, nature and quality of the data. The detailed analysis procedure is presented along with the results of each analysis and discussion in Chapters Five and Six. In addition, graphical procedures were preferred in the entire study because : (1) of their ability to introduce subjective judgements which are important when dealing with hydrological data; (2) the background knowledge on data accuracy and for regional coherence is easier to introduce information with graphical procedures; (3) the commitment to a particular distribution is reduced and (4) graphical procedures are helpful to fit annual runoffs and also simultaneously the values of other durations such as the three and six month durations.

4.6. LAND USE CHARACTERISTICS IN THE WATERSHED

The primary objective of the study was to examine and describe the influence of the perceived changed land use on the river's hydrologic regime. To achieve this, it was necessary to determine the extent of land use change that occurred between 1960 and 1990. Consequently a demographic and physiographic survey of the area was undertaken. An examination of the agricultural, urban and forestry development was undertaken as described in the following subsections using various methods that range from ground-truth transect surveys, rapid rural appraisal, aerial and topographic maps to the analysis of remote sensing imageries and vegetation indices.

4.6.1. Historical Land Tenure System

Prior to early 1960, the land within Enjoro river watershed area was mainly under a large scale conservative (traditional) agriculture and forestry. However, in the mid 1960s, data obtained from the Ministry of Agriculture (GOK, 1986) show the farms were purchased by land buying companies. They were then subdivided into smaller units of an average of 50 ha and housing individual families of between 4 and 6 persons per household.

The plots were further subdivided into smaller units of 0.2 to 5 ha. by the late 1980s (Karanja et al., 1986, and Kimani et al.,1992). In 1990s, these units were too small for agricultural purposes. There were converted into peri-urban high density housing units to meet the higher demands of increasing urbanisation in the area. Ultimately, the land tenure system changed from a conservative large sale agriculture and forestry in the 1960s to urbanisation and intensive small scale agriculture in the 1990s. This uncontrolled fragmentation placed heavy demands on the watershed by upsetting its ecological and hydrological balance. Studies on the Njoro river between 1975 and 1985 reported in Karanja et al. (1986) reveal the effects of the changes included increased sedimentation and limited amounts of water supply to meet the local domestic demands.

Ground-truth transect survey conducted for this study in 1992 and 1993 confirmed the previous findings of increased land subdivision and deforestation which changed the land tenure system from large-scale agriculture in the 1960s to that of small-scale and urbanisation in the late 1980s and 1990s. The family size per holding increased from 6 to 8 during the same period. Human population increased tenfold from 25 persons per km^2 to 250 persons per km^2 on the average.

This change in land holding size and subsequently land use is reflected in the entire Nakuru district. The distribution of land by holding size (large to medium scale) for the period 1975 to 1979 is presented in Table 4.5 for example, which emphasises the fact that the land holding size response to increased human population was on the decreasing trend. Post-impact assessment of the demographic, social and environmental changes are analysed and discussed in Chapters V, and VI. From the table however, small size holdings of less than 20 ha were unavailable because of a rapid nature of land subdivision and a dynamic change of ownership.

Table 4.5. Distribution of land holding size from large to medium scale farms in Nakuru district from 1975 to 1979 (Holdings in each size group)

Group Size in ha	1975	1976	1977	1978	1979	1975/79 change
less 19	84	84	84	87	90	+7.14
20 - 49	68	79	82	83	111	+63.24
50 - 99	49	51	53	56	58	+18.37
100- 199	95	92	93	92	119	+25.26
200- 299	75	76	80	76	72	-4.00
300 - 399	63	66	62	64	67	+6.35
400 - 499	46	47	46	44	47	+2.17
500 - 999	122	121	123	127	126	+3.29
1000 - 1999	46	46	45	47	33	-28.28
2000 - 3999	28	27	27	27	26	-7.14
4000 - 19999	13	14	13	13	15	+15.38
Greater 20000	1	1	1	0	0	-100.00

Source: Central Bureau of Statistics, Agricultural census of large-scale farms, Nairobi, Kenya.

The notable changes in land holdings over 5 years is outstanding in the category of farms between 1000 and 2000 ha which reduced by 28%, resulting in a 25% increase in the 100-199 category and more dramatic 63% increase in the 20-49 ha category size. A closer examination of the data show a decreasing trend in land sizes in the district. This trend is more pronounced at divisional and locational levels such as the Enjoro river watershed. There was therefore an increase in land subdivision into smaller units. These units are susceptible to intensive cultivation, denudation and degradation, with an adverse effect on the watershed moisture retention and streamflow recharge.

4.6.2. Estimating Area Under Different Land Uses

Land use changes in the watershed were obtained from a study of the aerial, topographic maps, and satellite imageries. In addition, data were obtained from the Department of Resource Surveys and Remote Sensing (DRSDRS) of the Ministry of Planning and National Development (GOK, 1979). The aim was to estimate the historical changes of different land use areas in terms of land fragmentation, human density, forested areas, urbanisation, and agricultural development.

Land use maps at different periods between 1970 and 1990 were delineated and respective areas computed as percentages of the total watershed area. The maps were digitised in order to derive the area associated with each land use. Detailed historical land use area extraction procedures from the aerial and satellite imageries is provided in Epp et al. (1983), Agabia (1985), and Ottichillis (1985). Specific and relative land use areas for 1970, 1973, 1979 and 1987 were planimetered and analysed to establish the amount of land that changed from agriculture and forestry to urbanisation and other impervious grounds. Changes in land use areas were assumed to have a direct implication both on the hydrology of the area and particularly the rivers' flow regime.

During the field survey, the catchment areas of two chosen subwatersheds in the river basin were delineated for detail analysis. The subwatershed were distinct and intertwined, because the upper subwatershed I historically has been under forest cover and subwatershed II under agriculture and of recent times urban development. Their estimated areas are 125 km^2 and 168 km^2 respectively (Table 4.6). Satellite imagery estimated percentages of different land use areas in the subwatersheds for the period between 1970 and 1990 are given in Tables 4.6 and 4.7. For SWSI, land area under agriculture increased by about 38%, urban increased by 11% and forestry decreased by about 59%.

Table 4.6. The percentage (%) of different land use area between 1970 and 1990 in subwatershed I draining through RGS 2FC05 (125 km^2)

Year	Agriculture Area Km^2	(%)	Urban Area Km^2	(%)	Forestry Area Km^2	(%)	Other Areas* Km^2	(%)
1970	25.00	21.00	1.87	1.50	97.50	78.00	0.62	0.50
1973	43.25	34.50	10.50	8.40	71.25	57.00	0.12	0.10
1979	65.62	52.50	14.25	11.40	45.00	36.00	0.12	0.10
1987	73.13	58.50	14.08	11.90	23.00	18.50	0.12	0.10
%change	48.13	37.50	10.68	+10.40	74.50	-59.50	0.50	-0.40

*others includes bodies of water, paths and open lands

Table 4.7 presents the different percentages of land use areas in SWSII which also includes SWSI. The data shows that there has been a tremendous change in land use in less than three decades. There was a 30% increase in agricultural land, about 45% decrease in forested area and a 20% increase in urban area. These changes are assumed to have contributed to water shortages in the area and therefore an analysis of change in the river's hydrologic regime was necessary. This rapid change in land use is depicted more evidently in Figure 4.6 for subwatershed I.

Table 4.7. The percentage of different land use area between 1970 and 1990 in SWSII (168 Km^2)

Year	Agriculture area (km^2)	(%)	Urban Area (Km^2)	(%)	Forestry Area (Km^2)	(%)	Other Areas* (Km^2)	(%)
1970	58.80	35.00	8.40	5.00	100.13	59.60	0.67	0.40
1973	67.20	40.00	26.71	15.90	73.92	44.00	0.17	0.10
1979	82.32	49.00	36.96	22.00	48.55	28.90	0.17	0.10
1987	108.36	64.50	42.00	25.00	24.19	14.40	0.17	0.10
Change	49.56	29.50	33.60	20.00	-75.94	-45.20	-0.50	-0.30

4.6.3. The Use of Normalized Difference Vegetation Indices

The application of remote sensing techniques is an attractive alternative to traditional survey methods because of its lower labour requirements. With minor modifications the techniques described by Epp et. al., (1983) were used to analyse images from successive satellite flights. The techniques are however, very expensive and can only be utilised in some priority projects in Kenya.

Vegetation cover is associated with pervious surfaces, therefore mapping the vegetated areas should result in an approximate division of a watershed into pervious and impervious areas. Green vegetation is distinguished by strong reflection in the near infra-red (IR) and strong absorption in the visible wavelengths of the electromagnetic spectrum. Other surfaces such as soils, do not have as much difference. This characteristic is then used to estimate the change in vegetation cover in a watershed.

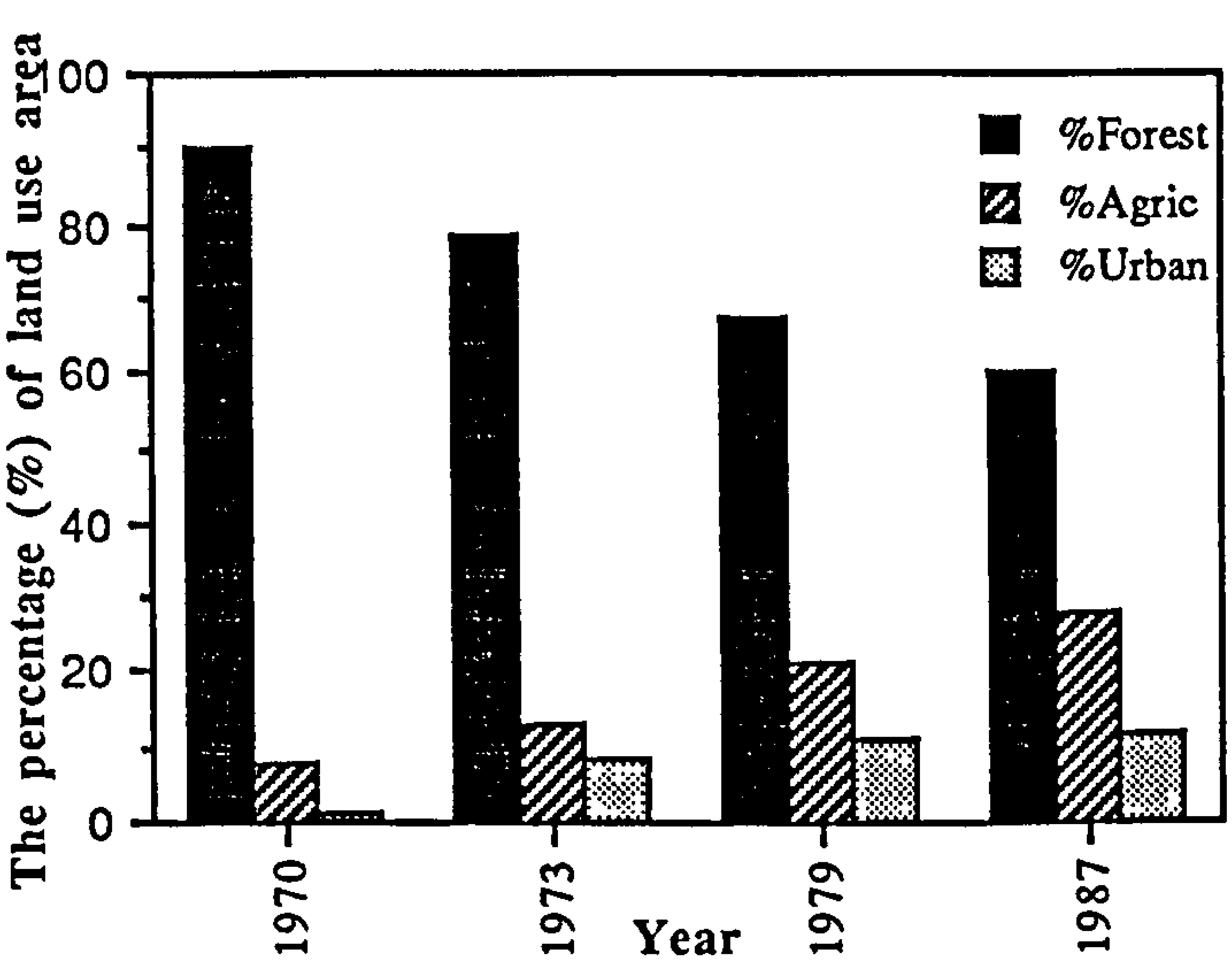


Figure 4.6. The percentage (%) of different land use areas in SWSII (1970-1990).

The normalized difference vegetation index (NDVI) discussed in Chapter III is suitable for global and local monitoring of vegetation cover because it partially compensates for changing illumination conditions, surface slope and viewing angles. The NDVI data obtained from the Regional Centre for Remote Sensing, Surveys, and Mapping (RCRSM) in Nairobi, Kenya was thus abstracted from the global storage systems for locations defining the watershed boundaries and parameter measuring sites Three-day data were averaged to monthly values (Appendices D.3) for the chosen sites from 1982 to 1990. The 1989 values were unavailable because data for that year were destroyed by volcanic eruptions in the Philippines which blurred the entire outer space.

Since the NDVI data indicate both the temporal and spatial changes in vegetation cover, they were utilised to provide comparison of historical trends in land use in the watershed. In this way, the indices could be used to support inferences and evidence of any apparent changes in streamflow pattern, afforestation/deforestation, urbanisation and any expansive agricultural development in the study area. The specific sites in the study area in which the indices were abstracted are given in Table 4.8.

4.7. DEMOGRAPHIC TRENDS IN THE WATERSHED

The demographic picture of the Njoro division in Nakuru district is characterised with an increasing trend of rural-urban out-migration of human population. The population growth trends in the division relative to the entire district is presented in Table 4.9. and the rural-urban out-migration in Table 4.10 respectively and perhaps Figure 4.7 would depict the clearly the observed demographic changes.

Table 4.8. Normalized Difference Vegetation Indices (NDVI) abstraction sites in the Enjoro River Watershed for the period 1981 to 1990 (MSS*).

NDVI Data Site	RGS	Latitude	Longitude	Resolution(Km)
Little Shuru	2FC11	0.3694° S	35.9027o E	7.6x7.6
Enjoro Egerton	2FC05	0.3750° S	35.9236o E	7.6x7.6
Egerton Univ	Met.stn	0.3667° S	35.0153o E	7.6x7.6
NPBRS, Njoro	Met.stn	0.3333° S	35.9333o E	"
Central Part	Middle	0.3333° S	36.0014o E	50x50
Kirobon Farm	2FC09	0.3167° S	35.9500o E	7.6x7.6
Ronda Estate	2FC10	0.3083° S	36.0011o E	"
Lake Shore	2FC16	0.3172° S	36.0796o E	"
Nakuru Met.	Met.stn	0.2667°S	36.0011o E	"

*MSS = Multispectral Scanner and Middle= mid of watershed

Table 4.9. Demographic trends of Njoro division in Nakuru District.

Population Area	1969	1979	1988	1989	1990	1991	1993	Annual(%)
Kenya (10^6)	10.9	16.1	23.51	24.3	27.21			3.56
Nakuru District(10^6)		0.52	0.85	0.89	0.93	0.99	1.05	5.10
Njoro Division (10^6)		0.082	0.133	0.14	0.147	0.154	0.169	5.28
Njoro Urban (10^3)		6.10	16.49	24.41	26.12	27.95	32.00	12.55
Njoro Urban/Division (%)		7.63	12.68	17.44	17.41	17.47	18.82	2.38
Njoro Division Density		249	404	423	445	466	512	5.28

Source: Central Bureau of Statistics, Nairobi, Kenya.

A tremendous explosion of human population into the Enjoro urban area is presented in Tables 4.9 and 4.10 respectively. The average annual growth rate of urbanisation in the division was around 12.55%. The total population growth in the division which covers the whole of the Enjoro river watershed was 5.28, a little higher than the Nakuru district annual rate of 5.10. The urban area exploded between 1988 and 1989, with an increase from 16490 to 24410; a 48% rapid influx of human population.

Further investigation using a rapid participatory rural appraisal approach (PRA), revealed that several people moved and registered in the Njoro urban areas during the 1987/88 period possibly as a result of the 1988 general election. The data seems to indicate that most of immigrants remained in the area. In the 1980s and around the same period, explosive growth of industries, and educational and research institutions was observed. Coupled with a 6.94% annual rate of rural-urban migration into the area, the ratio of urban to total divisional population rose from 7.41% in 1979 to 18.88% in 1993 which is equivalent to average land use conversion rate from rural to urban of roughly 2.38% per year (i.e =12.55/5.28)

Table 4.10. Rural-urban out migration in Nakuru District.

Receiving Urban Area	1969	1979	1989	% Annual Rate
Nakuru Municipality	47151	92851	182652	7.00
Naivasha Municipality	6920	11491	22605	6.10
Njoro Urban Areas	3037	5803	11415	6.84
Nakuru District Urban	21984	38941	69737	5.94

Source : Central Bureau of Statistics, Nairobi, Kenya.

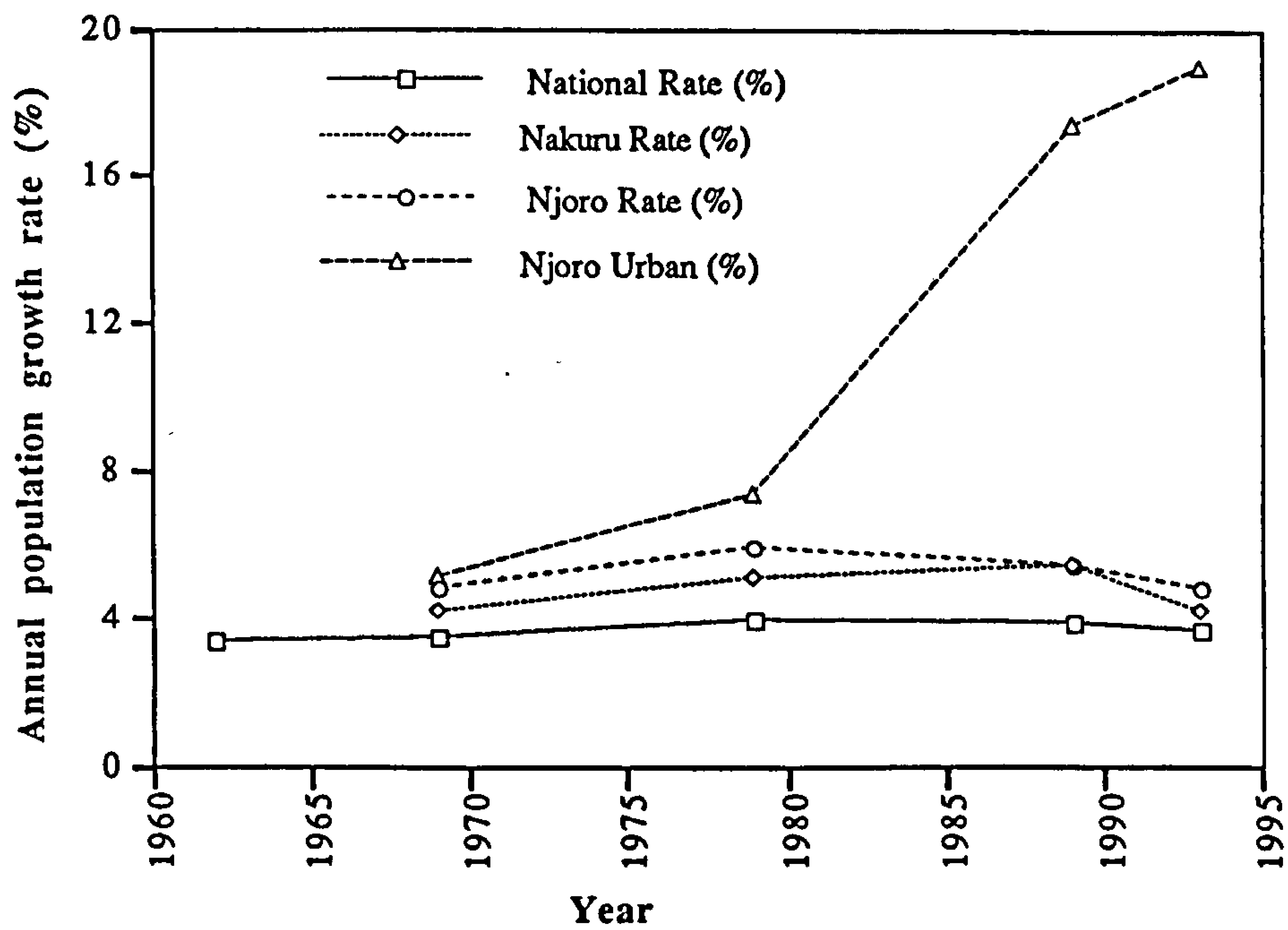


Figure 4.7. The demographic trends in the Njoro division of Nakuru district (Central Bureau of Statistics, Nairobi, Kenya, 1991).

4.8. DATA QUALITY CONTROL

4.8.1. Homogeneity Tests

In order to identify the temporal changes in hydrological responses due to land use change, the data must be cleared of any errors to be sure of the causal mechanism. The data were therefore checked for temporal and spatial homogeneity. Subjective and statistically based objective methods were used. The hydrometeorological data were collected, processed and stored by different people and institutions hence errors could have been introduced. Consequently, any inconsistencies were identified before the analysis of the data. The best technique for testing homogeneity in time series is the double mass curves , which is considered together with the difference in the actual time of occurrence of selected events. Time consistency is important since certain temporal parameters were considered for comparison.

The double mass analysis technique is well known and commonly used to test rainfall and streamflow records for nonhomogeneity (Schultz, 1976, Linsley et al., 1982). This can identify periods of changing trends in which case the cause is sought. It has also been used to identify the effects of land use on hydrological regimes (Braune and Wessels, 1980). It involves plotting the accumulating totals of one time series against one (or more) stations or series. The data series being tested should, if possible, be compared to at least 4-5 base station records of either rainfall or streamflow. One of the series is assumed to be homogeneous, and if the plot is an acceptable straight line, the other series is also assumed to be homogeneous. A change in slope and an identified clear break in the data series, means that a search for an explanation should be undertaken. The changes are confirmed by applying statistical tests of the means and variances at different time periods (partial series) of the data and examining the rate of change of their slopes.

The rainfall time series homogeneity tests were carried out by considering mean values from the upper (higher elevation) and lower reaches of the entire watershed as homogeneous and independent from each other. This assumption was developed from the examination of the watershed relief and topography which is a very important factor in rainfall distribution and streamflow generation. Therefore, the accumulated totals of annual mean rainfall series from the higher elevation stations were plotted against the corresponding series from the rainfall recording stations in that lower elevation part of the watershed. The results are presented in Figures 4.8 for rainfall stations in the upper reaches of the watershed. The slopes of the double mass curves for these stations are given in Table 4.11.

Table 4.11. The slopes of double mass curves for three selected periods for the upper watershed rainfall against the adjacent rainfall recording stations.

RAINFALL STATION	SLOPE	CORRELATION
NPBRS	+ 1.9941	$R^2 = 0.999$
EGERTON	+ 1.0411	$R^2 = 1.000$
NESSUEIT	+1.1455	$R^2 =0.999$
TERET	+1.0940	$R^2 = 0.999$

Similarly, the lower watershed accumulating mean annual rainfall was plotted against the corresponding accumulations of annual mean rainfalls from Ogilgei, Technology Farm and Nakuru rainfall recording stations in the lower elevation reaches, with the plots presented in Figures 4.9.

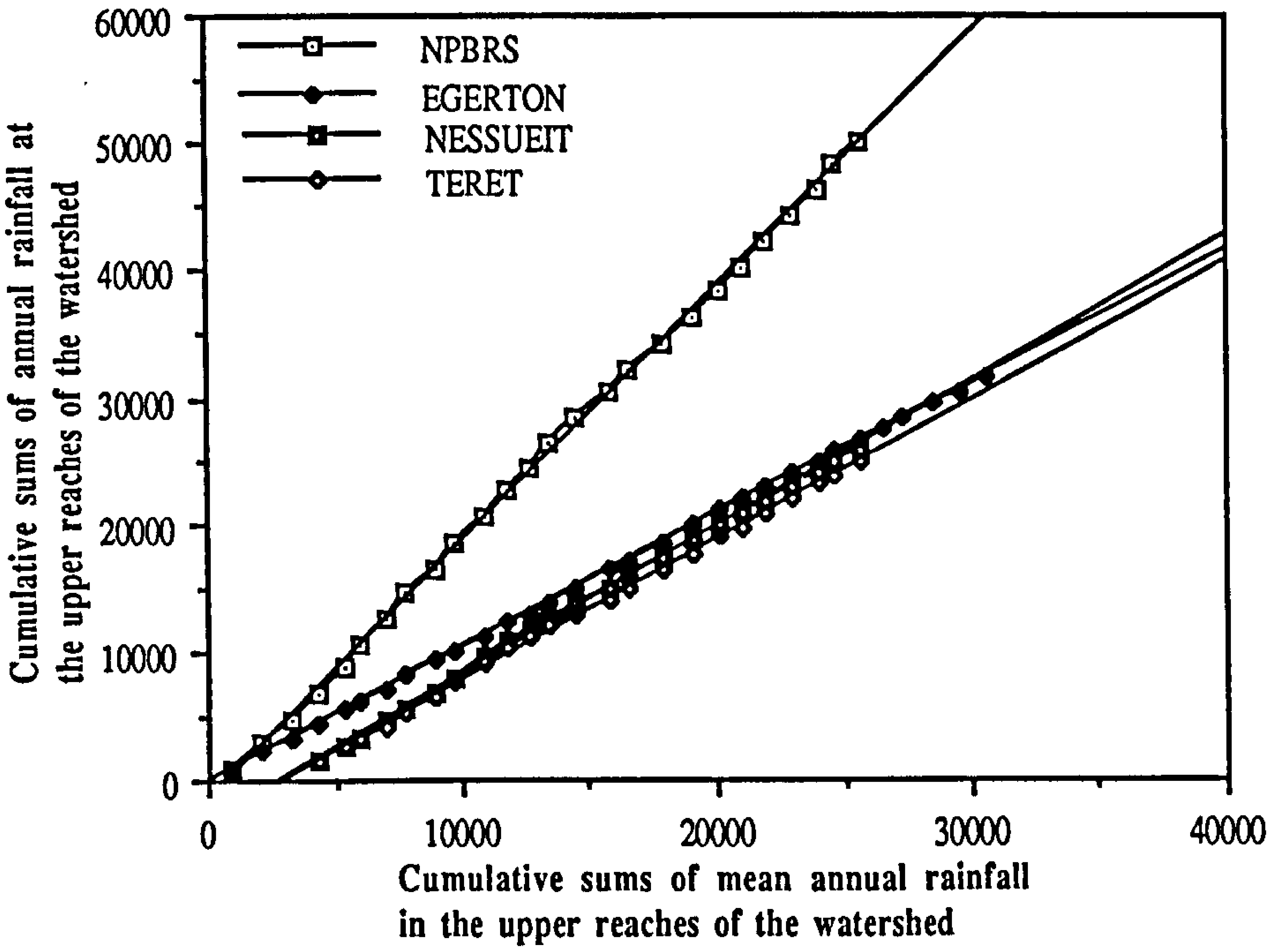


Figure 4.8 Double mass curve of accumulated mean annual rainfall for the upper Enjoro river watershed and NPBRS, Egerton, Nessueit and Teret rainfall stations respectively.

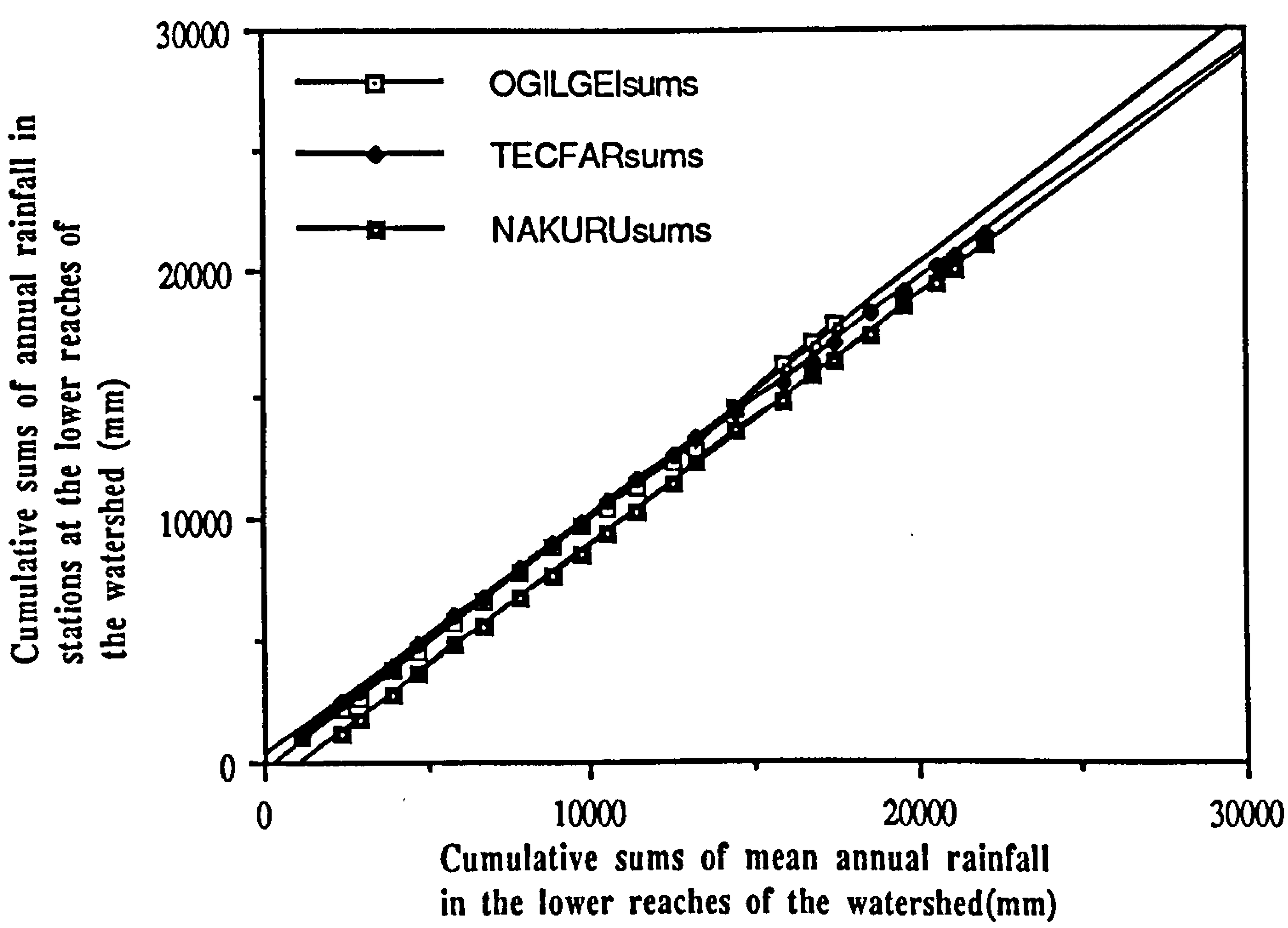
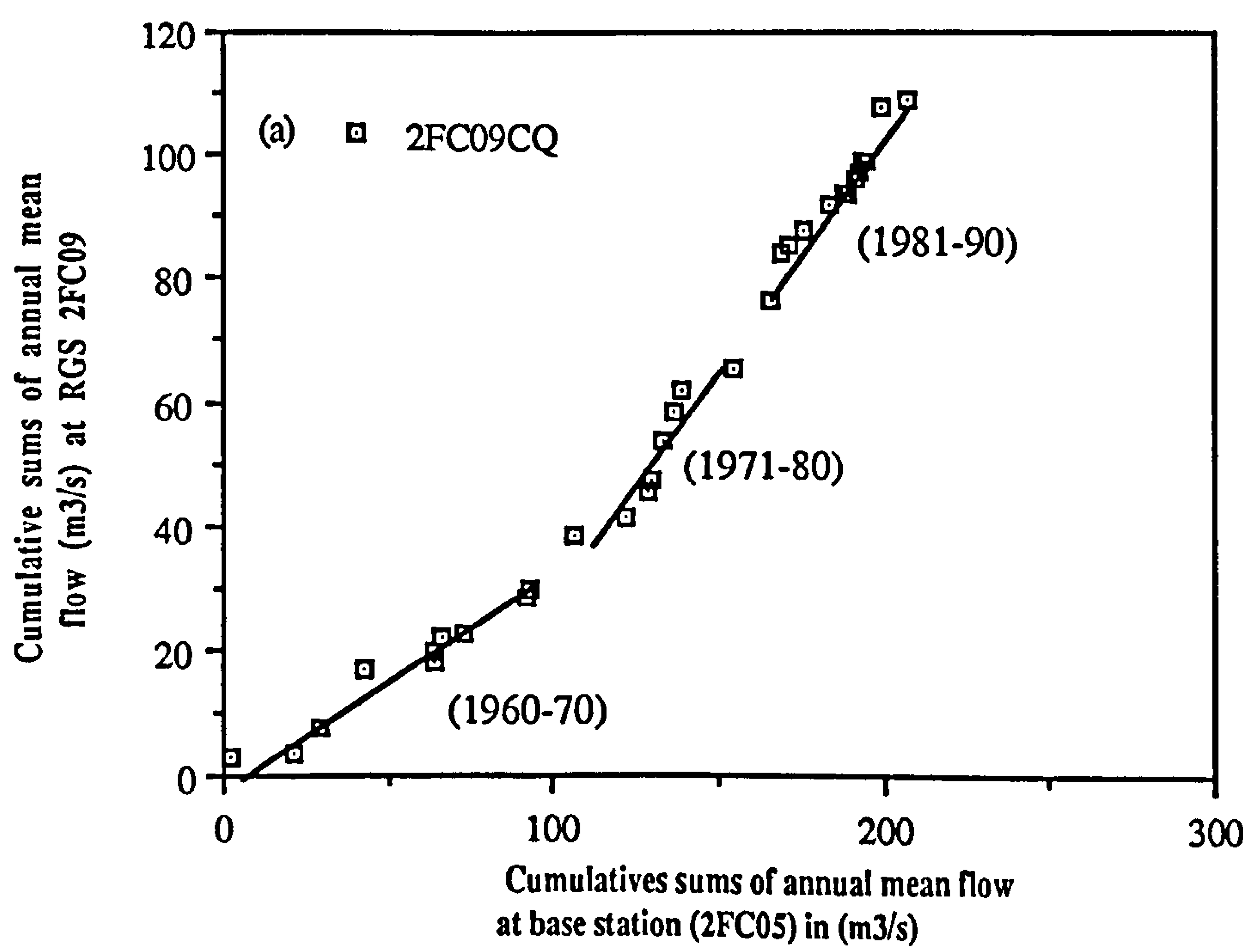


Figure 4.9. Double mass curves for accumulated mean annual rainfall at the lower (disturbed) parts of the watershed and accumulated mean annual rainfall from Ogilgei, Technology and Nakuru rainfall recording stations.

Table 4.12. The slopes of double mass curves for 1960 to 1990 period for the lower watershed mean rainfall and adjacent rainfall stations.

RAINFALL STATION	SLOPE	CORRELATION
OGILGEI	+ 1.0238	$R^2 = 0.999$
TECHNOLOGY FARM	+ 0.96757	$R^2 = 1.000$
NAKURU	+ 1.0013	$R^2 = 1.000$

The graphicall representation in Figures 4.8 and 4.9 in addition to the statistical results in Tables 4.11 and 4.12, all describe clearly the quality and consistency of the rainfall data in the watershed. A correlation of the double mass curve, R^2 close to 1.0 in the upper and lower parts of the watersheds confirms the good quality data. In the upper reaches however, NPBRS station has more rainfall totals and higher slopes (1.9941) than the rest of the stations. The lower reaches show a consistent homogenous data. In addition this analysis has shown that the upper reaches has more rainfall than the lower reaches.
Similar approach was applied to the streamflow data. Because of gaps in the data series, the streamflow series from 2FC05, located in the upper forested part of the watershed, was considered as the base station. The cumulative annual total streamflow from 2FC05 series were plotted against the corresponding accumulated series from 2FC09 for the period (1960-1990). The resulting plot is presented in Figure 4.10. For these partial series, a straight line was fitted which is used for comparison. The slope of the straight lines from this figure were extracted and presented in Table 4.13.



Figures 4.10. Double mass curves of accumulated streamflows for 2FC05 and 2FC09 series for the periods 1960-1990.

Table 4.13. The slopes of the double mass curves (plots) for the three selected periods for 2FC05 and 2FC09 flow series.

PERIODS	SLOPE	CORRELATION
1961-1970	0.3373	$R^2 = 0.96$
1971-1980	0.8460	$R^2 = 0.96$
1981-1990	0.7413	$R^2 = 0.91$
1960-1990	0.5648	$R^2 = 0.96$

The double mass curve in Figure 4.10 show a break in 1963/65, 1970/73, 1978/80, 1984/85 and 1989/90. A slight change exist in mass plot slopes during these periods although it is smoothened when regressed. These have been isolated and regressed individually for the periods between 1960 and 1990. The results show that the two data series do not belong to the same population. They however seem to move together as shown in their close R^2 values, although at different scales and magnitudes.

Temporal fluctuations of the data series were further examined to ascertain whether they could reasonably be expected in a natural condition, and spatially, the data from nearby stations or adjacent station were examined to check whether they behave simultaneously, within an appropriate tolerance range. To achieve this, a validation window method recommended in Herschey et. al. (1978) and Herschey (1985) was used. The method recognises the fact that the continuous form of a river's stage hydrograph is suitable for a time series examination, which can then be designed to detect spurious behaviour. Since variations are large, quality control was necessary to achieve a control procedure which filters out any significant errors from the data.

The use of validation window procedures to identify suspect streamflow data, fixes natural variation of annual flow to be between -60 and +60 % deviations from the long-term mean value. Flow series falling between -60 to -150% and +60 and 150% indicates that erroneous data are present for individual periods and hence, a detailed data inspection is required. Any variation below -150%, and above +150% , indicate, invalid data, and the particular values are rejected, unless there is a good explanation, such as from rainfall data that confirms that events of the observed magnitudes occurred during those periods.

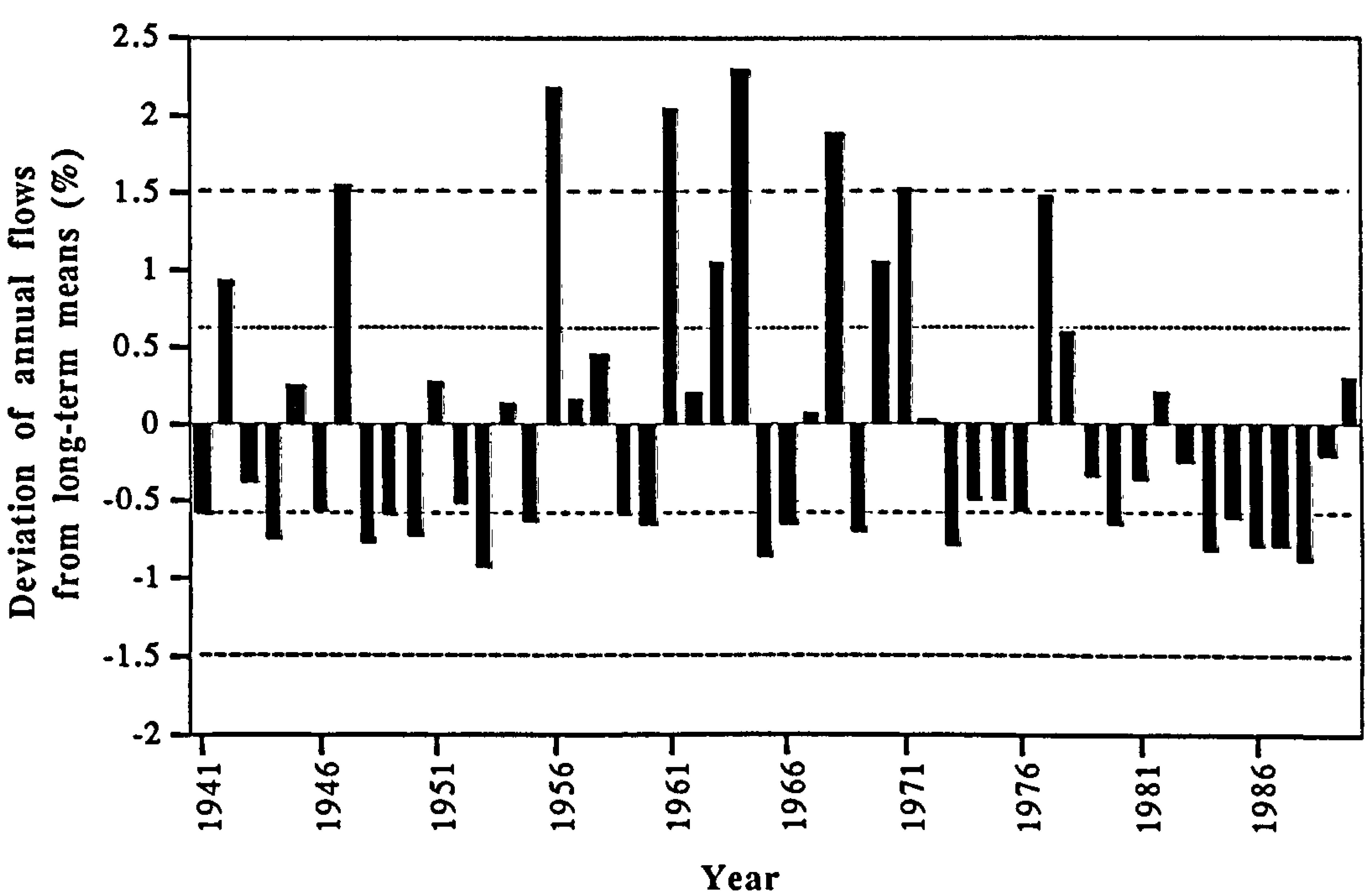
However, gross errors are often easily recognised. Hence, by considering, the probable causes of faulty data, simple validation procedures were designed to pinpoint corrective action. This procedure was applied to the streamflow data series and the results are presented in Figures 4.11 and 4.12 for gauging station 2FC05 and 2FC09 respectively. In addition, physical checks to ensure that each batch of data series contained the correct number of days and the monthly mean flows was indeed, the average of the flows on the specific days. Correction and harmonisation of the units and detection of misplacement of decimal points in the rainfall and streamflow series was carried out.

The results shown in Figure 4.11 suggests occurrence of invalid flow data in 1957, 1962, and 1968 with values greater than the 150% deviations. Similar observations and corrections were made for the 2FC09 flow series shown in Figure 4.12 where 1964 and 1978 were identified. These however, were physically checked and confirmed as reliable data that corresponded to high rainfall totals during those years.

4.8.2. Analysis of Gauging Stations Stability

4.8.2.1. Analysis of Rating Curve Stability

The purpose of analysing the gauge station and rating curve stability was to examine the historical trends of the rating curves and establish any discontinuity in the flow series which might have affected the rated discharges. If found, it is then necessary to establish whether the changes could have been caused by human-induced interventions. Cross-sectional reduced levels (RLs) at RGS 2FC05 in 1952 were compared to those measured in 1992. The same procedure was applied to RGS 2FC11 for the period 1974 and 1992. The historical cross-section for the 2FC09 gauge station was not available for analysis. The results are presented in Figures 4.13 and 4.14 for 2FC05 and 2FC11 gauging stations respectively.



Figures 4.11. Validation window of annual streamflow residuals at gauging stations 2FC05 in the Enjoro river watershed .

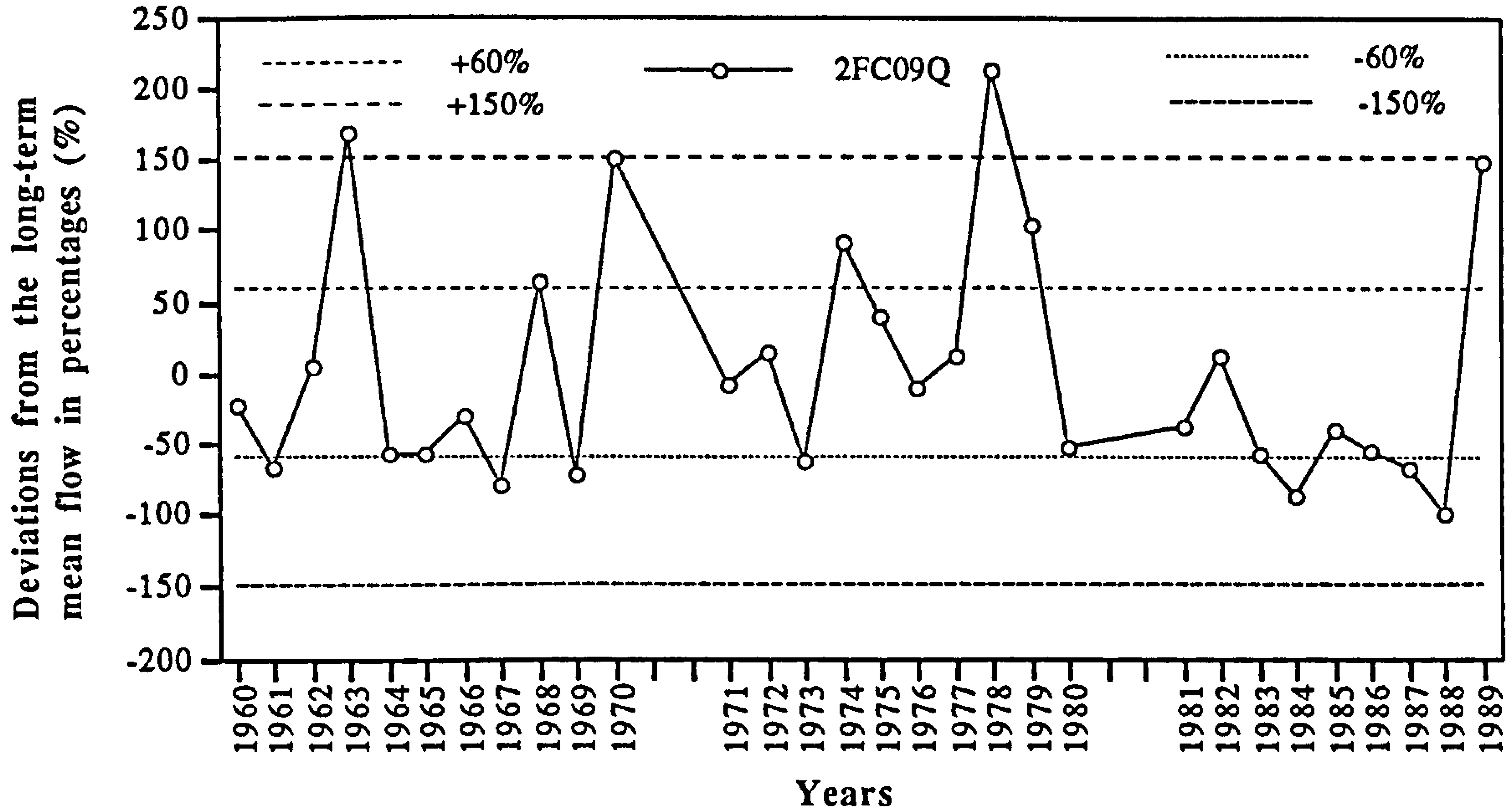


Figure 4.12. Validation window of the annual streamflow residuals at gauging station 2FC09 in the Enjoro river watershed.

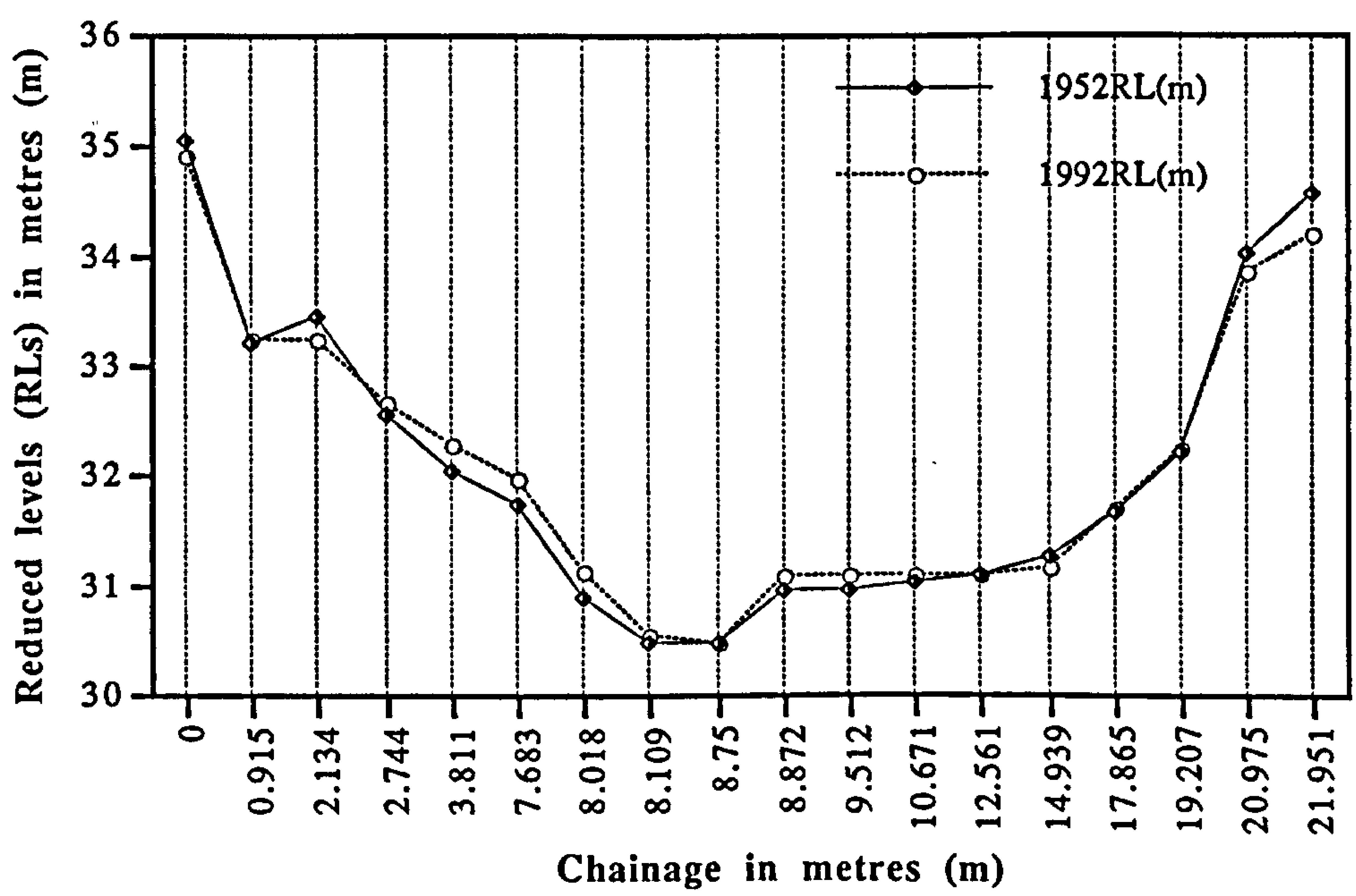


Figure 4.13. Cross-sectional changes of 2FC05 for the period 1952 and 1992.

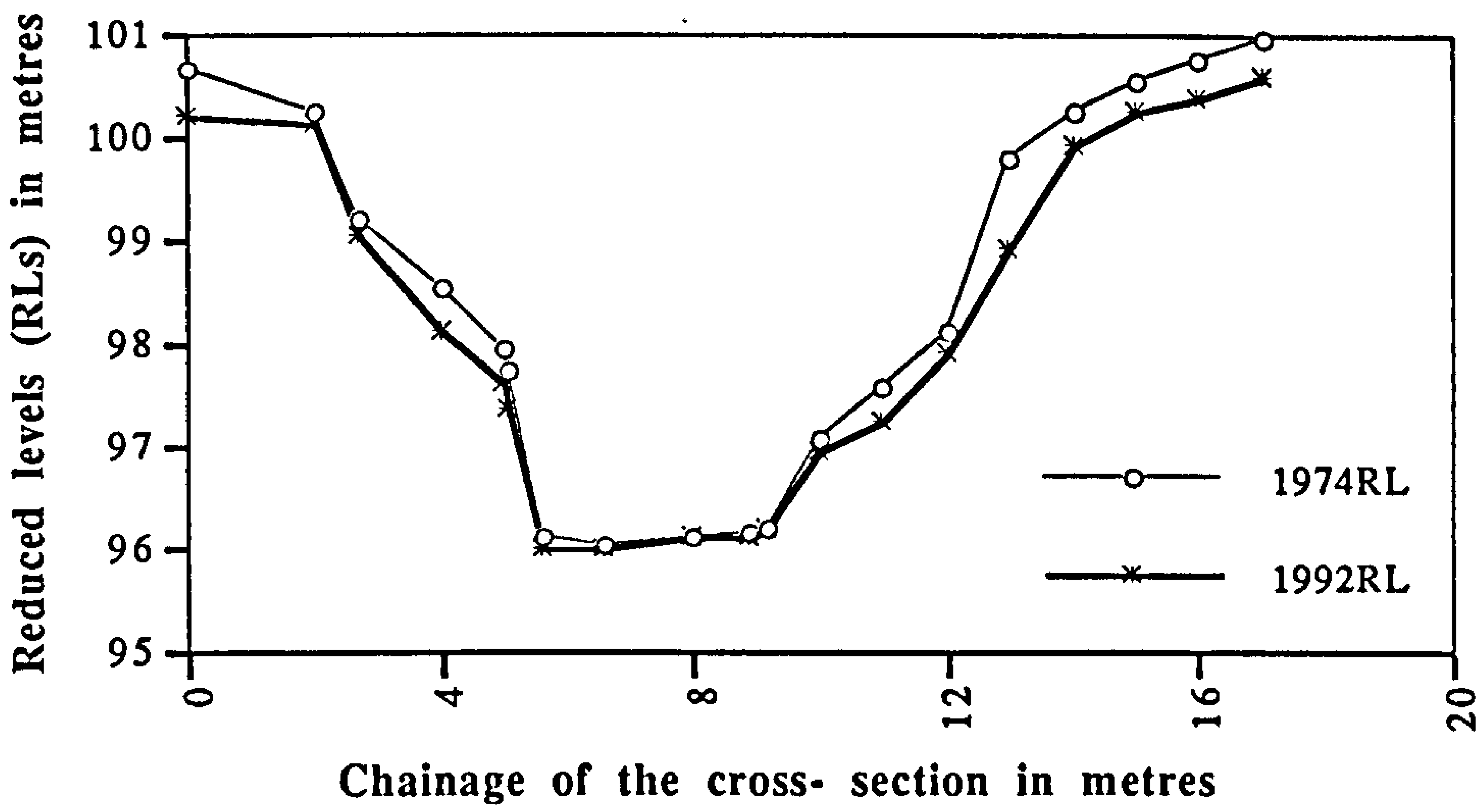


Figure 4.14. Cross-section changes of RGS 2FC11 the period 1974 and 1992.

The difference (∂H) in the RLs in the two cross-sections during the periods examined have no significant deviations at 95 % significance level. It was therefore, accepted that the stations were stable and the rating curves developed and used to compute the discharges were reliable. Differences in the data series will be assumed to have been caused by external factors other than the structural changes of the gauging stations.

4.8.2.2. Tests of Changes in Hydraulic Characteristics

To confirm the results from the cross-sectional variations discussed, a further test was necessary. This involved analysing the relationship between the river stage (H) and actual current meter measurements (CMS) for the entire record. A set of 77 current meter discharge measurements and stages for RGS 2FC05 were examined and graphically presented in Figure 4.15. Similarly, 63 sets of current meter measurements and corresponding stages were examined in Figure 4.16 for RGS 2FC11. The results from the RGS 2FC09 flow series are given in Figure 4.17.

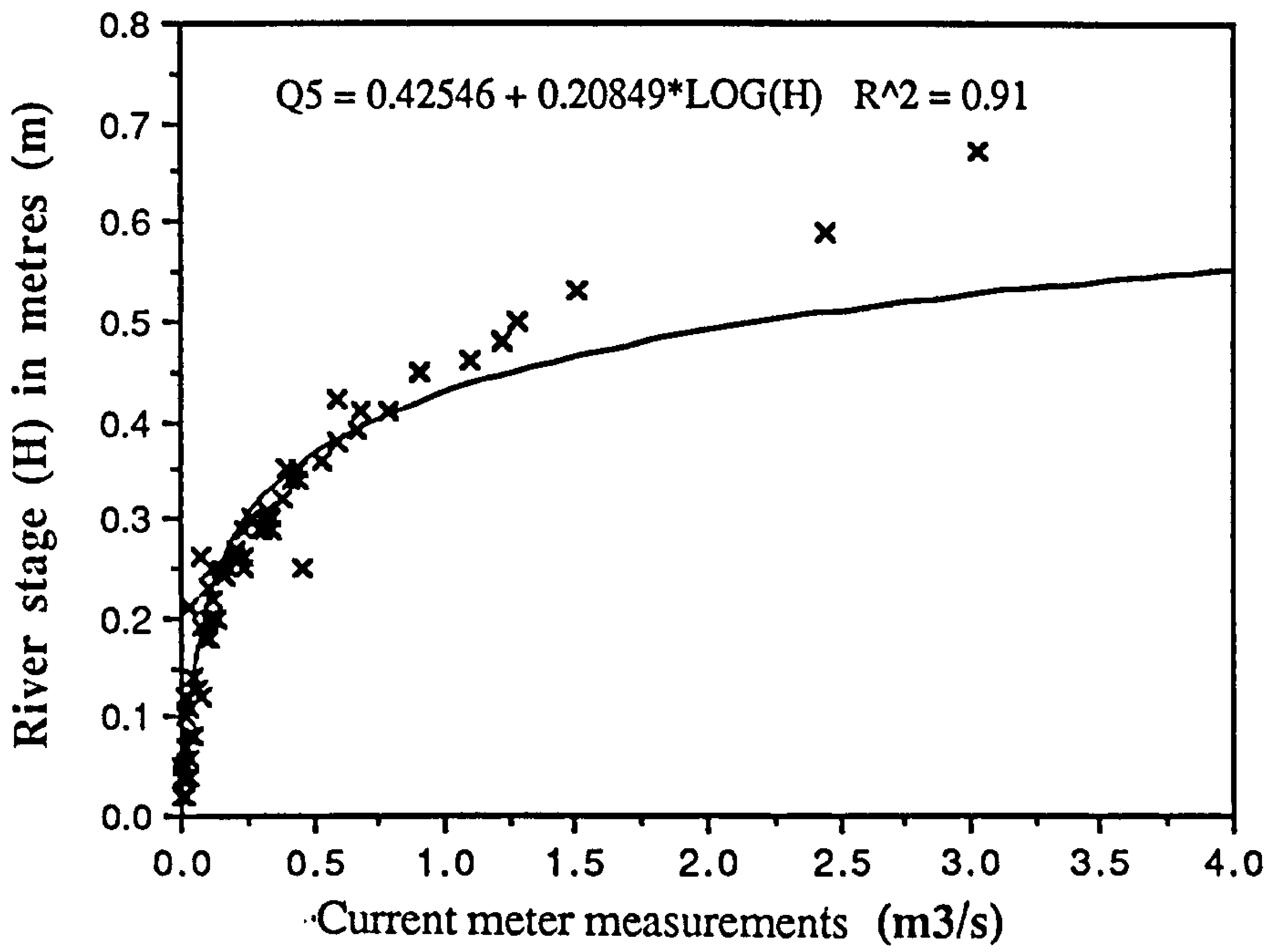


Figure 4.15. Stage-discharge relationship in RGS 2FC05 for the period 1949-1992.

Figures 4.15, 4.16, and 4.17 show no major change in the scatter curve, except a single outlier in the 2FC05 series, which was ignored and removed from the data set. There was therefore no major change in the curves during the period considered, and hence, the rating curves and equations used were considered stable. The same is said for the 2FC11 except that, two sets of equations were used in this case for $H \leq 0.4$ and $H > 0.4$ m where the slope of the curve assumed a different distribution.

The 2FC09 series however was stable and maintained a single distribution during the period under examination. Prior to this period(1974), the station used a Cipolleti weir whose rating were based on a standard Cippolleti weir equation. Therefore, the analyses further confirmed the previous findings from the stream cross-sections, that the stations were relatively stable throughout the period. Hence changes observed in the flow series would be attributed to external factors such as the anthropogenic influence of the land use and changes in the climate.

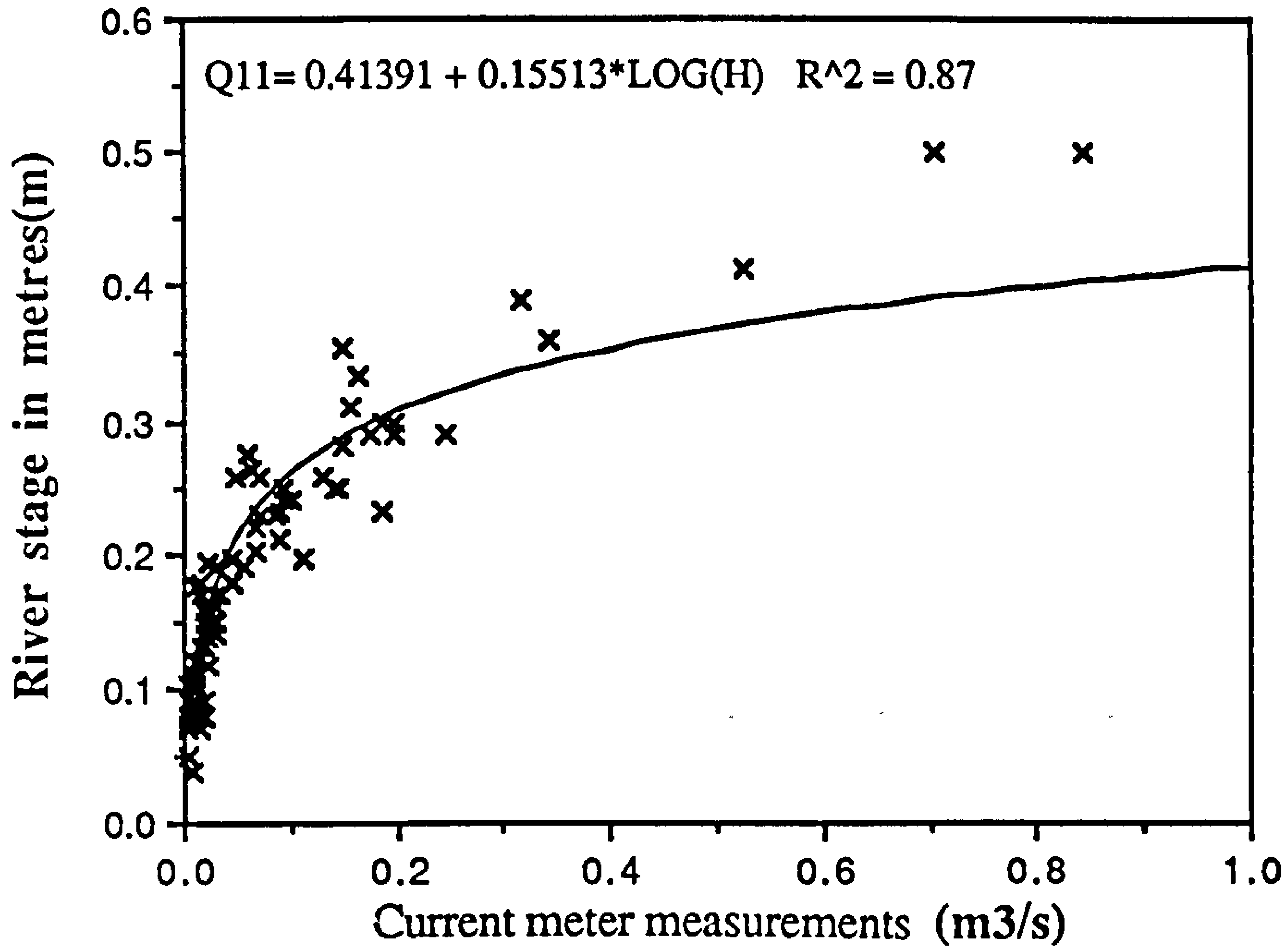


Figure 4.16. Stage-discharge relationship in RGS 2FC011 for the period 1966-1992.

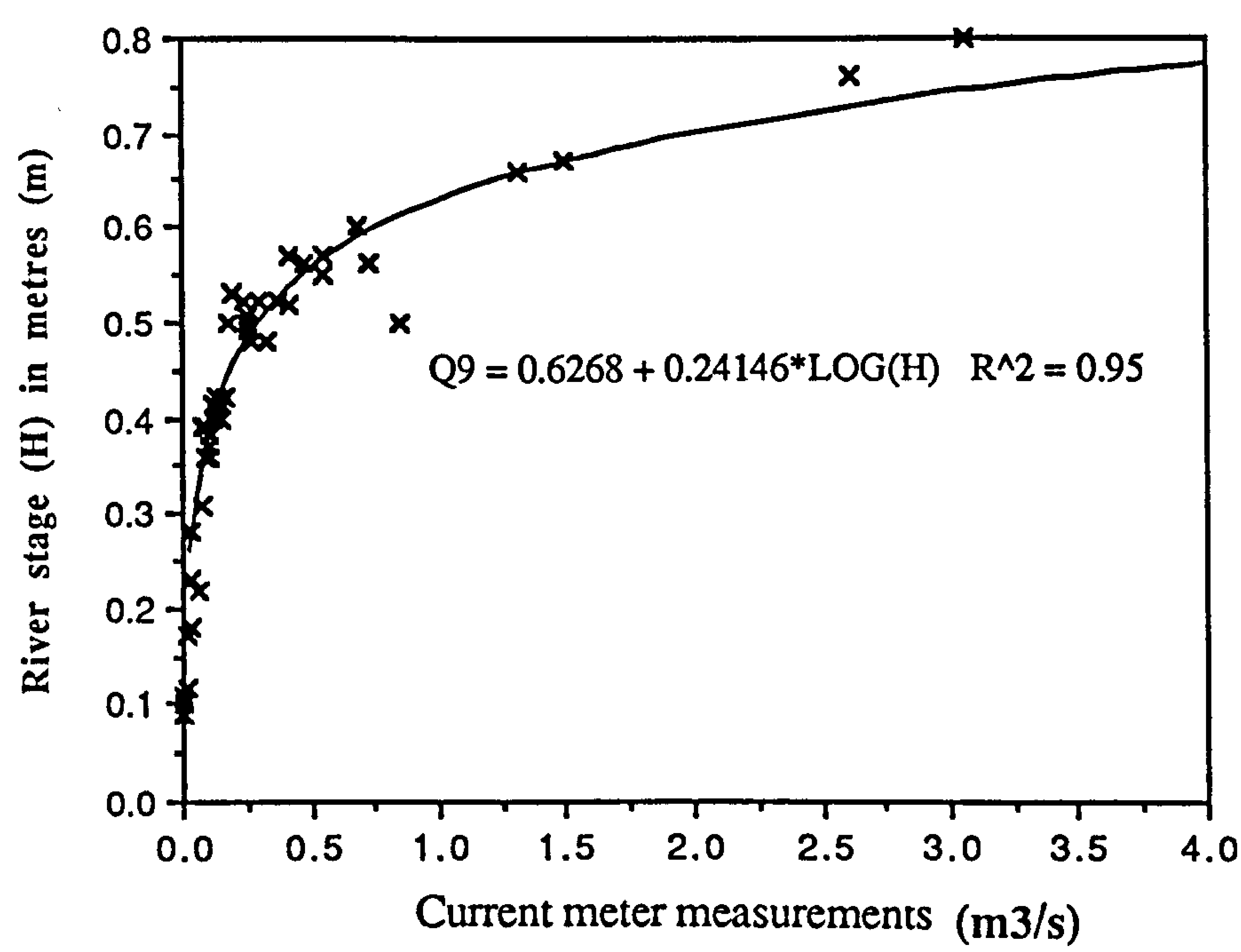


Figure 4.17. Stage-discharge relationship in RGS 2FC09 for the period 1974-1992.

These results however, were not accepted without examining the nature and periodic behaviour of the gauging stations cross-sectional hydraulic characteristics. An analysis that takes into account the whole section, its slope and roughness coefficient was necessary. To achieve this objective, the Manning's equation was substituted into the WMO (1980) hydraulic coefficient, κ

$$\kappa = \frac{S^{1/2}}{n} \tag{4.1}$$

and with the continuity equation $Q = A*V$ (4.2)
where V is the Manning's velocity

$$V = \frac{R^{2/3}S^{1/2}}{n} \tag{4.3}$$

$$\text{which gives } Q = \frac{A * R^{2/3}S^{1/2}}{n} \qquad 4.4)$$

and by combining equation 4.1 and 4.4 it provided an estimated hydraulic coefficient from measured cross-sectional variables hence,

$$\kappa = \frac{S^{1/2}}{n} = \frac{Q}{A * R^{2/3}} \tag{4.5}$$

From equation 4.5 the coefficient κ, that combines the hydraulic characteristics and roughness coefficient can be established using the current meter discharges, the cross-sectional area, the river stage (H), and the computed hydraulic radius of the river's cross-section.

where, Q is the current meter measured discharges (m^3s^{-1})
R is the hydraulic radius of the river's cross-section (m)
A is the cross-sectional area of the river (m^2).
n is the Manning's roughness coefficient ($sm^{-1/3}$)
κ is the overall hydraulic coefficient ($s^{-1}m^{1/3}$)
V is the profile mean velocity (ms^{-1})

Thus the value of the coefficient (κ), which combined the hydraulic slope and roughness coefficient was deduced from each observation of the river stage (H) and the rate of flow (Q). The sample means and coefficients of variations (CV) of κ in four successive periods(1949-1992) for 2FC05 series,, three periods (1966-1992) for 2FC11, and two periods (1974-1992) for 2FC09 data series were analysed using the Student's t-test (Chebotarev, 1974) to confirm that each subperiod κ-coefficients belonged to the same population and hence, there has been no major change in the cross-sectional hydraulic characteristics. The results from the tests are presented in Tables: 4.14, 4.15, and 4.16 for the hydraulic coefficients (κ) from RGS 2FC05, 2FC11, and 2FC09 respectively. The student t-statistics for 2FC05 RGS κ -coefficients for a paired degrees of freedom (df) of 20 and a two-tailed t-test at 5% level of significance is 2.228. All the computed t-values for the selected periods fall within t<-2.228

and t> 2.228 and within the 1949-1992 (df=77, 5%) t-value of 1.990 and -1.990. The coefficients therefore are accepted as belonging to the same population.

Table 4.14. Student's t-test values of k-coefficients from RGS 2FC05.

Partial Series	Mean	Std Dev.	CV	1961-1970	1971-1980	1981-1990
1961-1970	0.140	0.092	0.658	1.000	0.126	0.524
1971-1980	0.105	0.103	0.978	0.126	1.000	0.111
1981-1990	0.087	0.093	1.067	0.524	0.111	1.000

The results from the 2FC11 cross-section are presented in Table 4.15. The one-tailed t -test at $t_{\alpha/2} = 5\%$, df = 63 is 1.998. The null hypothesis is rejected if t < -1.998 or t > 1.998. Secondly the subperiod paired t-test for df=9 and 5% siginficance is 2.228. The t-values ranged between -0.804 and 1.975 for the periods examined. These are within the critical ranges and hence, there is no reason to reject the null hypothesis that the means of the current meter discharge measurements have not changed in the course of time. The data set therefore belongs to the same parent population.

Table 4.15. Student's t-test values of the hydraulic (k) coefficients from RGS 2FC11

Partial Series	Mean	Std	CV	1966- 1975	1976- 1985	1986-1992
1966-1975	0.056	0.046	0.821	1.000	- 0.804	1.427
1976-1985	0.071	0.083	1.169	-0.804	1.000	1.975
1986- 1992	0.036	0.031	0.860	1.427	1.975	1.000

The t-statistics for the 2F09 cross-section are shown in Table 4.16. For $t_{\alpha/2} = 5\%$ and df=44 (for the whole period) t-statistic is 2.014. The paired t-value for df=9 and same level of significance is 2.228. The critical levels range is between t < -2.014 or t > 2.014. Since the calculated t -statistic of -1.844 fall within the above range, the null hypothesis of equal means of the k values cannot be rejected. The CV values ranged between 0.8 to 1.0, an indication of a high variability which is uniform over the years. This study assumes that this variability resulted from the daily fluctuation of the flow. Hence CV may not be a good indicator for detecting changes in hydraulic patterns of rivers experiencing human interventions.

Table 4.16. Student's t-test values the hydraulic (k) coefficients from RGS 2FC09

Partial Series	Mean	Std	CV	1974-1983	1984-1993
1974- 1983	0.091	0.088	0.967	1.000	-1.844
1984- 1993	0.153	0.137	0.895	1.844	1.000

On the basis of the students' t-tests and the graphical representations, it was accepted that the hydraulic characteristics of the RGS cross-sections did not change significantly during the period of data record. The river stage and rating equations used by the Ministry of Water Development (MOWD) to compute discharges were accepted as accurate for use.

4.9. CONCLUDING REMARKS

1. The data collection and validation procedures were extensively explored and used. Quality checks and control of the data were achieved and homogeneity tests carried out to ensure a reliable and accurate interpretation of the rainfall and streamflow data series. It was then possible to identify and isolate the external factors that may have influenced the hydrologic regime during the period under study. Although a complete accuracy in data validation is often difficult to achieve, especially when dealing with data collected by several different persons and institutions at different times, these procedures have provided a clear understanding of the Enjoro river watershed physical and demographic characteristics. As a result, it is possible to understand its hydrological regime and hence achieve the objectives of the study.

2. Several methods and approaches were attempted to isolate factors other than human-induced interventions in the watershed. The homogeneity tests and trend analysis of the current meter measured discharges in addition to the gauging station hydraulic characteristics enabled an evaluation of the possible change in the river cross-section and validated the data obtained from rating curves and equations used. The demographic results revealed the watershed to have experienced increased urbanisation and deforestation. Consequently, it was possible to establish the rural-urban migration equivalent to a land use conversion rate of 2.38% annually. Assuming a linear trend and rate, the area is anticipated to be urbanised in less than 22 years. Detailed analyses in chapters V and VI are expected to offer a clearer picture of the scenarios observed since the chapters will relate the changes in land use to observed rainfall and streamflow patterns.

CHAPTER V

RAINFALL AND STREAMFLOW DATA ANALYSIS

5.1. INTRODUCTION TO THE ANALYSIS

The variations in river flows are controlled primarily by the spatial distribution of rainfall, evapotranspiration, and changes in land use. At watershed level, the annual variability of runoff is a key descriptor of these hydrological variables and determines the characteristics of low flow frequency (Gustard and Gross, 1989). The temporal variations of these variables are therefore analysed in this chapter to understand the hydrological processes in the watershed set in objectives 3 and 4. The respective hypotheses are later addressed in chapters six and seven.

Since the behaviour of a river varies from year-to-year, a difficulty exists in distinguishing significant changes over time from short-term variability. Coupled with short and discontinuous records, several analytical approaches were thus attempted. One approach was to examine data from several recording sites and draw inferences about their spatial trends using the temporal patterns of change. This approach is followed in section 5.3 to 5.4 for rainfall regime and section 5.6 for the streamflow regime. Other approaches included the estimation of evapotranspiration that was used with rainfall and streamflow values to compute the temporal water balance in the watershed (section 5.6) and a temporal examination of the rainfall-runoff relations in section 5.7. The study also recognised that while the adopted approaches could assist in the understanding and identification of change in the flow regimes, several changes at a range of spatial and temporal scales occurred simultaneously. Some of these factors that lead to temporal changes in flow regimes are summarised in Arnell (1989) as:

- Changes in the characteristics of the watershed such as; changes in land cover, may produce changes in peak and low flow behaviour downstream,
- Changes in upstream water use: over a period of time both the use of water and the return of effluents may change low flows ultimately,
- Changes in climatic inputs; year-to-year variability in hydrological characteristics are clearly related to variability in climatic inputs

This chapter thus considered the temporal and spatial variations of rainfall and streamflows and drew implications of the results with respect to these limitations. The effects of the watershed characteristics on streamflows are discussed in chapter six.

5.2.1. Basic Rainfall Statistical Descriptors

The annual rainfall descriptors were computed for the seven rain gauging stations to understand its natural pattern and regime. The results are then presented in Table 5.1.

Table 5.1. Basic statistical descriptors of the annual rainfall in the Enjoro river watershed (mm)

Station Name	Mean $\overline{X}$	Std Dev,	Variance s^2	Skew C_s	Kurtosis C_k	Coef. Excess $=C_k-3$	Range	Min	Max	Cv $\frac{s}{\bar{x}}$
Teret	1078.61	282.20	79638.03	-0.131	2.367	-0.633	1041.1	583.7	1624.8	0.26
Nessuiet	1123.31	229.60	52717.74	0.041	2.933	-0.067	915.70	716.1	1631.8	0.20
Egerton	1031.41	237.46	56385.58	0.389	3.520	0.520	1129.4	602.3	1731.7	0.23
NPBRS	950.22	207.37	43003.00	0.051	1.945	-1.055	760.7	552.9	1313.6	0.22
Ogilgei	993.69	320.06	102435	1.078	4.689	1.689	1298.1	540.0	1838.1	0.32
TecFarm	936.56	203.44	41386.24	-0.121	2.637	-0.363	706.30	533.5	1239.9	0.22
Nakuru	955.26	193.26	37573.29	-0.541	2.622	-0.378	651.05	584.1	1235.2	0.20

These basic descriptors are the mean rainfalls, $\bar{x}$ the standard deviation, s, the coefficient of variation, C_v, skewness, C_s, and kurtosis, C_k. The mean is the commonest measure of the centroid value, and is computed from the sum of all the rainfall observations and divided by the number of years of record, N:

$$\overline{X} = \frac{\sum_{i=1}^{N} X_i}{N} \tag{5.1}$$

$\bar{x}$ becomes a better estimate of the mean rainfall as N increases, and it approaches the population mean, μ as N approaches infinity. The standard deviation descriptor is also a useful measure of the spread of the hydrological data series. For a sample of N observations, $X_1, X_2, X_3, \ldots X_n$, the standard deviation (s) is:

$$s = \frac{\sqrt{\sum_{i=1}^{N} (x_i - \bar{x})^2}}{(N-1)} \tag{5.2}$$

where $\bar{x}$ is the mean of the hydrologic variable (s) becomes a better estimate of the population standard deviation, σ, as N increases towards infinity. The coefficient of variation (Cv) is represented as the ratio of the standard deviation to the mean rainfall ($s/\bar{x}$). It is one of the diagnostics used to measure the variation of hydrological data series. The parameter ranges from zero for no variation to a number greater than one for an extreme variation.

The most preferred measure of the coefficient of skewness (Cs) use the third power of the deviations of the mean denoted as:

$$\mu_3 = \frac{\sum (x_i - \mu)^3}{N} \tag{5.3}$$

while the coefficient of kurtosis (Ck) measures the peakedness of the data series distribution which is assessed from the fourth moment, μ_4:

$$\mu_4 = \frac{\sum (x_i - \mu)^4}{N} \tag{5.4}$$

A better descriptor related to C_k is the coefficient of excess which defined as C_k-3. If the coefficient of excess is negative or positive, the corresponding frequency distribution of the data series is termed as leptokurtic and platykurtic respectively.

A graphical plot of the Cv against the mean of the annual rainfall for the seven rainfall recording stations is presented in Figure 5.1. It shows a certain amount of scatter and a rather decreasing trend with increasing means of annual rainfall. The distribution of the coefficient of skewness (Cs) over the watershed assumes the characteristical pattern shown in Figure 5.2 and more or less similar to that displayed by the Cv, although it is slightly skewed to the left: an existence of no trends to the annual mean rainfall. These variations however, may be considered insignificant because of the concentration of the scatter points around zero values. Ogilgei is the only station with an exceptionally high skew value. The unusual behaviour of this station from the other stations was attributed to an exceptionally high and localised convectional rainfall in 1978. The Cs however, do not have definite patterns with respect to mean annual rainfall. The high Cs and low Cv observed in the stations could be attributed to their being located along the equator. The results correspond to patterns of rainfall characteristics along the equator and around Lakes Victoria, Kyoga and Albert catchments with CV values ranged of between 0.18 and 0.22 (WMO, 1974 and Shahin, 1984).

The distribution of the Cv of the annual rainfall ranged from the highest value of 0.322 at Ogilgei station to the lowest of 0.203 at Nakuru station. The values from the other stations fall in between. There seem to be not much difference between the Cv values from station to station except in the Ogilgei station. On removal of the 1978 annual rainfall values in the Ogilgei station, the Cv becomes closer to the other stations.

Hence, these suggest that the spatial variability of the annual rainfall may influence the river flow regime downstream. Altogether, the Cv values are a little bit higher in the lower elevation stations than in the higher elevation stations where the rainfall regime is more consistent and less variable.

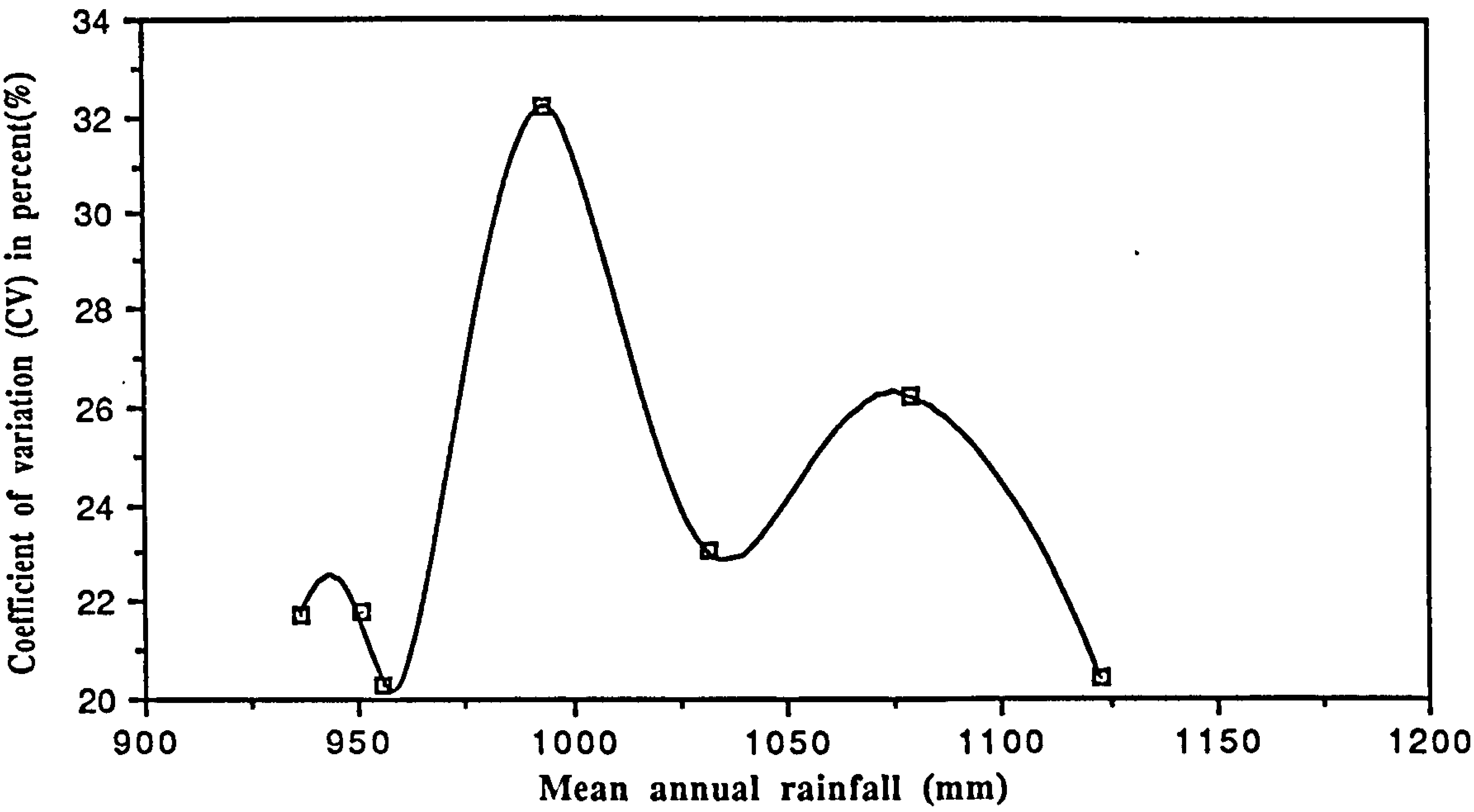


Figure 5.1. Variation of the coefficient of variation of the annual rainfall in the watershed.

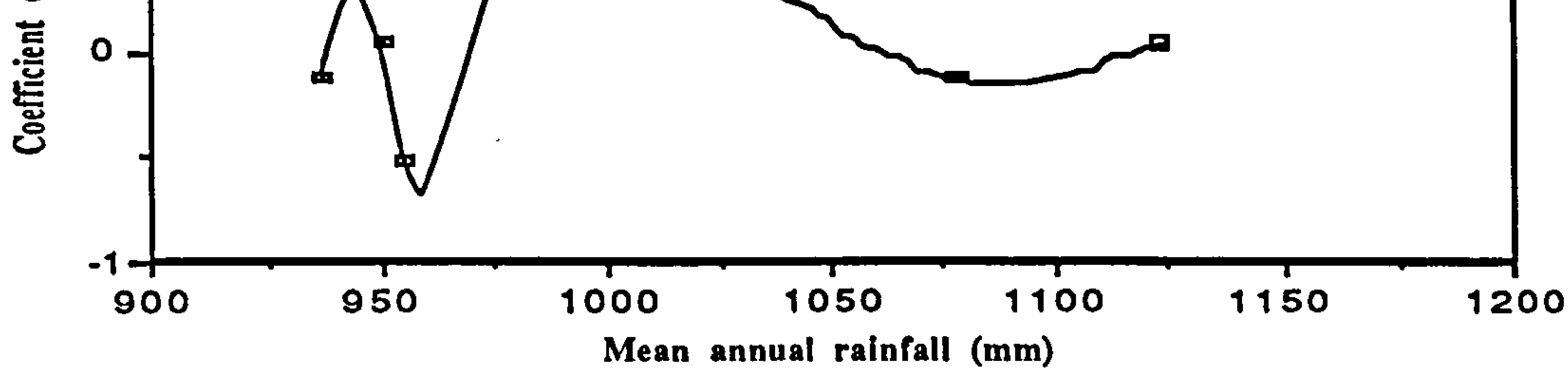


Figure 5.2. Distribution of the coefficient of skewness (Cs) of annual rainfall in the watershed

The variation of the coefficient of kurtosis at the rainfall recording stations is plotted in Figure 5.3. No well defined peakedness except that larger Ck values cluster between 950 and 1030 mm yr^{-1} which is relatively larger than the watershed mean annual rainfall of 850 mm yr^{-1}. Since the coefficient of excess (C_k-3) in Table 5.1 for Egerton and Ogilgei stations are positive, they may be considered to have platykurtic frequency distribution while rest of the stations assume leptokurtic frequency distribution.

On the basis of the results from observing these basic statistical descriptors, it can be concluded that, a higher annual rainfall occurs in the higher elevation stations (Teret, Nessuiet, Egerton and NPBRS) than in the lower elevation stations (Technology farm, Nakuru) except at Ogilgei whose rainfall values deviated because of a single year (1978) with a very high rainfall event. Consequently, the rainfall variability is lower in those stations with higher rainfall values and higher in those with less rainfall amount. The 10-year mean rainfall in the upper parts of the watershed is 839.89 mm/yr in the 1960s, 1074.72 mm/yr in the 1970s, and 810.75 in the 1980s. The higher decadal value in the 1970s is due to the unusually very high rainfall in 1978.

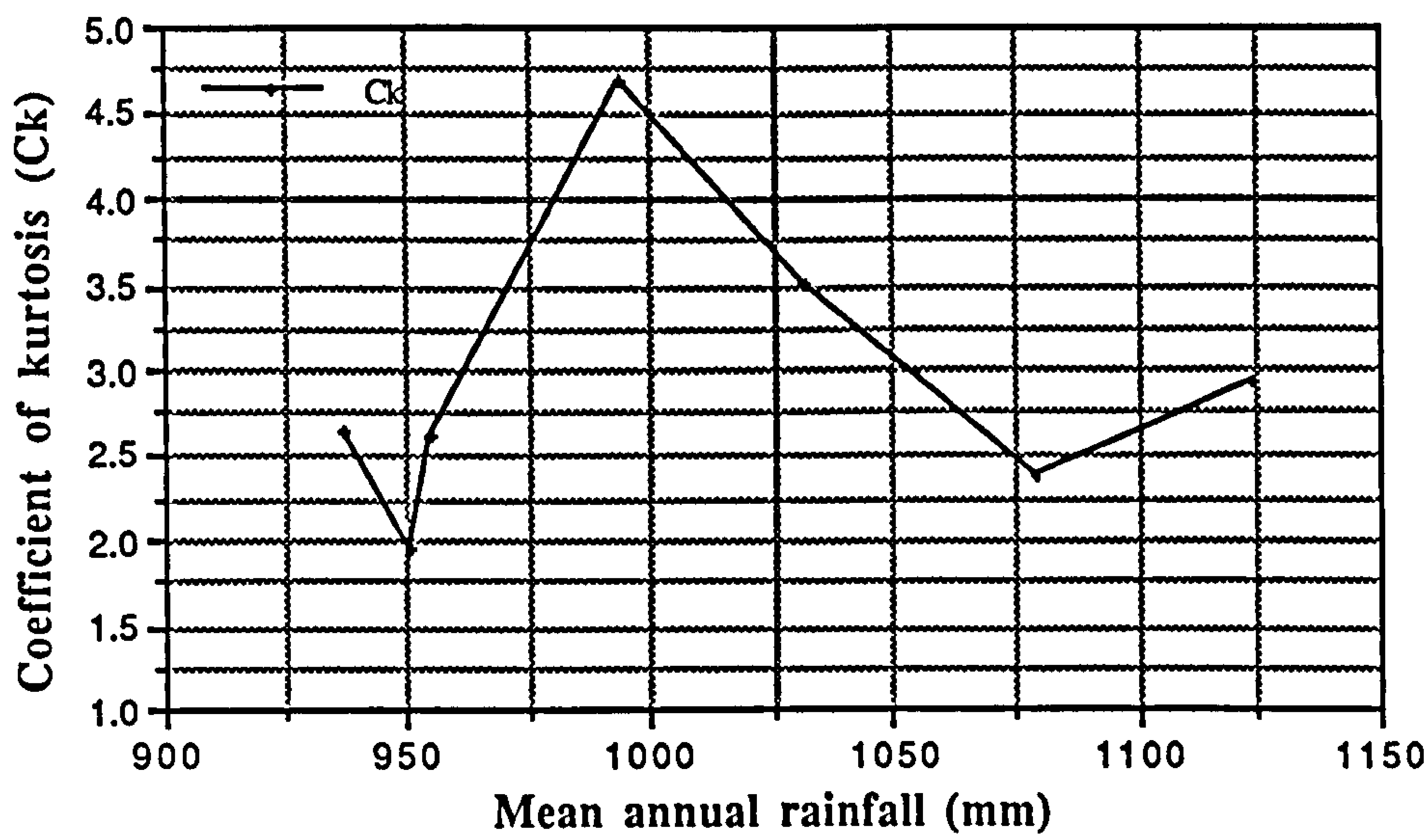


Figure 5.3. Variation of the coefficient of kurtosis (Ck) of the annual rainfall in the watershed

5.3. ANNUAL RAINFALL VARIATIONS

In order to describe the degree of change or effect of the rainfall in streamflow generation, several hydrological indices were used to characterise the intensity of different events. The first index compared the depth of rainfall in a calendar year (R_i) with its long-term mean, $\overline{R}$. The estimation of the rainfall index (RI) from the Ri and $\overline{R}$ followed the representation in equation 5.5

$$RI = \frac{R_i}{\overline{R}} = \frac{mean(R_i stn_1 + R_i stn_2 + \ldots R_i stn_n)}{mean(\overline{R}stn_1 + \overline{R}stn_2 + \ldots \overline{R}stn_n)} \quad (5.5)$$

where stn 1, 2..n are the rainfall recording stations

A given year therefore was described as wet or dry if the ratio of the rainfall index $\frac{R_i}{\overline{R}}$ was less or greater than unity (Beran, 1985). On the basis of the results from the preliminary analysis, the Teret, Nessuiet, Egerton and NPBRS stations were ascribed as higher elevation stations and Ogilgei, TechFarm and Nakuru as the lower elevation stations. The results of this analysis are presented in Figures 5.4.

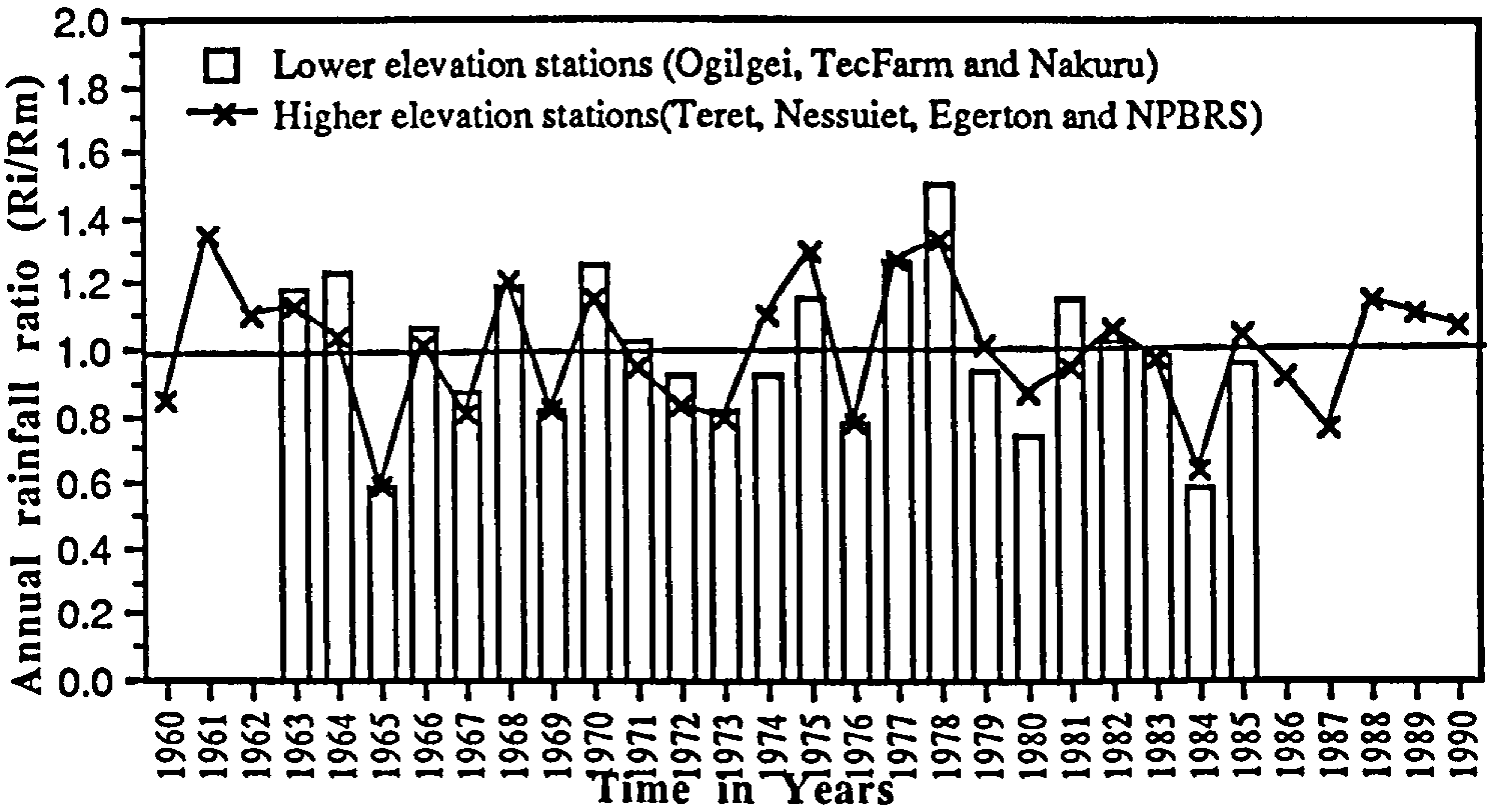


Figure 5.4. Time series of the annual rainfall index ($Ri/\overline{R}$) for the higher and lower elevation rainfall recording stations in Enjoro river watershed.($\overline{R}$.=R_m)

The severity of the rainfall variation was classified as dry year when RI is ≤1.0 and wet year when RI is >1.0. On the basis of on this classification, the time series of the index in Figure 5.5 shows outstanding dry years as 1965, 1971, 1972, 1973, 1976, 1980,1, 1983, 1984, 1986 and 1987. The wet years are 1961, 1964, 1963, 1968, 1970, 1975, 1978, 1989 and 1990. The index ranged from a lowest value of 0.64 in 1984, and highest value of 1.52 in 1978 at both elevations. There is a pattern of increasing number of dry years towards the 1970s and 1980s. The 1980s, however, experienced more drier years than the rest of the study period. On the average, the ten-year mean RI is 1.0 for the 1960s, 1.02 in the 1970s, and 0.97 in the 1980s which indicate that the 1970s was the wettest decade. However, when the unusually very heavy rainfall in 1978 is removed, the RI reduces to 0.99.

A closely related measure is the fractional deviation of the annual rainfall from its long-term mean $\frac{(R_i - \overline{R})}{\overline{R}}$. This measure has been used with success to rank events in a long-climate record and to infer deviations of rainfall patterns due to large scales of human activities (Beran, 1985). The results from the use of this index in this study are given in Figure 5.5 and 5.6 for both elevations. The fractional deviations are below zero during the identified dry and wet years. It clearly identifies the general decreasing trends of the rainfall in most of the 1980s before recovering in 1989-1990 period.

The fractional deviations have also demonstrated that during the driest years, the lower elevation reach of the watershed received relatively lesser rainfall than the higher areas. This is particularly displayed during the long dry years of 1971-1973 and the first half of the 1980s.

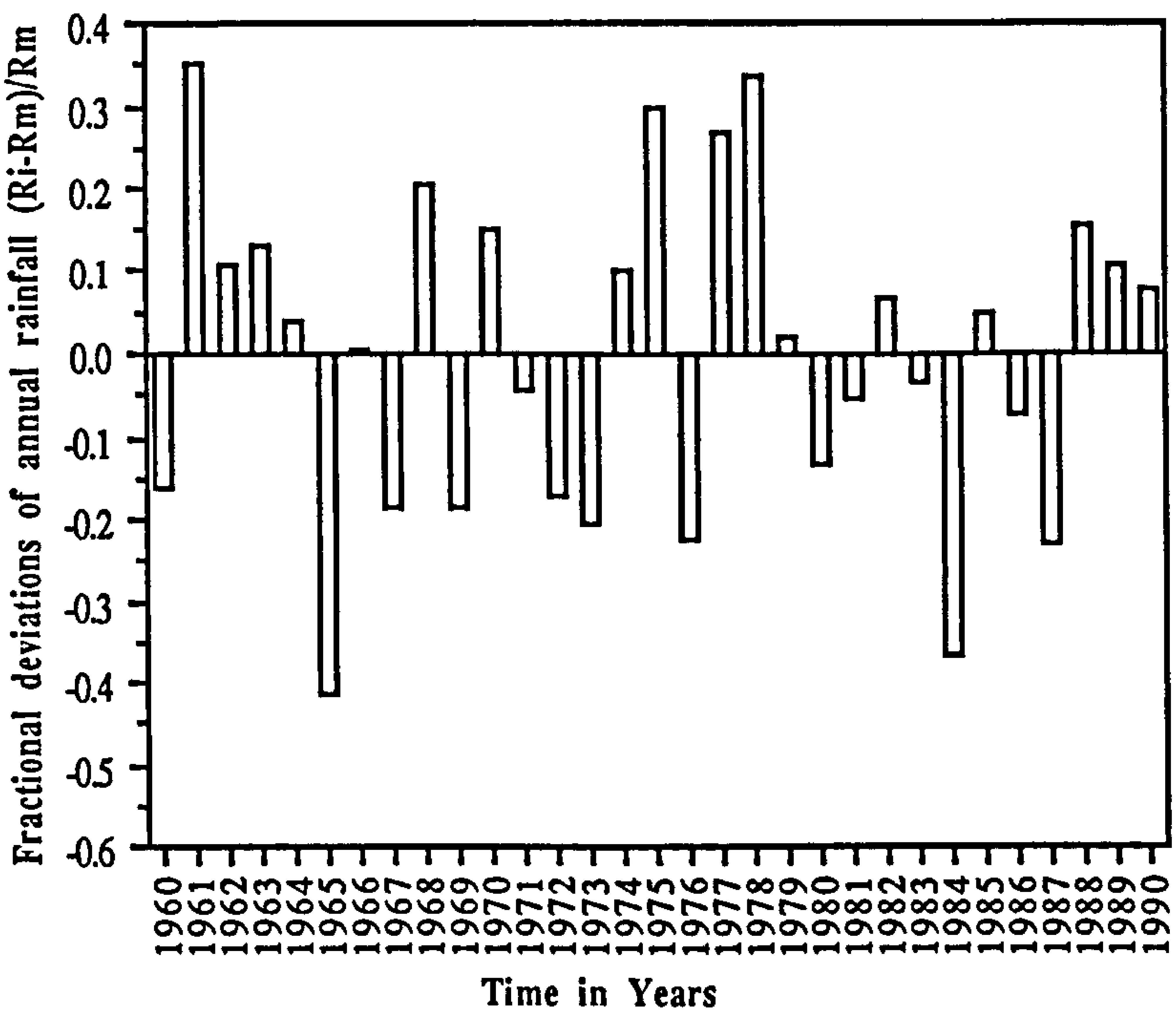


Figure 5.5 Time series of the annual rainfall fractional deviations $\frac{(R_i - \overline{R})}{\overline{R}}$ in the higher elevation reaches of the watershed (in the plot Rm=$\overline{R}$.)

5.3.1. Z-score Values of the Annual Rainfall in the Watershed

The drawback in the (Ri/Rm) approach is its inability to describe adequately the relative severity of the rainfall deficits, which could have helped to detect changes in the rainfall patterns due to human or climatic changes. Hence, there was a need to compare the drought evolution at different locations and times using measures of shortfalls. This approach enables the identification of the different variability at the locations under considerations (Beran et al., 1985). To achieve this, the annual rainfall in the watershed was expressed in terms of the number of deviations above, or below the average in a standardised Z-score value:

$$Z_i = \frac{(R_i - \overline{R})}{s} \tag{5.6}$$

where, Z_i is the standardised rainfall values in the ^{i}th year

R_i is the annual rainfall (mm) in the ^{i}th year,

s is the standard deviation of the annual rainfall,

$\overline{R}$ is the long-term annual mean rainfall (1960-1990 mean)

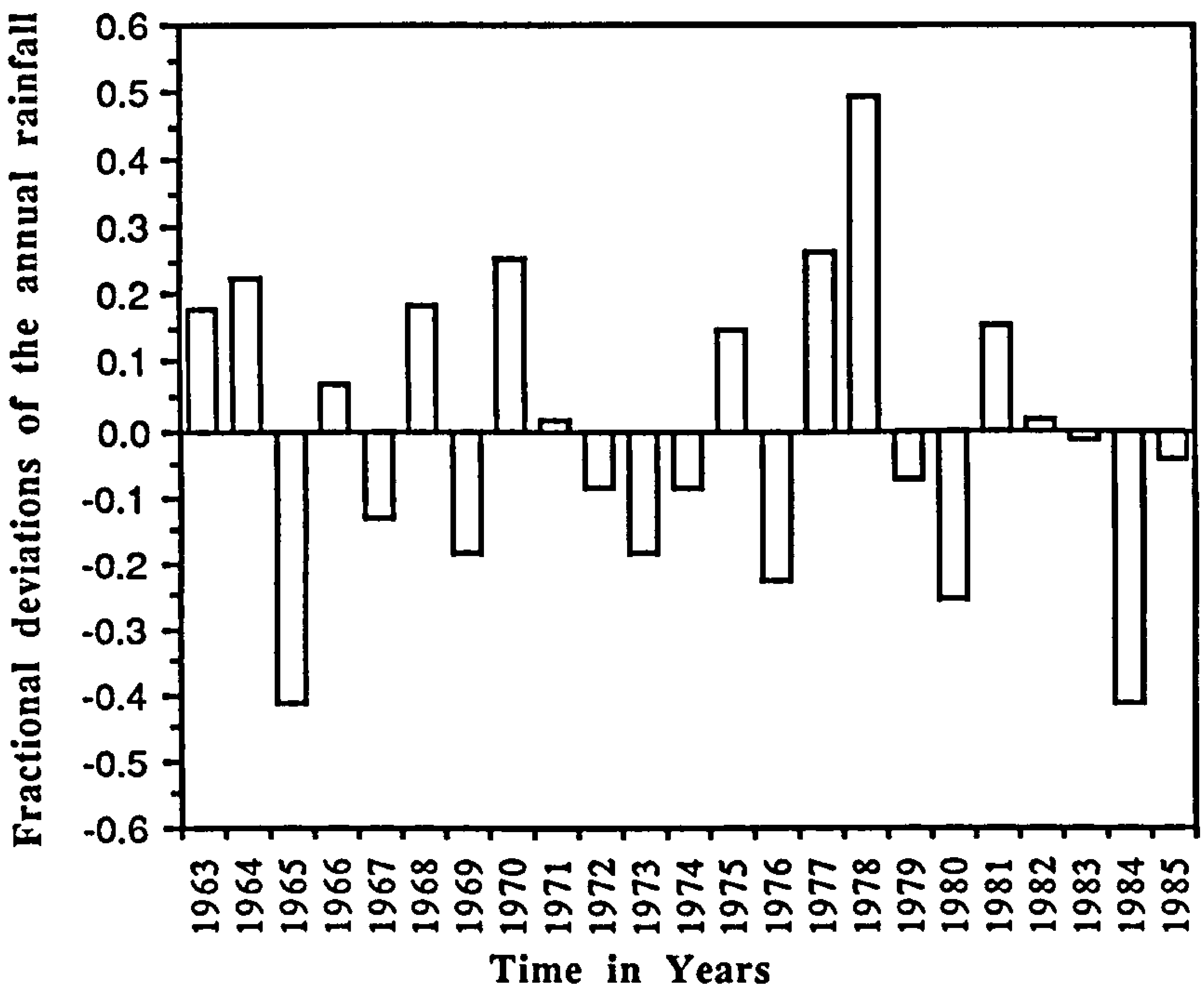


Figure 5.6. Time series of the annual rainfall fractional deviations in the lower reach of the watershed.

Z-score values greater than zero indicate that the ^{i}th year rainfall was wetter relative to the 1960-1990 means. Figures 5.7 and 5.8 were produced which identify the 1960s to have higher Z-score values than the rest of the study period. The Z-scores tend to decline in the 1970s and plummeted in the 1980s. The years 1965 and 1984 are clearly the driest years with Z-score value equal or less than -2 in the entire watershed. The wettest years in the watershed are 1961 and 1978 with Z-score values greater than 1.7

in the higher elevation areas and 1978 with Z-score value of greater than 2.25 in the lower elevation parts of the watershed, an exceptionally wet year.

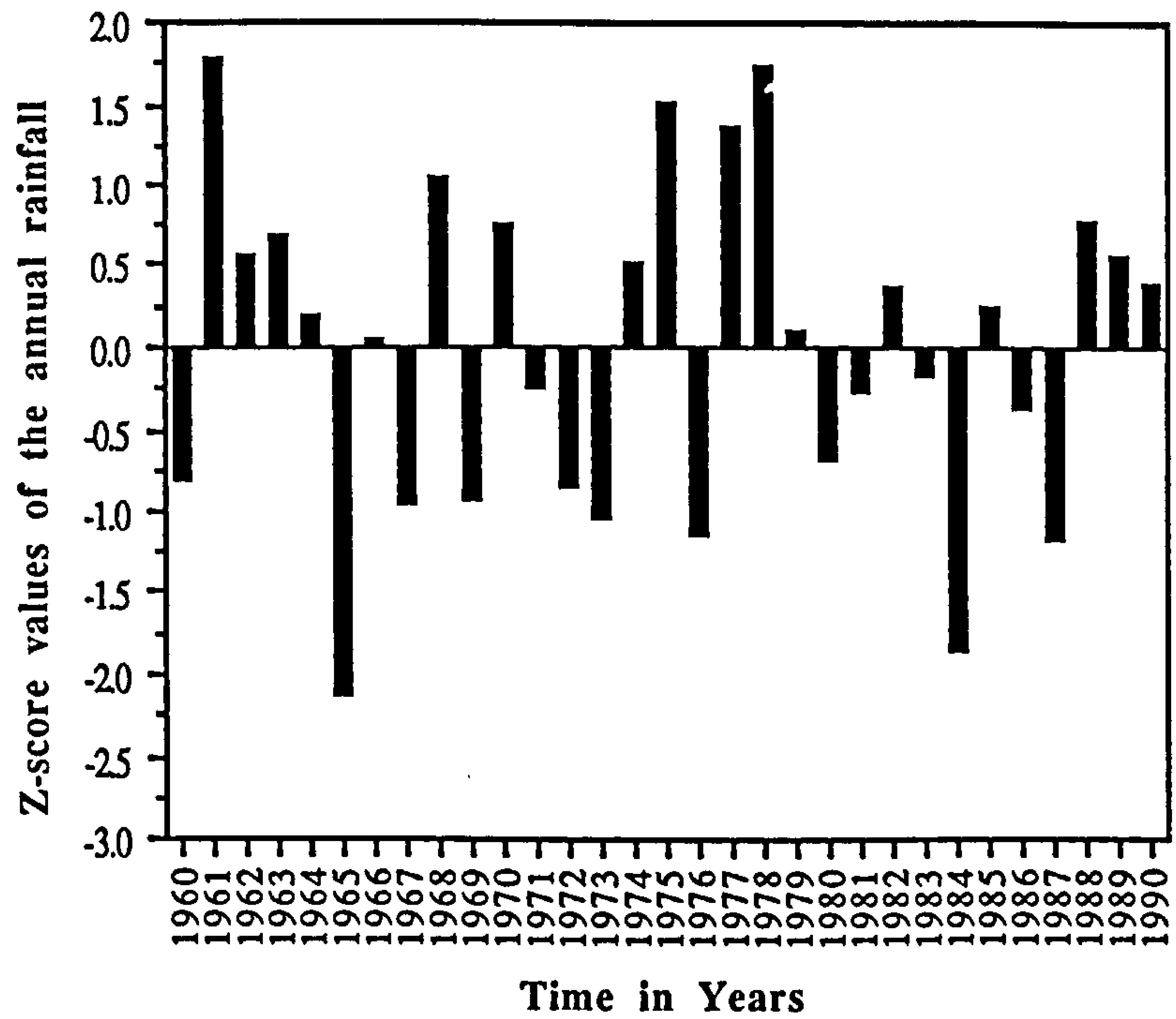


Figure 5.7. Time series of the annual rainfall Z-scores in the higher elevation reaches of the watershed.

A moving average in equation 3.43 helps in the understanding of the temporal behaviour of the series. The averaging of adjacent measurements eliminates the random fluctuations, with the result converging to a description of the systematic trend in the data. It would therefore be possible to isolate the time periods of unusual trends and periodicity's whose cause can be inferred. Thus the rainfall data series were further examined by plotting its time series and a 3-year moving averages. This representation is presented in Figure 5.9. The three- years smoothing interval provided the best separation of the systematic and non-systematic variation and hopefully detect secular trends in the data. This smoothening identified a rather weak 9-10 periodic trend of drought seen from 1965/66, 1973/75 and 1984/85. There is however, no well defined rainfall periodicity, but a rather decreasing trend can be observed.

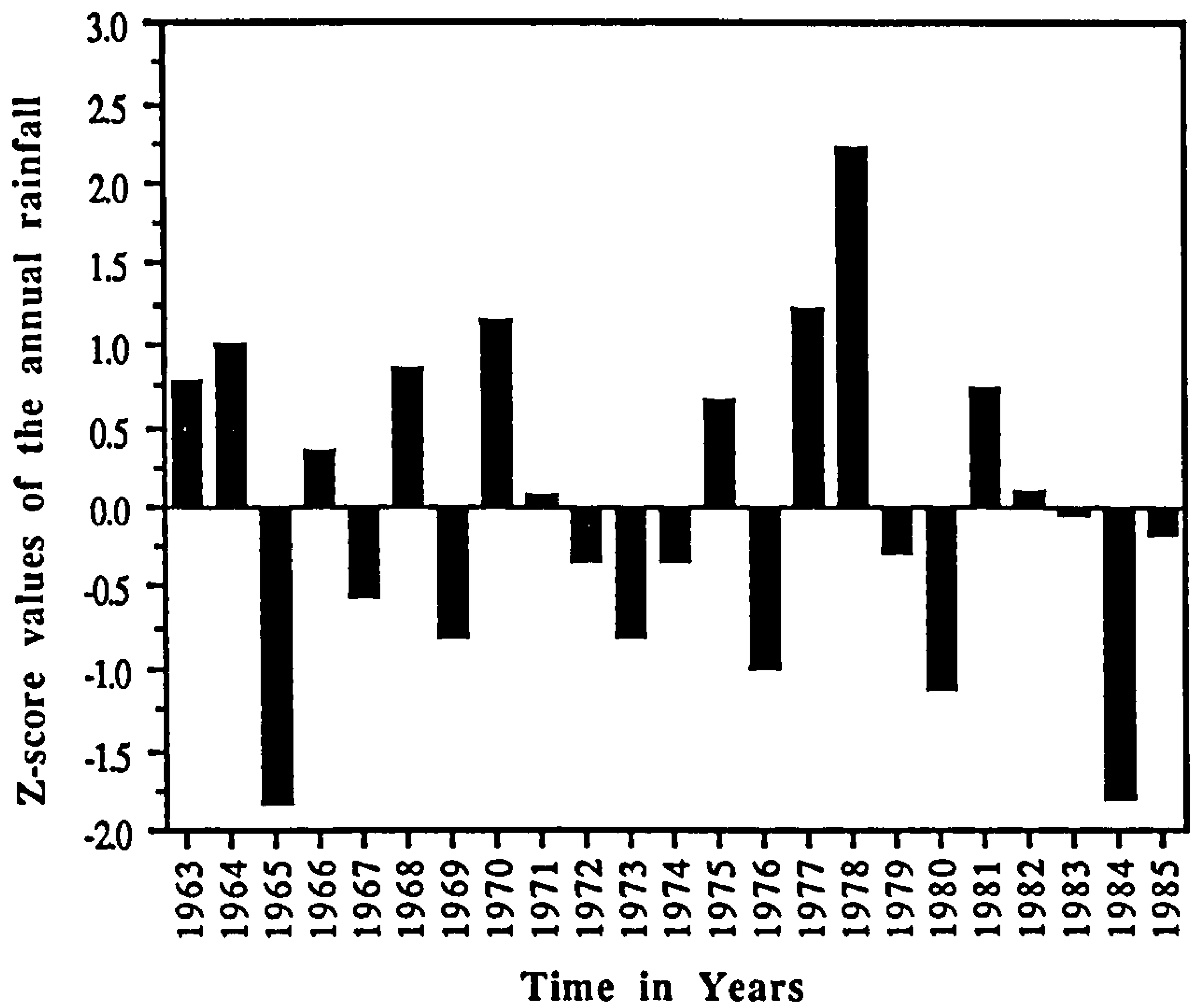


Figure 5.8. Time series of the annual rainfall Z-scores in the lower elevation reaches of the watershed.

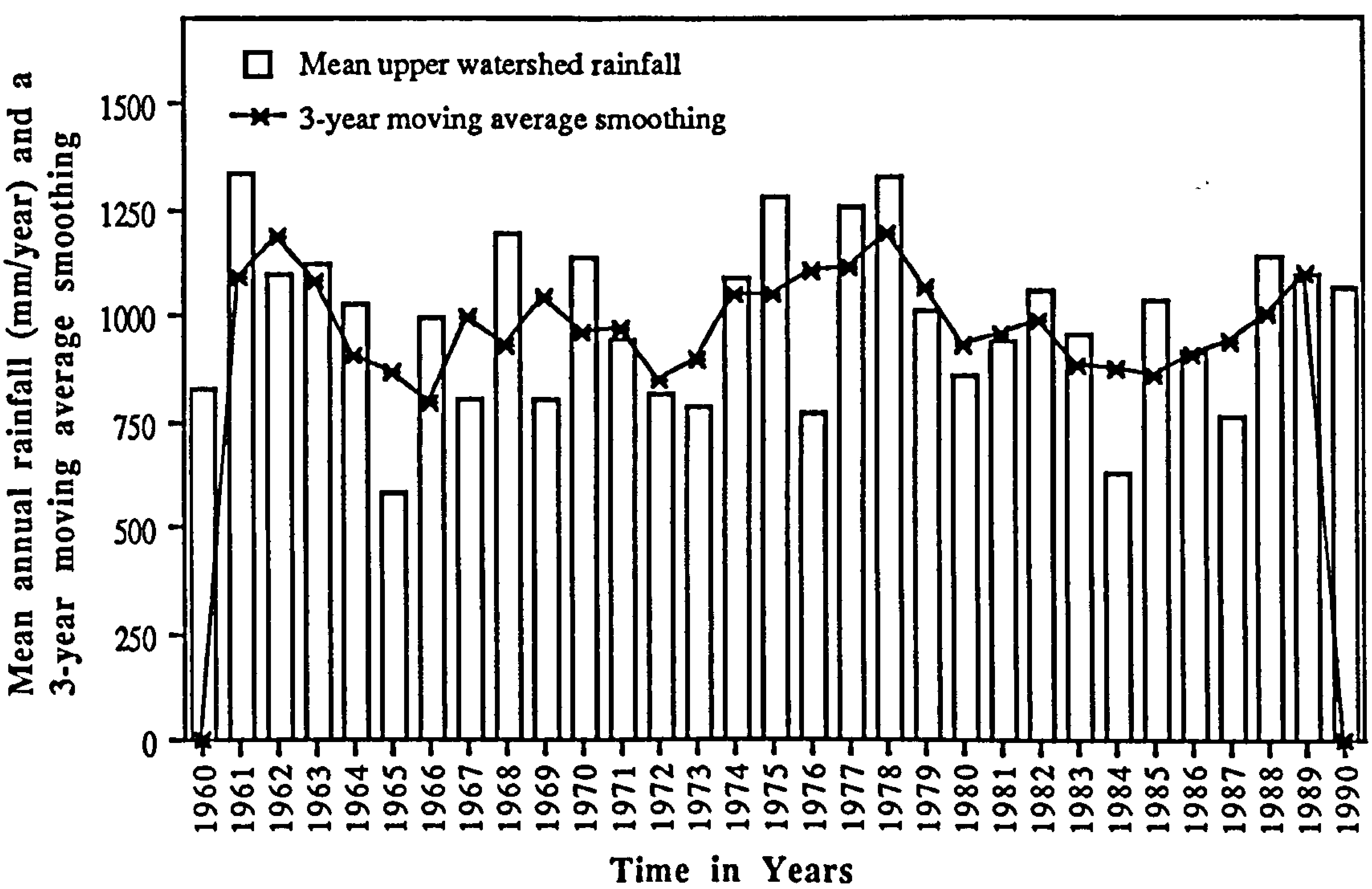


Figure 5.9. Variations of the annual rainfall and a three-year moving averages in upper reaches of the watershed.

5.3.2. Mass Curve Analysis of the Rainfall Data

The running sum of the annual rainfall is one of the homogeneity tests used to detect existence of trends in hydrologic data series. Sometimes referred to as the cumulative sums, they are good indicators of detecting temporal change in the rainfall regime. The approach has also been used to detect non-homogeneity of rainfall data during a particular time period (Chapter IV). It involves plotting accumulating totals of the annual rainfall against other series for selected periods for a particular location. In this way, it is possible to detect changes in their slopes. For a mass curve of the parameter versus time:

$$\sum_{t=0}^{t} R_i \text{ vs. time, where } R_i \text{ is the value of the parameter at time, } t \qquad (5.7)$$

Figure 5.10 presents the mass curve of the mean annual rainfall at the higher and lower elevation parts of the watershed. Rainfall at the lower elevation areas is lower than at the higher elevation stations. The lack of a complete straight line that intersects at y =0 in both the series however, indicates an existence of trends or breaks whose cause and location should be established using a double mass curve.

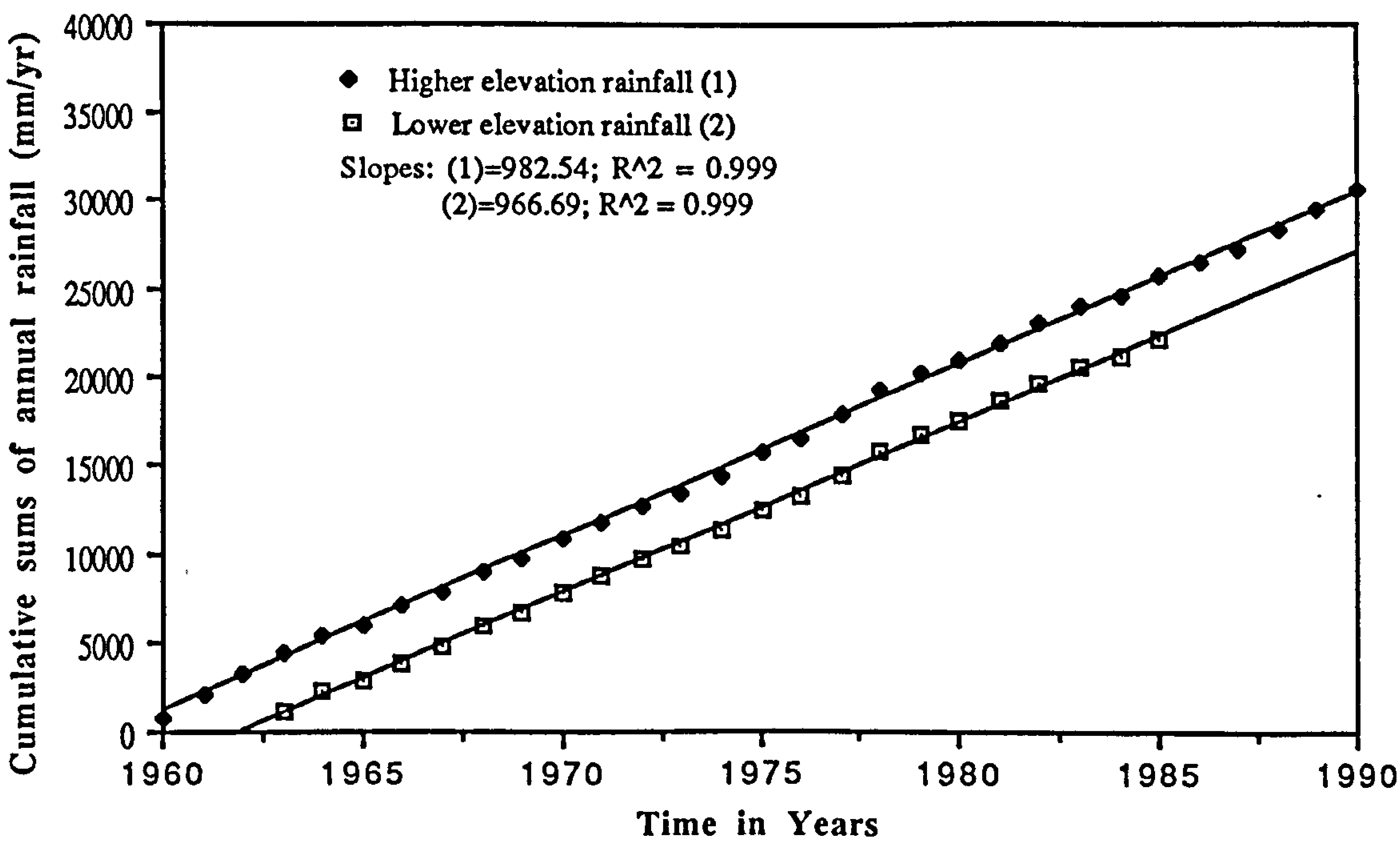


Figure 5.10. A mass plot of the annual rainfall in the upper and lower parts of the watershed.

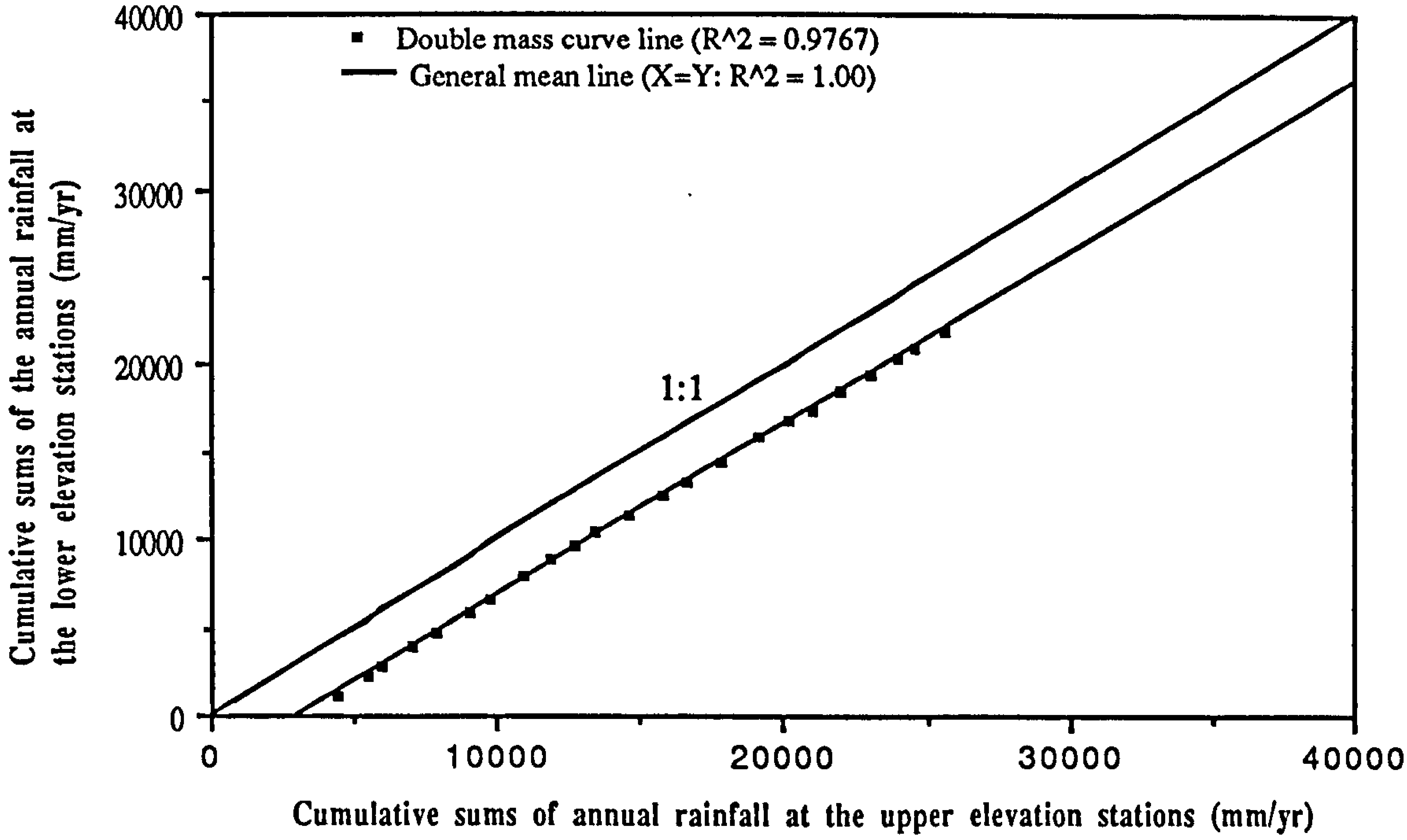


Figure 5.11. Double mass curve of the annual rainfall at the upper and lower parts of the watershed (where $\overline{Xu}$ and $\overline{Y}$L are the cumulative sums at upper and lower reaches).

The double mass curve in Figure 5.11 gives an idea of the amplitude of the suspected trend, because it appreciates their changing slopes. The higher elevation rainfall is considered homogeneous and hence used as a control, and its general means line is used to assess the magnitude of change of the means of the other rainfall stations' values. The departure to the right, from the general line ($\overline{Xu}=\overline{Y}$L) means that the mean of the lower elevation rainfall is less than that of the higher elevation areas. A simple fit of these curves gave a slope of 0.9767 at R^2 = 1.00 compared to the normal slope of 1.00 of the general mean line, which is not a large variation.

5.3.3. Annual Mean Rainfall Residuals

The cumulative function can also be used in the same way as the double mass curve to further analyse temporal trends, by observing the cumulative totals above or below the long term mean rainfall (R_m) in form:

$$CUSUM(Rt) = \sum_{i=1}^{t=N}(R_i - \overline{R}) \qquad (5.8)$$

where, (R_t) are the cumulative sums in t-year, N is the total record in years, R_i is the ith year rainfall, and $\overline{R}$, the mean (1960-1990) rainfall in mm/yr. The CUSUM (Rt) function is plotted as a function of time and the results are assessed by comparing those values above or below the zero line (y =0). The CUSUM function rotates the mean line of the plots so that they are brought to the horizontal. The deviations from the general line are therefore much more visible as they determine the scale of the graph in opposition to the mean line as in the double mass curve (Pirt, 1983, and Cluis, 1989).

Four situations are often searched for when using the CUSUM function: (1) If the curve intersects the y =0 line very often, then there is probably no significant trend; (2) if there are departures on only one side of the curve, it indicates a probability of a trend; (3) if the curve is parabolic, it suggests a presence of a monotonic linear trend in the data series and; (4) if there are discontinuous lines, this indicates a presence of stepwise trend. The results of these computations are presented in Figures 5.12 and 5.13

On the basis of these conditions, the rainfall in the higher elevation part of the watershed (Figure 5.12 displays a parabolic sequence that enables the study to conclude that it has a monotonic linear trend. This trend peaks at both the positive and negative sides of the curve with an amplitude of between 375 to 450 mm every 14-15 year.

The rainfall at the lower reaches of the watershed however, display a different pattern as shown in Figure 5.13. Here, the curve assumes the first condition of departures intersecting the y =0 axis more often, suggesting that there is probably no significant trend. It also seems to have more departures but not all. The conclusions drawn here is that, there exist a hidden trend in the series attributable to the unusually very heavy rainfall in 1978, which made the deviations to be on the positive side during the 1980s.

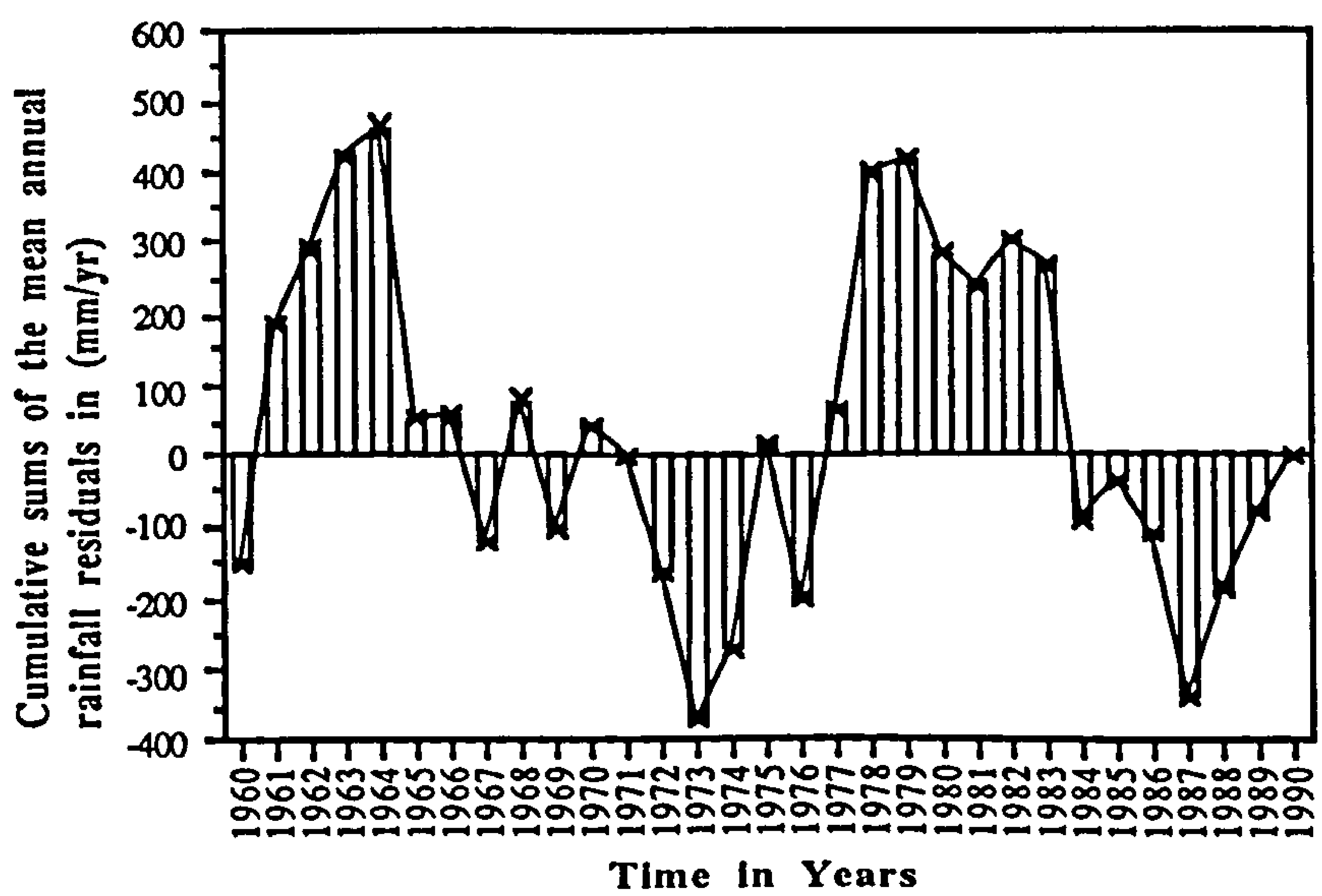


Figure 5.12. Time series of annual rainfall residuals in the upper reaches of the watershed.

5.4. SEASONAL RAINFALL VARIATIONS

The seasonal variation of the rainfall is of interest in environmental and particularly watershed studies as it affects the annual flows and hence annual concentrations of nutrients in the river system may vary considerably, due to biological activities and other conditions. The seasonality of the rainfall in the watershed is presented as regimes for the hydrologic system as well as for the higher and lower elevation reach of the watershed. Time series of the monthly rainfall in the system for 1960-1990 is presented in partial periods and in wet and dry periods, so as to isolate time periods of extreme fluctuations, variability and changes. These are then related to any observed and perceived human-induced changes in hydrological regimes.

In order to examine and compare the rainfall regimes, the monthly rainfalls (R_{mi}) expressed in percentage of the annual rainfall (R_{yi}) were calculated. The representation is occasionally referred to as the pluviosity of the rainfall (Beran et al., (1985) Egerton and NPBRS stations were used as baseline stations because of their long and consistent records. The seasonal distributions of the rainfall within the year assume different patterns as a result of the complex watershed relief and topography discussed in Chapter IV. The seasonal rainfall peaks in April, August, and November. A gradual downward trend starts at the end of November and ends in the mid of March.

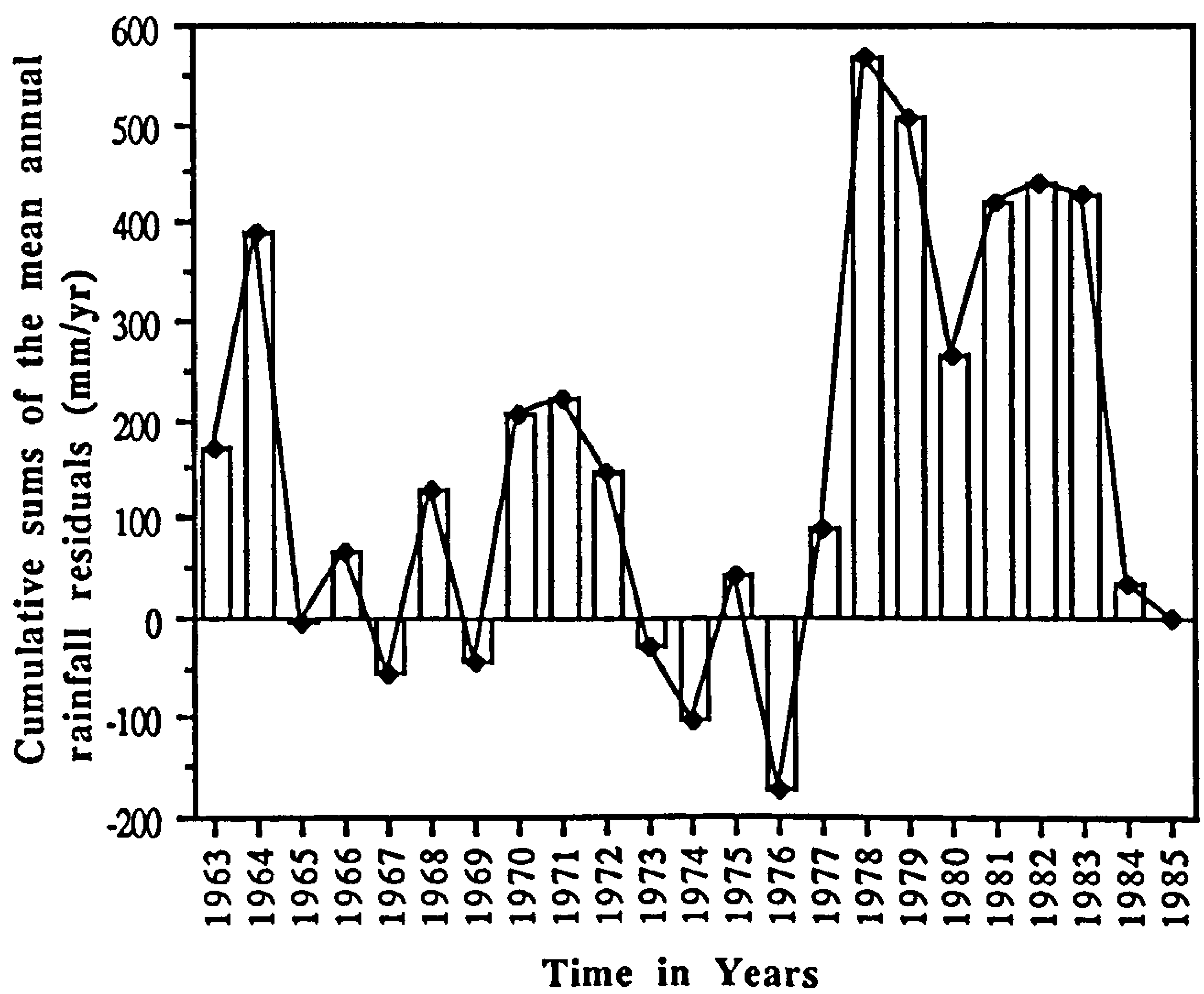


Figure 5.13. Time series of mean annual rainfall residuals in the lower reaches of the watershed.

The analyses of the monthly rainfall data for seven recording stations were grouped as before for the higher elevation and lower elevation reaches of the watershed. This would account for the rapid drop in watershed elevation and a corresponding change in vegetation cover. The stations to station variation in the monthly rainfall pluviosity (in percentage) are shown in Figures 5.14 and 5.15 respectively for the two different elevations reaches while the results for the whole study period are given in Appendix D.1.

The plots display a somewhat similar seasonal regime in both reaches, in that, the peaks in April, August, and November occur in all stations. The only difference is their magnitudes. For example, the highest seasonal pluviosity of 16 % occur in April in all the higher elevation stations, while the lower elevation stations have a 14 % maximum. The values in all cases assume decreasing patterns of April, August and November. The lowest value in the stations is experienced in January with a 2.5 %. The conclusion that can be drawn is that, the stations have similar seasonal peaks but there is more individual monthly rainfall in the upper reaches than the lower reaches when compared to their annual values.

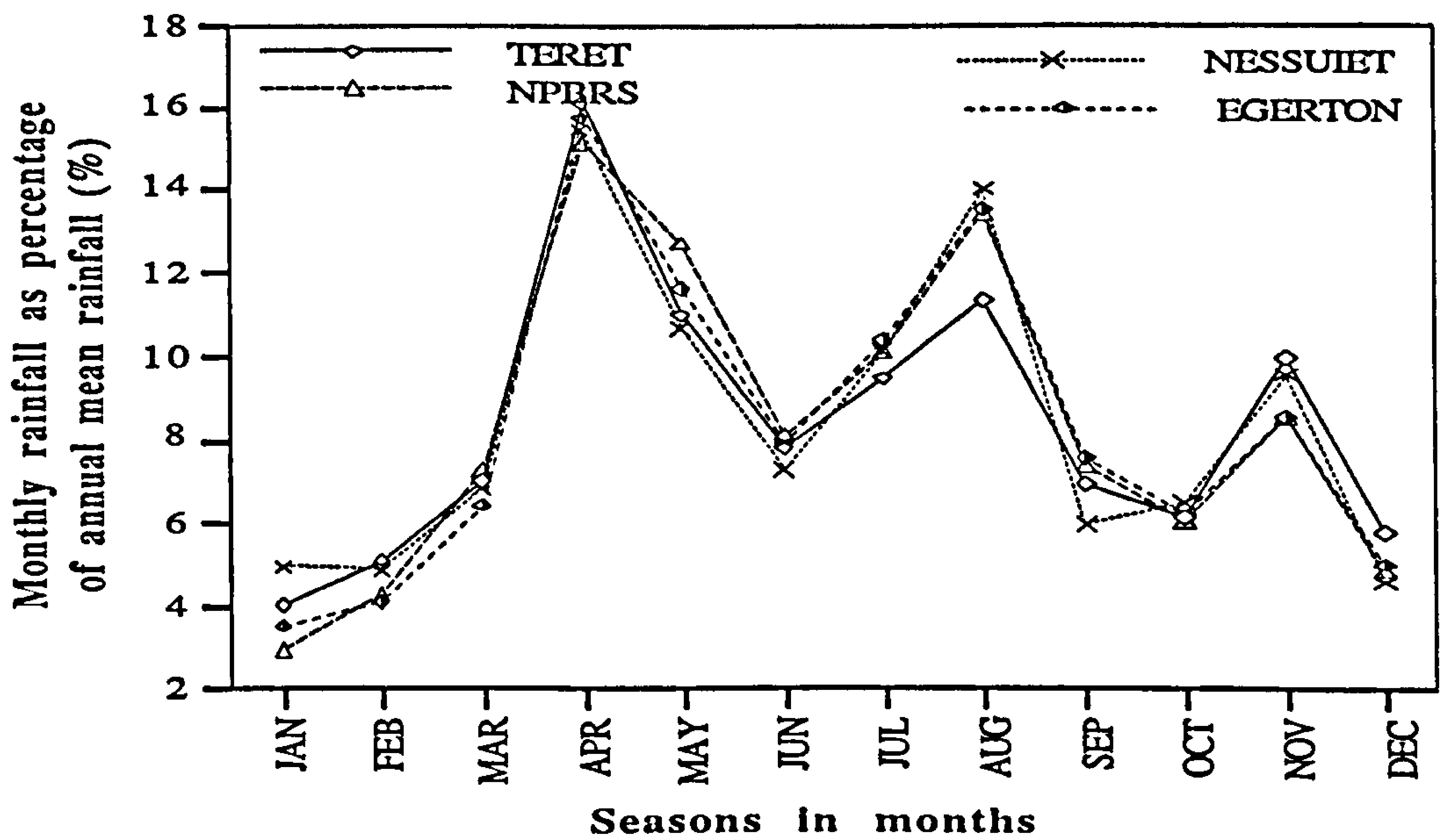


Figure 5.14. The seasonal patterns of monthly rainfall expressed as percentage of annual rainfall (R_{mi}/R_{yi} *100) in the higher elevation rainfall recording stations.

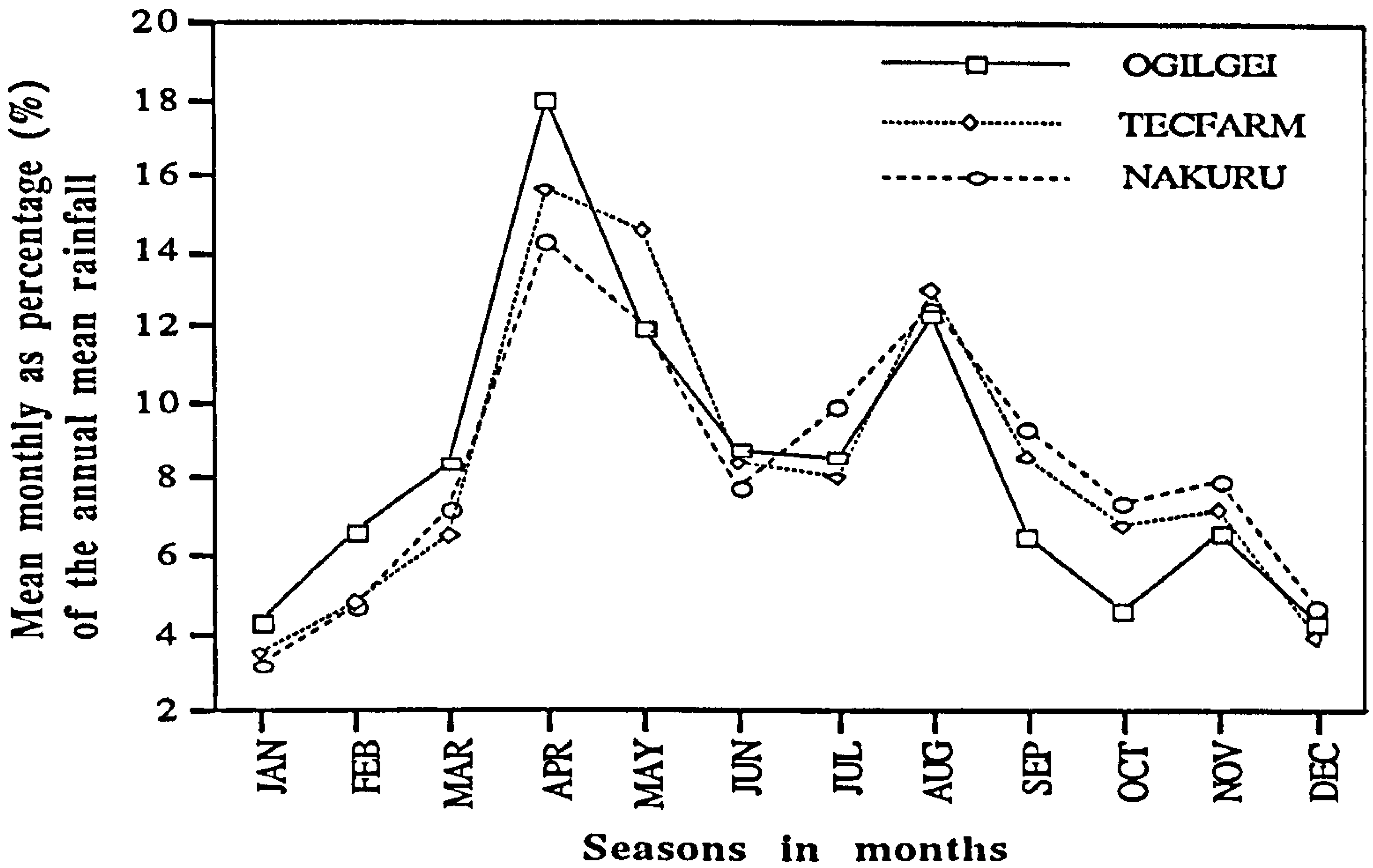


Figure 5.15. The seasonal patterns of the monthly rainfall expressed as (%) of the annual rainfall in the lower elevation rainfall recording stations.

In addition, the individual monthly rainfall values were examined as they deviated from the long-term seasonal means. The deviations describe the seasonal evolving regime in the form:

$$D_{mi} = (R_{mi} - R_m) \qquad (5.9)$$

where D_{mi} is the ^{i}th month deviation (mm), R_{mi} is the ^{i}th month mean rainfall (mm), and R_m is the (1960-1990) mean monthly rainfall (mm).

The procedure was applied as before to the respective seasonal rainfall in the higher and lower elevation stations. Results are presented in Figures 5.16 and 5.17 respectively. The deviations of the rainfall values display clearly the seasonal rainfall regime in the watershed. The peaks in April, August and November identified earlier are seen all over the watershed, but the lower reaches indicate that November value is lower than the long-term mean (1960-90 mean) value by about 40 millimetres. The plots also confirm earlier findings that January is the driest month. The decreasing trend of the rainfall approaches its lowest levels from December to March, where it is roughly 40 mm below its mean value. The peaks are conspicuous in the higher elevation reaches with a 70 millimetre positive deviation. The peak in November is however not outstanding at the lower elevation areas of the watershed.

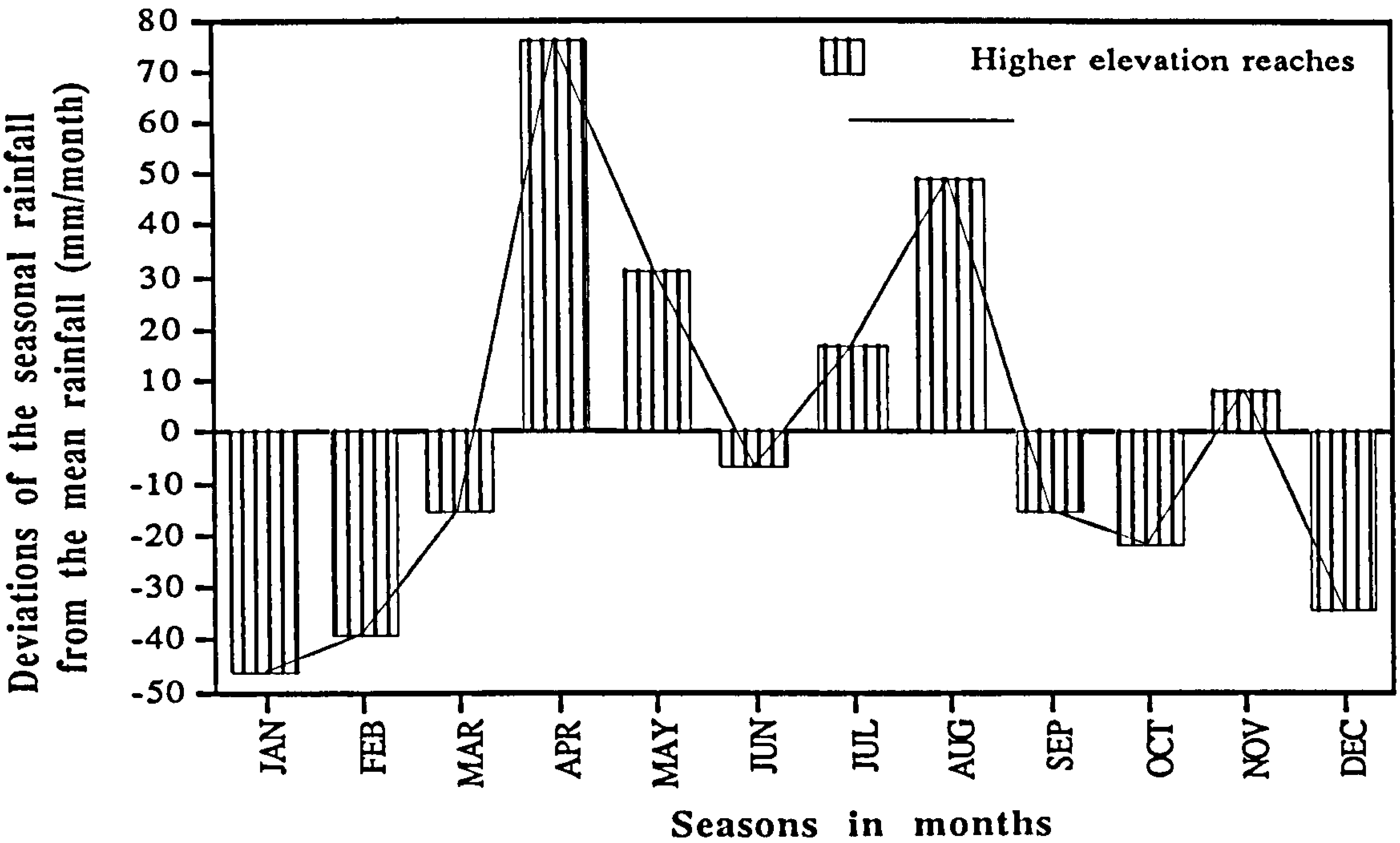


Figure 5.16. The seasonal deviation of the rainfall in the higher elevation reaches of the watershed.

Subjecting the monthly rainfall series to Z-score representation in equation (5.6), the seasonal rainfall variability is presented in Figures 5.18 for both the higher and lower elevation reaches confirmed the observed trimodal nature of the seasonal rainfall regime.

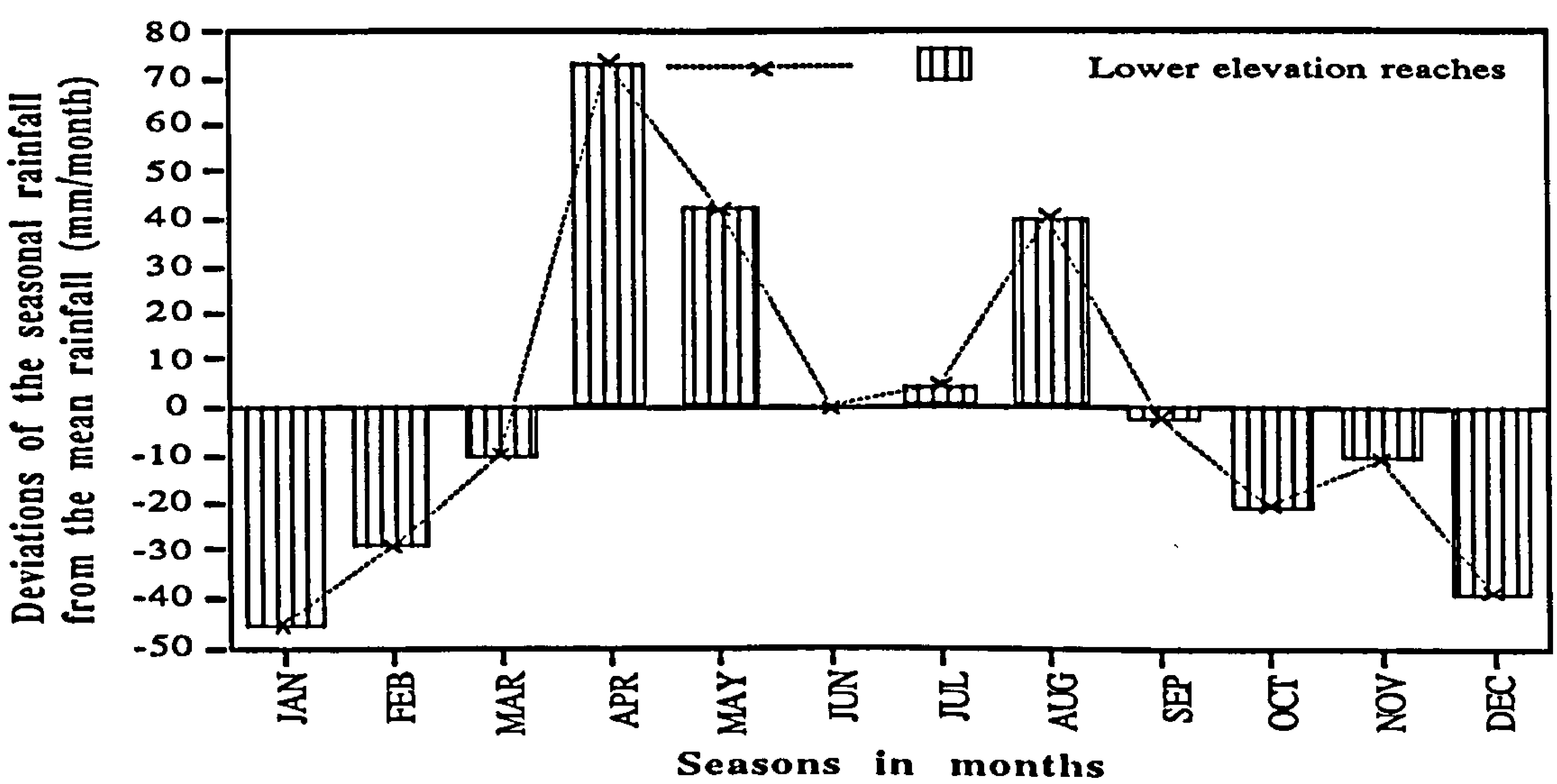


Figure 5.17. The seasonal deviation of the rainfall in the lower elevation reach of the watershed.

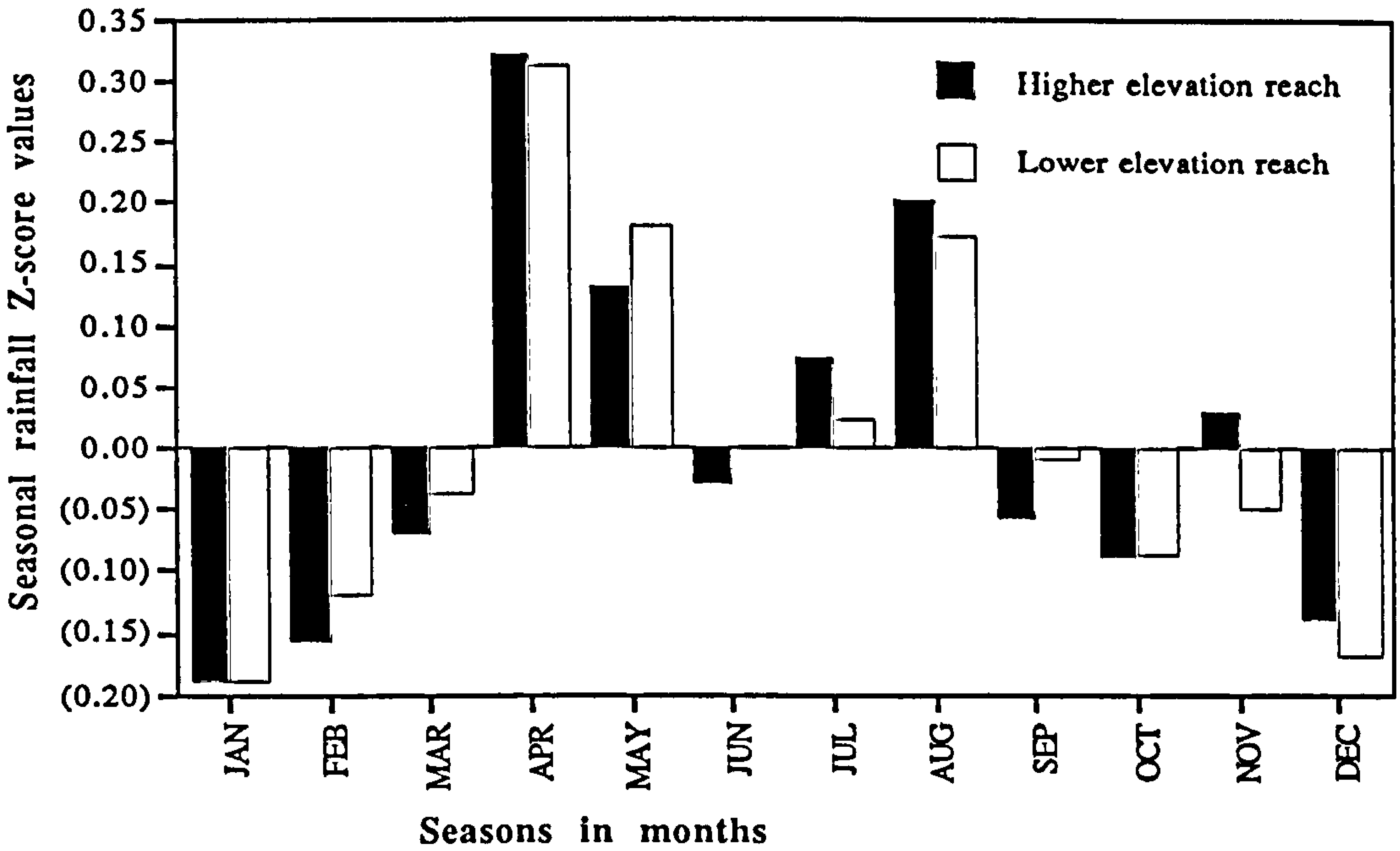


Figure 5.18. Seasonal variation of the Z-score values in the higher and lower elevation reaches of the watershed.

The cumulative sums of the mean (means from all the stations) seasonal rainfall in the upper reaches (higher elevation) was further examined on a ten-year mean monthly partial series to establish its seasonal trends. The results presented in Figure 5.19 show the rainfall patterns and trends in the 1960s and 1980s to have rather similar behaviour and magnitudes. The 1960s, however, stands out as the period with the highest seasonal mean rainfall. From this stage henceforth the analysis of the rainfall in the lower elevation stations is eliminated in order to concentrate in the upper elevation rainfall where SWSI and SWSII are located.

5.5. TEMPORAL AND SPATIAL VARIATIONS OF STREAMFLOWS

Streamflow in a river is the integration of many different tributaries flows, each of which is dependent on the climate and physiography of its drainage system (Higgs, 1990). The variability of this flow is established from several indices. The most common index is expressing each annual flow as the depth on its watershed area. This is referred to as the watershed specific discharge (m^3s^{-1} per km^2 or l/s per km^2). It is equivalent to the yield per unit area and is commensurate with its causal rainfall characteristics.

The tropical region of sub-saharan Africa produces intensive rainstorms that are highly variable. Coupled with extremes of temperatures and rapidly changing topography along the equator, these produce extreme temporal and spatial variation in streamflows. In addition, various combinations of climate, geology and change in land use combine to produce large diurnal variations in river flows.

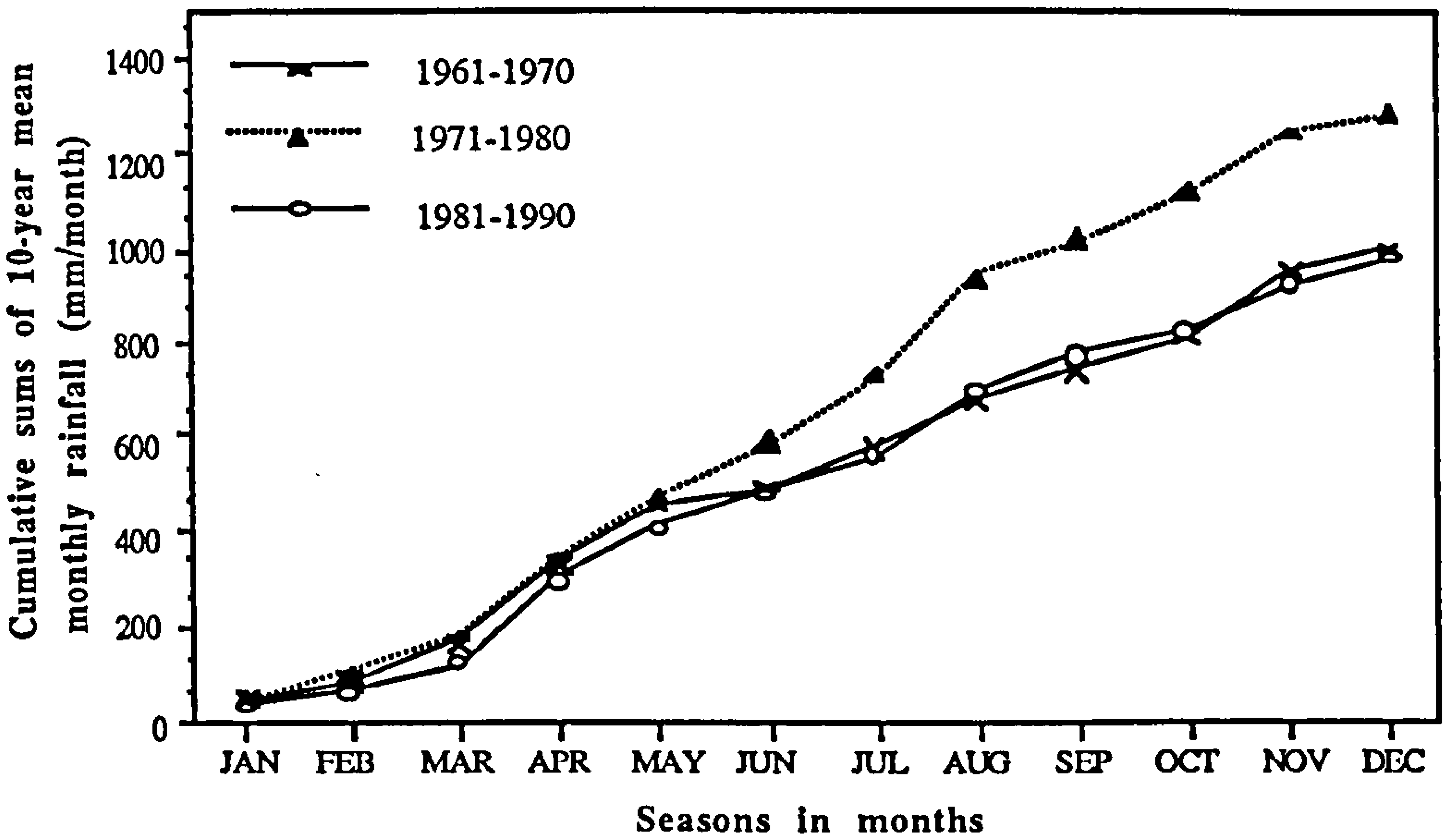


Figure 5.19. Trends of the rainfall in the higher elevation parts of the watershed.

In order to understand, the behavioural patterns of these variations in rainfall and streamflows, several hydrological indices were examined both in temporal and spatial dimension. The procedures of expressing the ith month flow to annual mean flow, percentiles, running sums, Z-score values, and residual analyses presented in equations (5.1- 5.8) were also used to evaluate the flow data.

5.5.1. Variations in the Annual Streamflows

The basic statistical description of the streamflows is discussed here. The nature and magnitude of streamflow variation, and the factors underlying this variation are explained. An attempt is made to explain the major climatic and physiographic factors underlying the respective variation by examining their basic descriptors, year-to-year variation, hydrologic indices, mass curves, residuals and statistical distributions.

5.5.1.1. The Basic Flow Descriptors

The use of statistical terms to describe hydrological variables is adopted. The mean flow describes the average flow for all complete years of record, and is expressed as a rate of discharge (m^3s^{-1}). The mean annual flow or sometimes means of annual flow are considered equal to the mean flow when expressed as a rate of discharge. Alternatively, the mean annual flow is expressed as the depth of water over the watershed area (mm/yr). The annual mean flow on the other hand is the mean flow for a particular year, and the monthly mean flow is the mean flow for a particular calendar month of the year. It is the mean of all recorded monthly mean flows for a particular month.

Using these terms, the statistical characteristics of the streamflows were evaluated. The statistical descriptors for the flow series; 2FC05, and 2FC09 of $\overline{Q}$, C_v, C_s, and C_{k-3} were calculated using equations (5.1 to 5.4). The two flow series were chosen because of their consistent and reliable data. They are the integrated outputs from SWSI and II located at the higher elevation reaches of the Enjoro river watershed.

The coefficient of C_v, C_s, C_k-3 values were plotted against the total annual flows. A statistical measure of the annual mean flow variation is provided by the coefficient of variation (Cv) whose interpretation has been discussed in section 5.2. The coefficients were derived from the 2FC05 and 2FC09 flow series and plotted against the annual mean flow in Figures 5.20 (a-c) and 5.21(a-c) respectively. The C_v value in Figure 5.20 concentrates between 0.6 and 1.8 for flow rates up to 0.6 $m3s^{-1}$ and reaches a maximum of 2.4 when the discharge increases to 1.6 $m3s^{-1}$. The C_s and C_k-3 displays a pattern of increases and decreases with no defined trend. From Figure 5.21, there Cv decreases with increasing discharge. The Cs and C_k-3 show not a well defined trend with respect to the mean annual flow. Since these basic statistical descriptors display no definite pattern, there is a need for further analysis.

5.5.1.2. Year-to-Year Streamflow Variation

An attempt is made to describe quantitatively the annual flow variations using various hydrological indices and graphical representations. Such an approach enables an effective detection of periodic and human-induced trends in the flow series. The natural variability of the rainfall from year-to-year was considered a major controlling factor in addition to the perceived human-induced factors. Despite its small size, Enjoro river watershed hydrological parameters seemed to have a high degree of variability in its time series. This is illustrated by the 5-year moving averages shown in Figures 5.22 and 5.23 for SWSI and II respectively. The smoothing clearly depicts a decreasing trend of the annual flows from the 1960s to the 1980s. A low flow pattern exists in a 10 to 12-year periodicity from 1965/66, 1973/76, and 1984/88. Altogether, the annual flow series, assume a year-to-year variation at each subwatershed in response to climatic changes, particularly to rainfall and watershed physical changes.

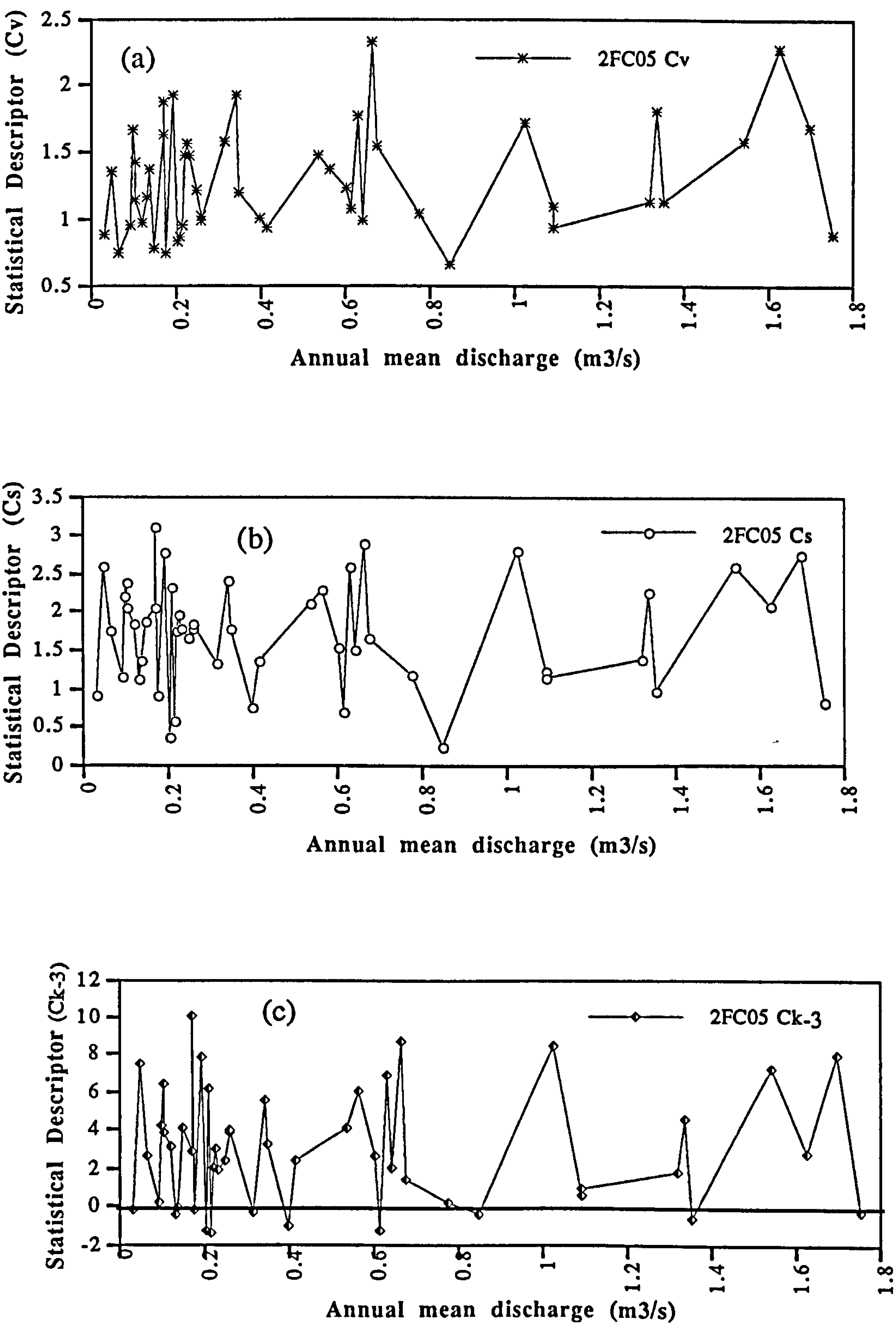


Figure 5.20(a-c) The basic descriptors of Cv, Cs, and C_k-3 against annual mean flow (2FC05 series) for the period 1960-1990.

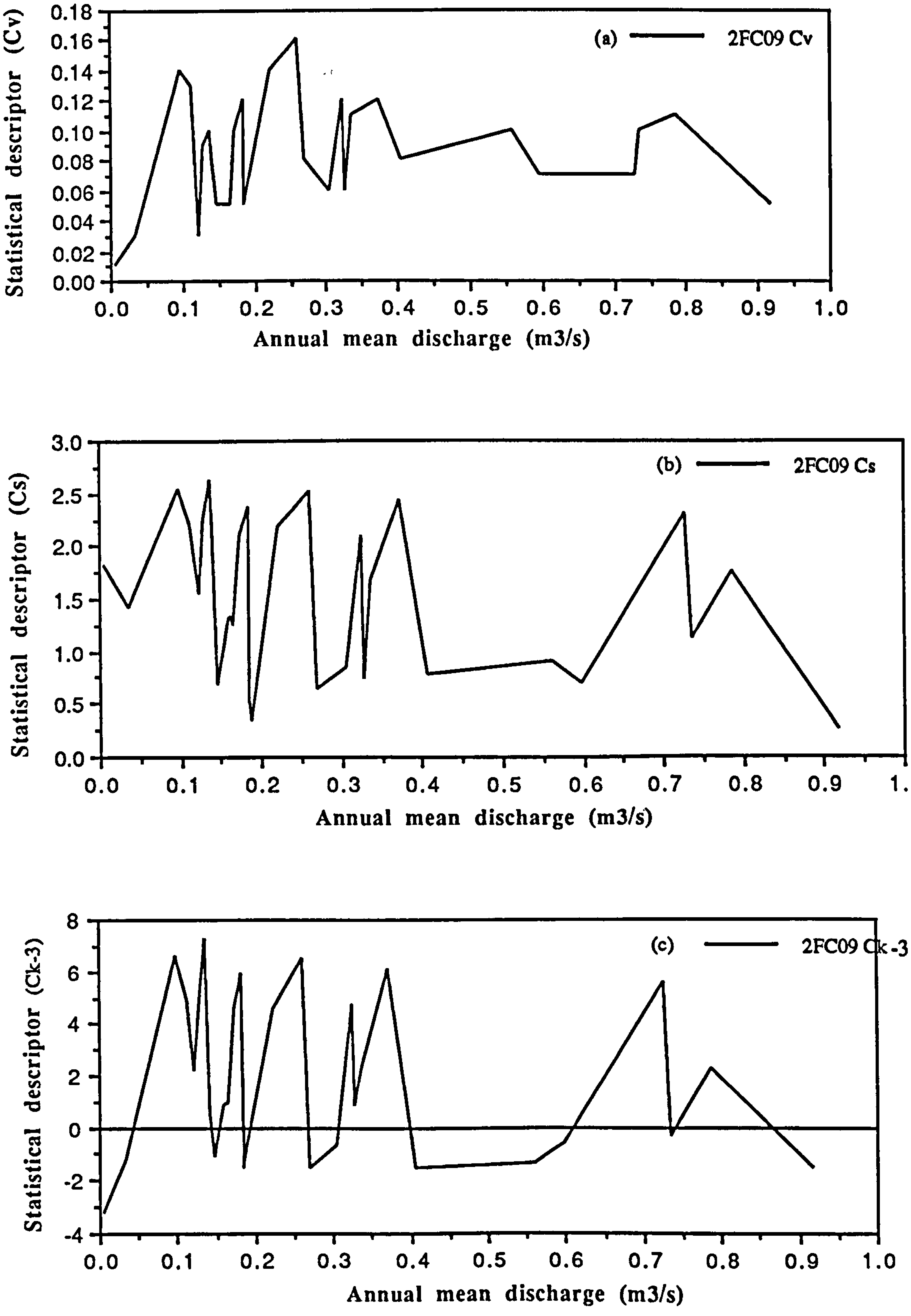


Figure 5.21(a-c) The basic flow descriptors against the annual mean flow (2FC09 series) between 1960 and 1990.

5.5.1.3. Annual Flow Variation Indices

Hydrological indices are often the observed indicators of selected hydrological components compared to standard values established for each component. It expresses the variability of the observed component in relation to its long-term mean value and indirectly to humans and their environment (UNESCO, 1985). The focus of this study is the variability over the month, year, and 10-year periods.

The development of these indices produced information on the hydrological and ecological system in a simple form that the evaluation of the impacts of human activities can be achieved without having to resort to very expensive controlled watershed experimentation. The validity of using the indices to evaluate these impacts however, is based upon the accuracy and quality of the available data. The development and uses of simple hydrological indices therefore have a potentiality in identifying changes due to activities in simpler terms. The use of several indices in this work was adopted because of a limited and inconsistent data set.

From the preceding discussion, the hydrological indices in common usage are those that relate the periodic series to the long-term mean values. The ^{i}th year flow (Q_i) for example was related to its long-term mean ($\overline{Q}$) in the relationship $(Ki = \frac{Qi}{\overline{Q}})$ and plotted against time, to give the relative runoff factors of the series. Secondly, changes of residuals of the annual flow $(Qi - \overline{Q})$ against time were examined.

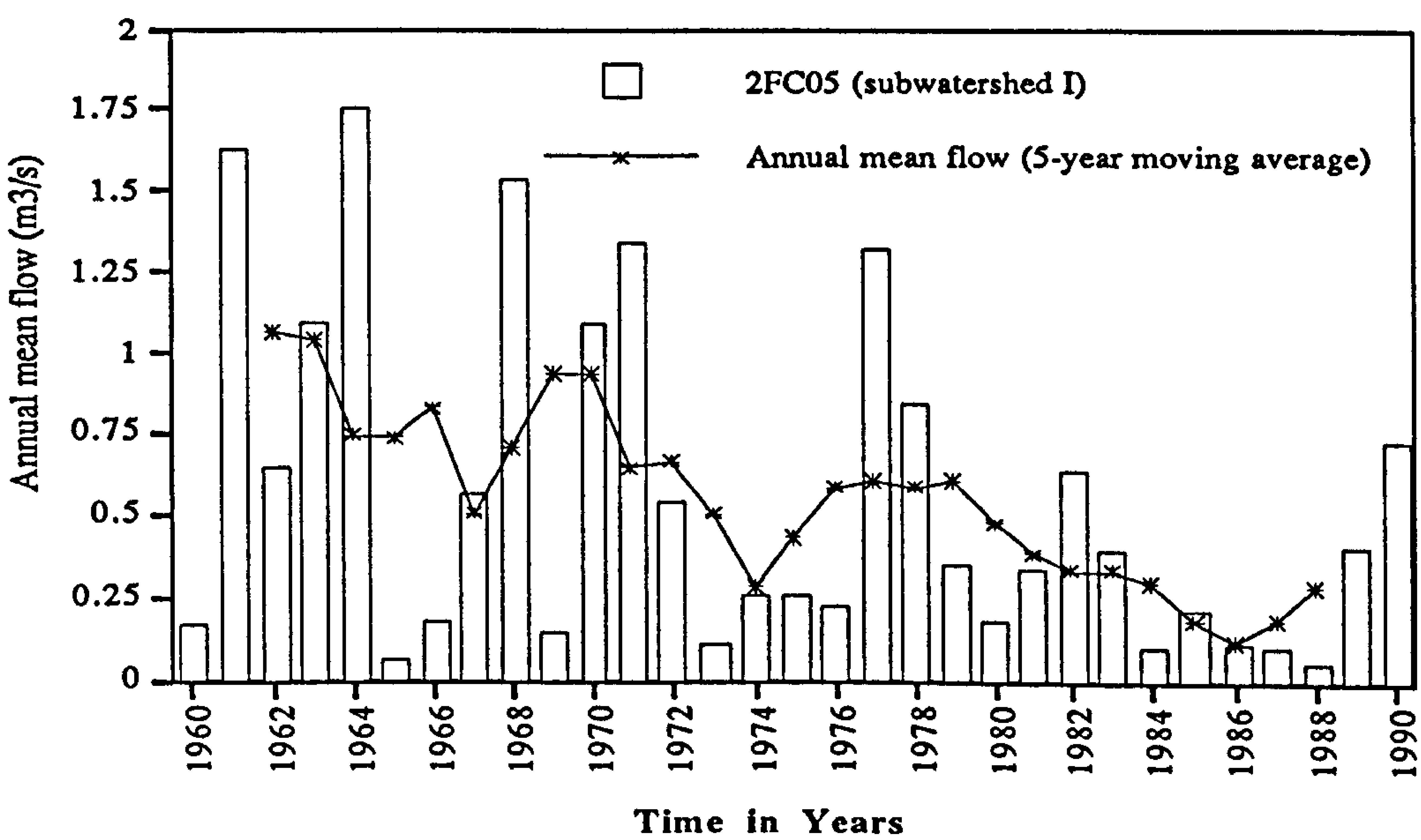


Figure 5.22. Annual time series and a 5-year moving average of the 2FC05 flow series in subwatershed I.

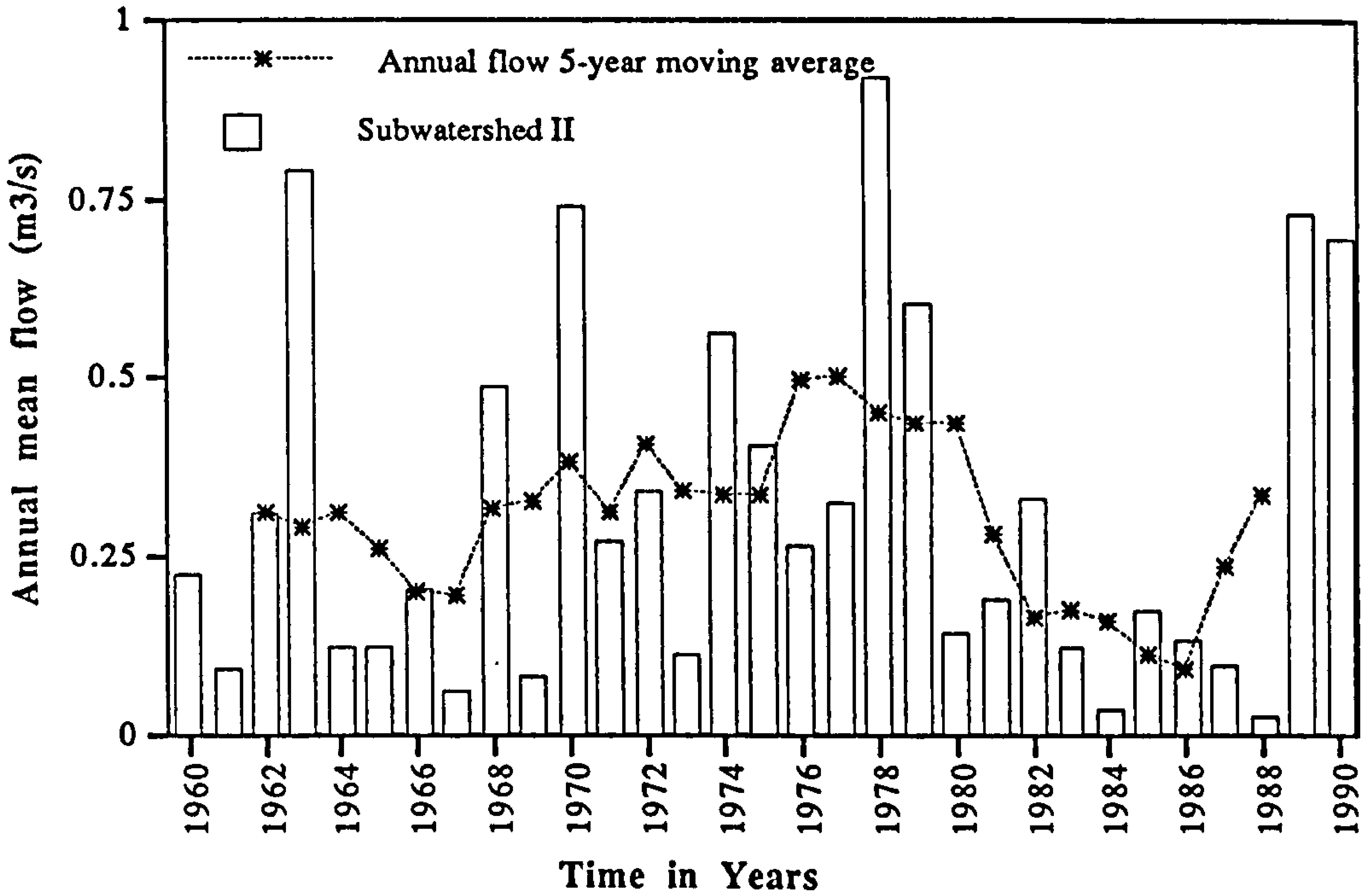


Figure 5.23. Annual time series and a 5-year moving averages of the 2FC09 flow series in subwatershed II

Thirdly, the residuals were examined by normalising the mean flow (Q_i-$\overline{Q}/\overline{Q}$) to give impression of the whole data population. Lastly, the data series were normalised to obtain the Z-score $(Qi-\overline{Q})/_{s}$ values that indicate flow variations from its norm. For all these indices, Q_i is the ith year mean flows (m^3s^{-1}), $\overline{Q}$ is the long-term annual mean flow (m^3s^{-1}), and s is the mean annual flow standard deviation. The time series plots of the indices from the 2FC05 and 2FC09 annual flows are given in Figures 5.24 to 5.27.

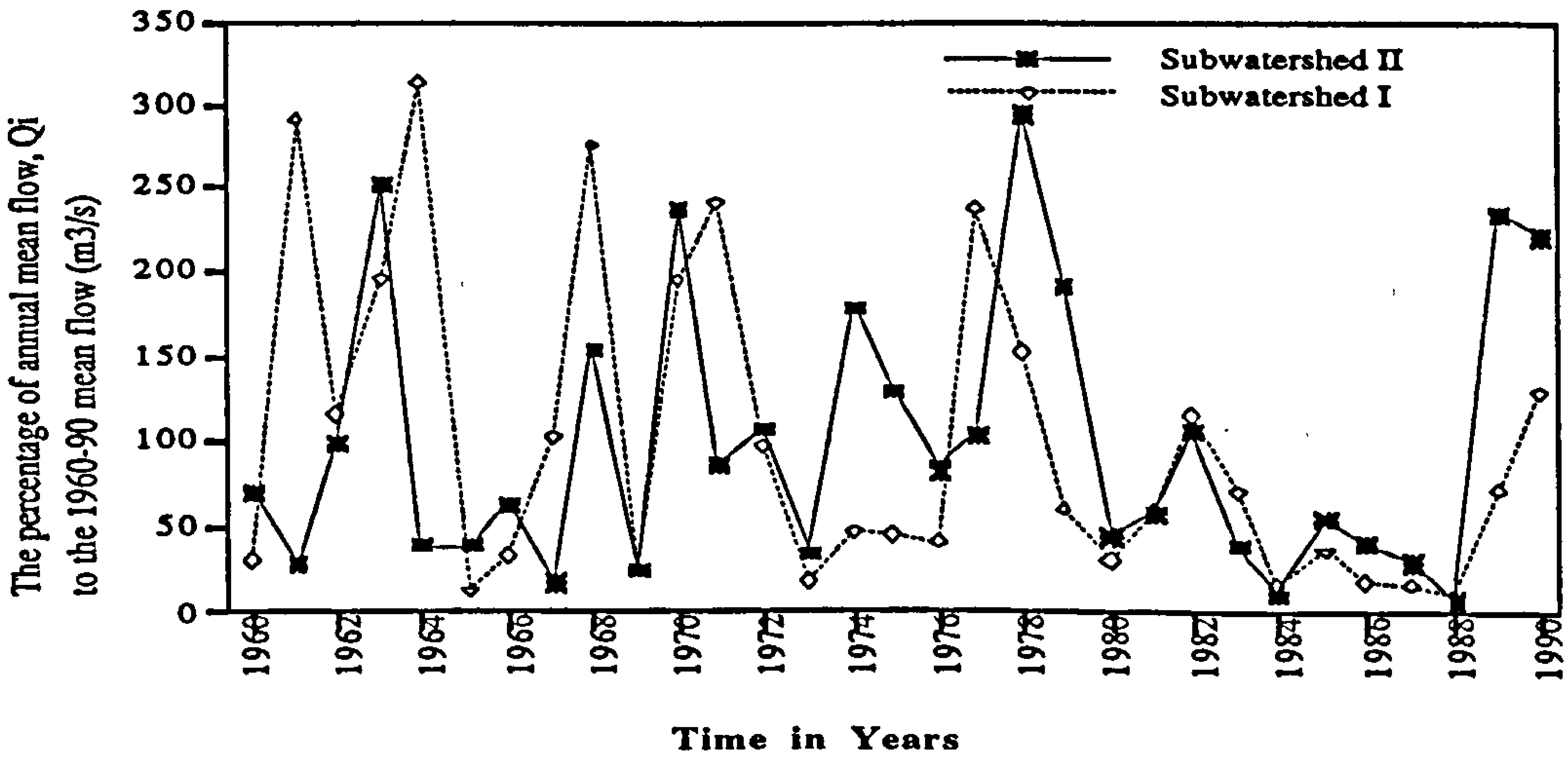


Figure 5.24. The percentage of annual flow to the mean flow (Qi/$\overline{Q}$)) in the series.

Figure 5.24 depict a similar pattern of rises and falls in the percentage of the mean annual flow to the 1960-1990 mean flow. Three peaks of more than 250 % are observed in SWSI during the 1960s while SWSII had only two in that range. A third peak falls between 200 and 230%. Both subwatersheds responded to the 1965/66 drought with a low percentage of flows SWSI seemed to experience the most effect.

The 1970s experienced higher flows than its mean value in SWSII with 3 years (1971, 1974, 1979) having more than 150%. The SWSI flow series had only two years (1971, 1978) with similar peak values. The 1980s on the other hand can be considered a decade with the lowest flow in both subwatersheds. Only in 1982 and 1989 did the flow exceed 50% of the mean flow while 1984, and 1988 had the lowest recorded flow values in 30 years of about 50%. Overall, the ^{i}th year values were closer to the mean flow. An examination of the annual flow deviations from its long-term value given in Figure 5.25 should provide a better understanding of the flow regimes in the watershed.

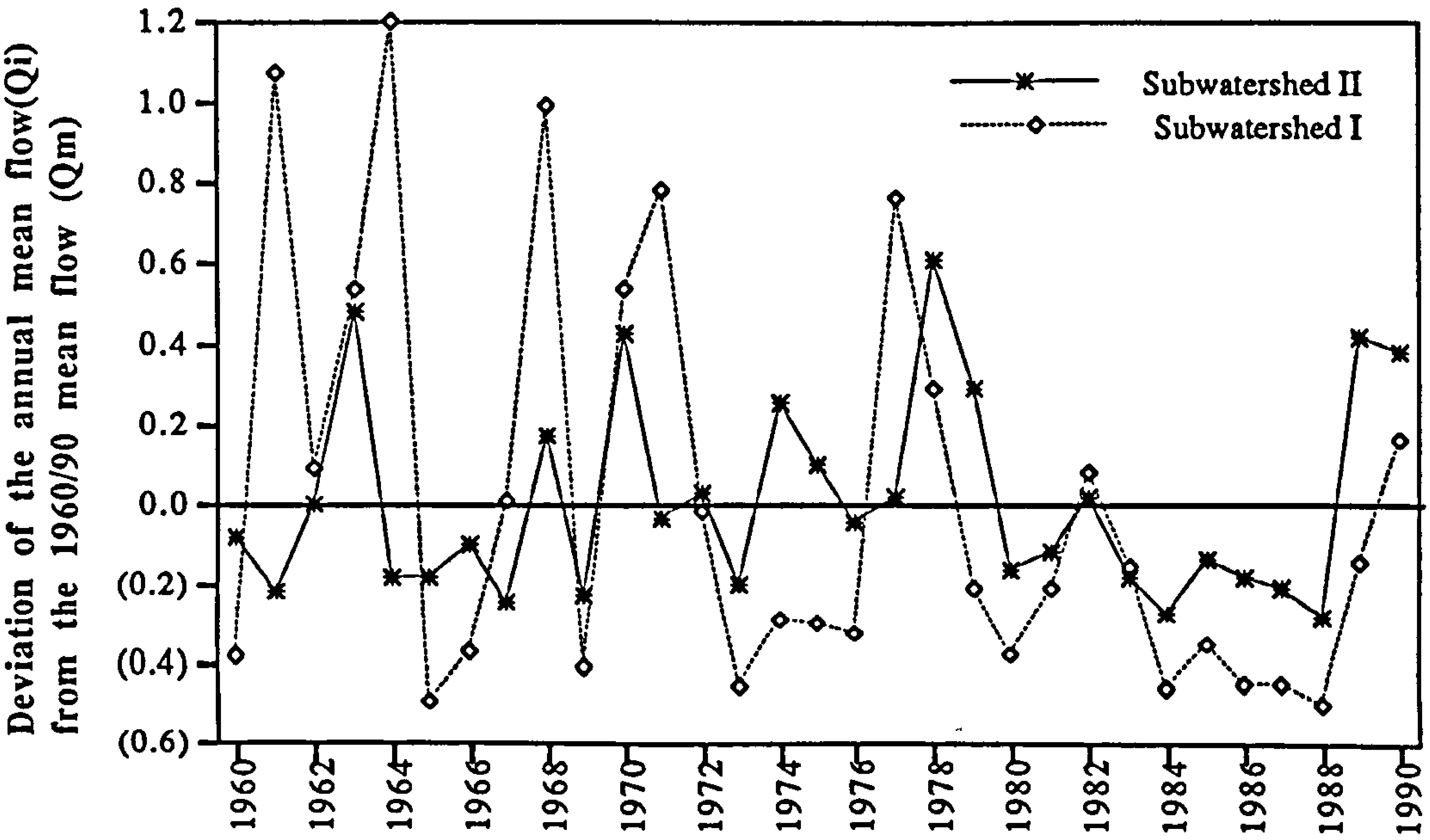


Figure 5.25. The deviations of the annual flow from its long-term mean

On examining the deviations, a clearer picture emerges. The flow series in SWSI still dominated the larger flow values. The deviations ranged from a maximum of 1 to 1.2 $m3s^{-1}$ in 1962, 1964, 1968, and 1978 to the lowest (negative deviations) of -0.5 $m3s^{-1}$ in 1965, 1973, 1980, 1984, 1986 and 1988. The flow series in SWSII seemed stable, or did not deviate as much from its long-term mean, in effect showing less deviations to the negative. The lowest deviation in the subwatershed is between -0.2 and 0 (no change), even in the driest periods of the 1980s.

The extreme positive deviations occurred only in 1963, 1968, 1971, 1975, 1978 and 1989 were about 0.4 $m3s^{-1}$. Together, the clearest picture of an existence of a hydrologic drought is between 1972 and 1975 and between 1982 and 1988, while the period with the highest flows dominated most years in 1960s, 1971 and 1978/79 in the 1970s. From 1979, until 1988, the decade experienced a hydrologic drought not previously recorded in the history of the river.

To confirm these observations, the flows were annualised by dividing the deviations by the long-term mean $(Qi-\overline{Q})/\overline{Q}$ so as to discover the actual temporal and spatial flow regime. This index defined clearly the periods of below and above average flow as shown in Figure 5.26. Both subwatersheds experienced below average situation in 1960, 1965/66, 197376 and literally the entire period of the 1980s except in 1989. The subwatersheds also depicted above average for similar lengths of time although at different magnitudes.

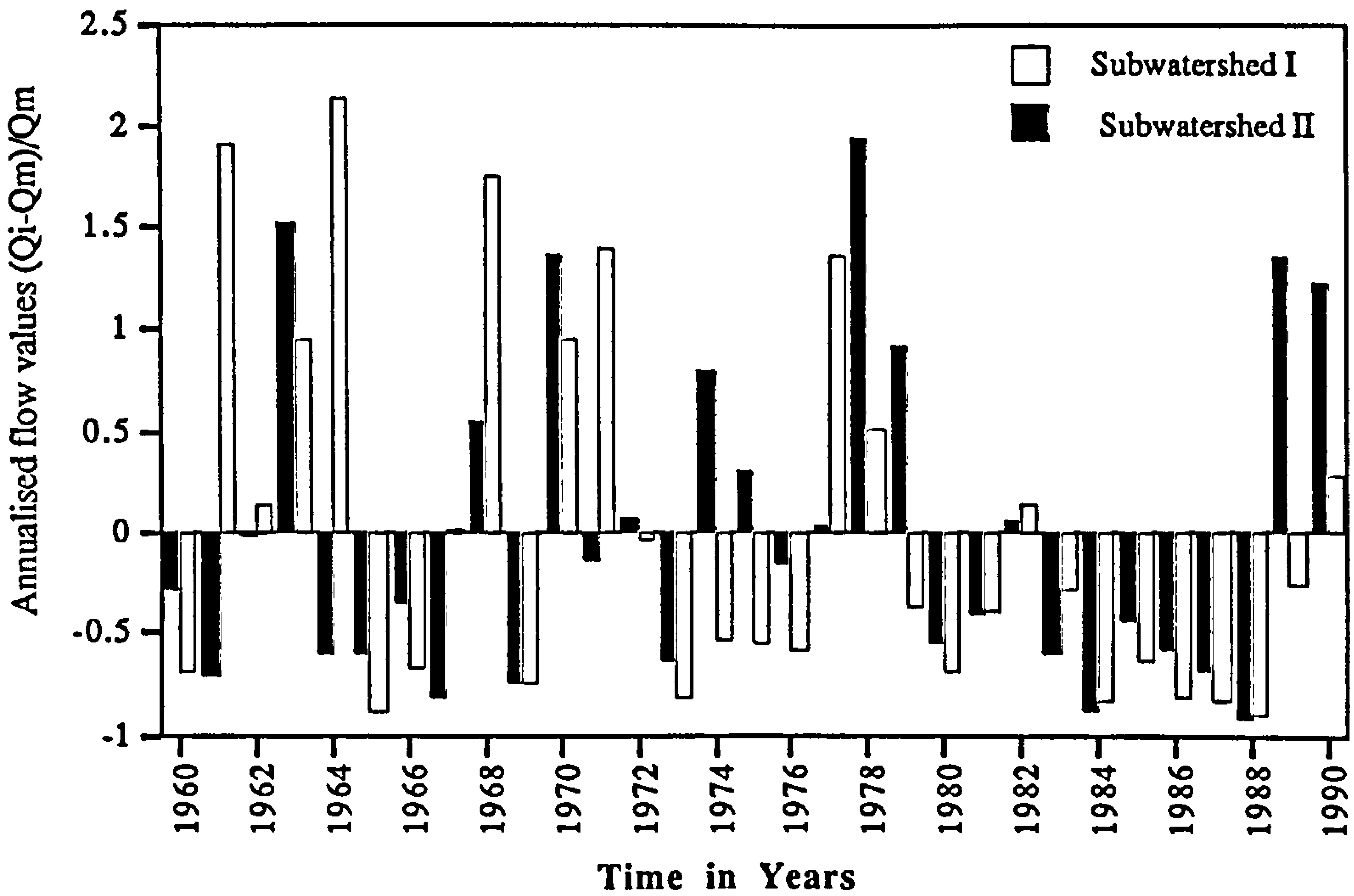


Figure 5.26. Time series of the normalised variations $(Qi-\overline{Q})/\overline{Q}$ of the two data series.

Between 1960 and 1970, three identifiable peak years can be identified in 1961, 1964, and 1968 from SWSI flow series. The flow series from SWSII had insignificant increases during the 1960s. During the 1971-1980 period, higher flow pattern in SWSII dominates particularly in 1971 and 1978 (although unstable increases can be deduced), with SWSI showing increased flow but at decreasing trends during the same period. The 1980s, again comes out as the period with the least flow with the index approaching a value of -1 in 1984, the lowest since 1965.

The flow series were further examined with the Z-score index that also normalises the flow shown in Figure 5.27. The Z-score index gives a better picture of the flow regime, and as seen, it is clearly established, that SWSI dominates the above average values especially in the 1960s and 1970s. It is interesting that the Z-score values in SWSII are relatively low even in the 1970s, when the other indices had shown larger values. There is however, a well-defined hydrologic drought that occurred among 1964-1966, 1972-1976, and the entire 1980s except in 1989. This again confirms earlier observations. The findings also correspond to the rainfall patterns, although at different magnitudes and rate of change.

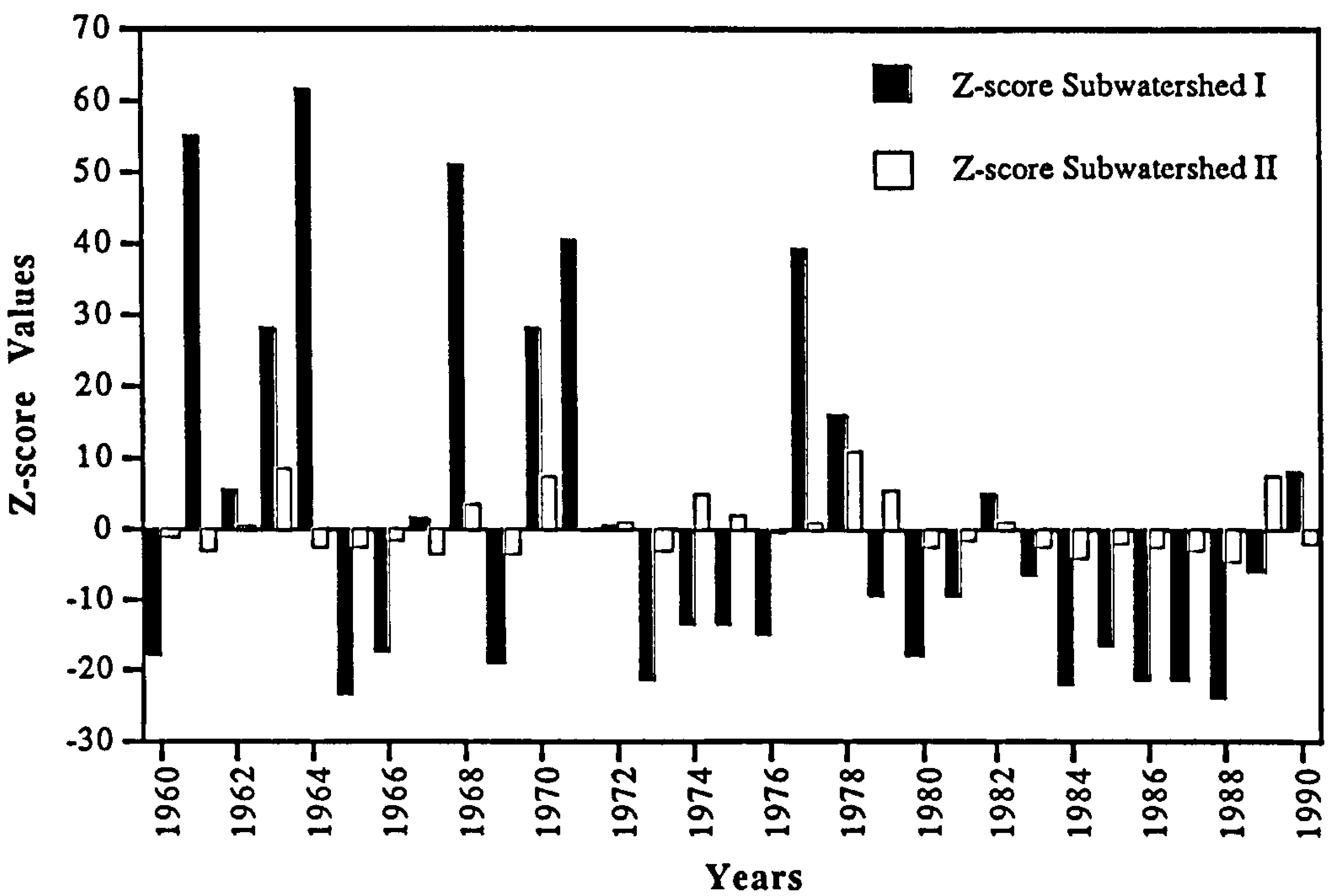


Figure 5.27. Normalized parameters of the flow series at the two subwatersheds

5.5.1.4. Double Mass Curves of the Annual Flows

To further examine the flow series, and confirm the observation in these analyses, a mass curve representing the cumulative sums of the annual flow volumes (m^3/yr) was plotted against time. The purpose was to detect the time, location and magnitudes in which changes of the flow took place since the double mass curve between the two series earlier had indicated a possible existence of a hidden trend (Chapter IV).

To isolate the observed trend, mass curves for three periods (1960-1970), (1971-1980), and (1981-1990) were further examined in Figure 5.28. The plots show more evidently that the flows in SWSI are higher on cumulative terms when compared to that of SWSII series. The arrows identify those periods during which the flow regime changed from high to low or vice visa. These occurred in approximately 1965, 1973, 1980 and essentially the entire years after 1983 when the total volumes in each case was below its respective best fit line.

A detail subperiodic series analysis was carried out by quantifying the rate of change of the mass curve slopes. The approximate value of the slopes at the different time periods is provided in Table 5.3.

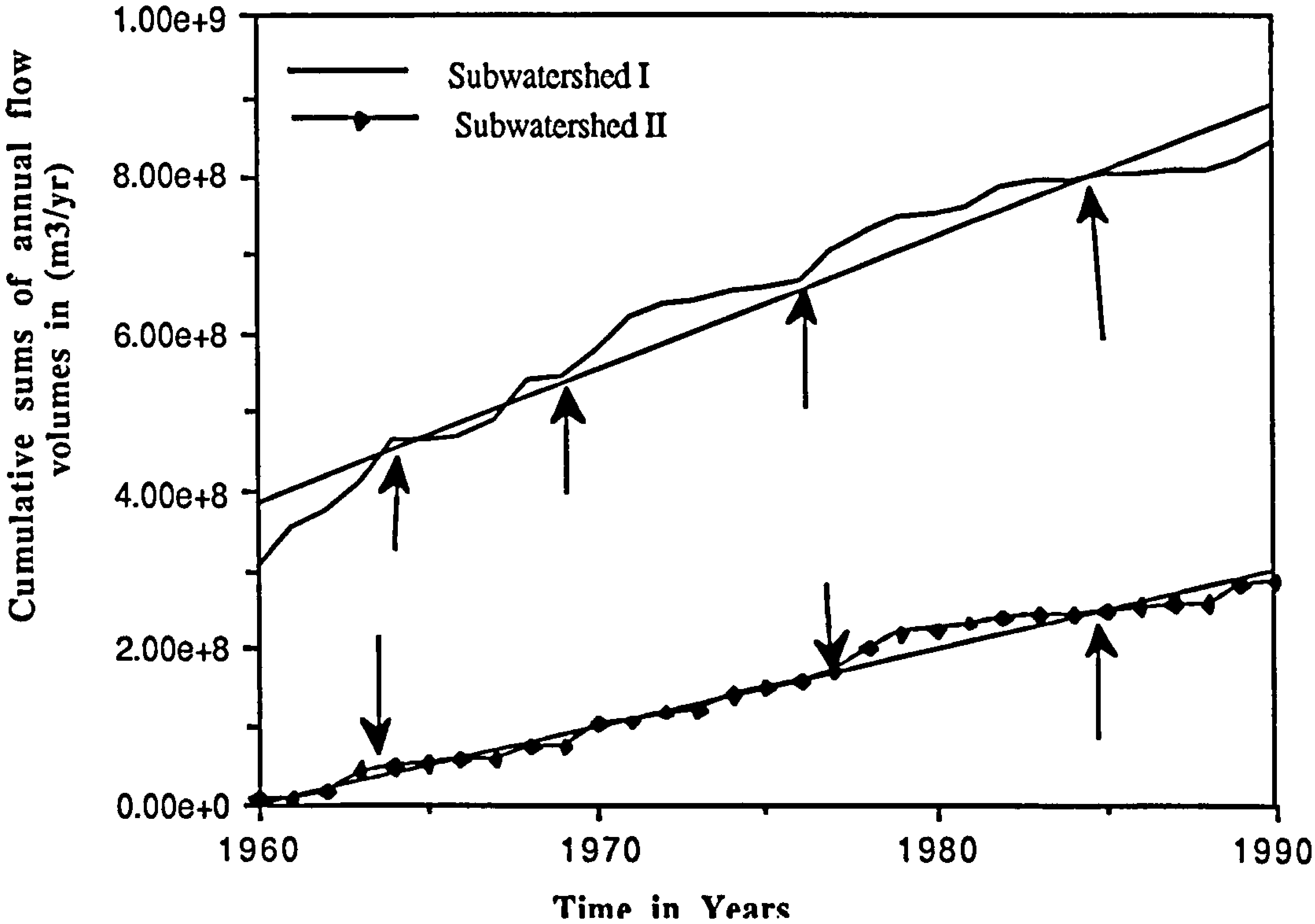


Figure 5.28. Time series of the annual volumes in the subwatersheds (1960/90)

Table 5.2. Slopes of the mass curve of the annual flows in SWS I and II

PARTIAL PERIODS	2FC05 FLOW SERIES(10^7)	2FC09 FLOW SERIES(10^7)
1960 - 1970	2.532900 $R^2 = 0.965$	0.86825 $R^2 = 0.955$
1971 - 1980	1.537580 $R^2 = 0.934$	1.353917 $R^2 = 0.974$
1981 - 1990	0.718917 $R^2 = 0.904$	0.543583 $R^2 = 0.903$
1960 - 1990	1.720210 $R^2 = 0.953$	0.99500 $R^2 = 0.982$

Changes in the slopes of the mass plots describe the changing behaviour of the river regime. From the tabulation above, the 1960s had steeper slopes in the 2FC05 series, and then begin to fall in the 1970s reaching its lowest level in the 1980s. There is roughly a 72 % drop in slope in this flow series, an indication of decreasing volumes per year. The 2FC09 (SWSII) series however, display a different pattern in the 1960s. It begins with a smaller slope, increases by 56 % in the 1970s and reduces by 60 % to reach its lowest in the 1980s. There was overall a 37 % decline in slope between 1960 and 1990. These findings therefore seem to indicate a clear decreasing trend of flows in both subwatersheds, with the subwatershed I (2FC05 series) having the largest effect. The rapidly changing nature of flows in the 2FC09 series could be attributed to the nature of land use change from forestry and agriculture to agriculture and urban.

5.5.1.5. Mass Curves of the Annual Flow Residuals

Similarly, the mass curve analysis was performed on the annual flow residuals using the CUSUM function in equation 5.8. The procedure was adopted because the residual flow diagrams present a hydrological snapshot of a watershed for a particular flow condition. As a result, each significant influence on the river is quantified and an overall picture of the total flow at any point of the system is presented (Pirt, 1987). The mass residual curves of the flow series shown in Figure 5.29 were used to quantify the flow at different locations and time with respect to the historical watershed characteristics.

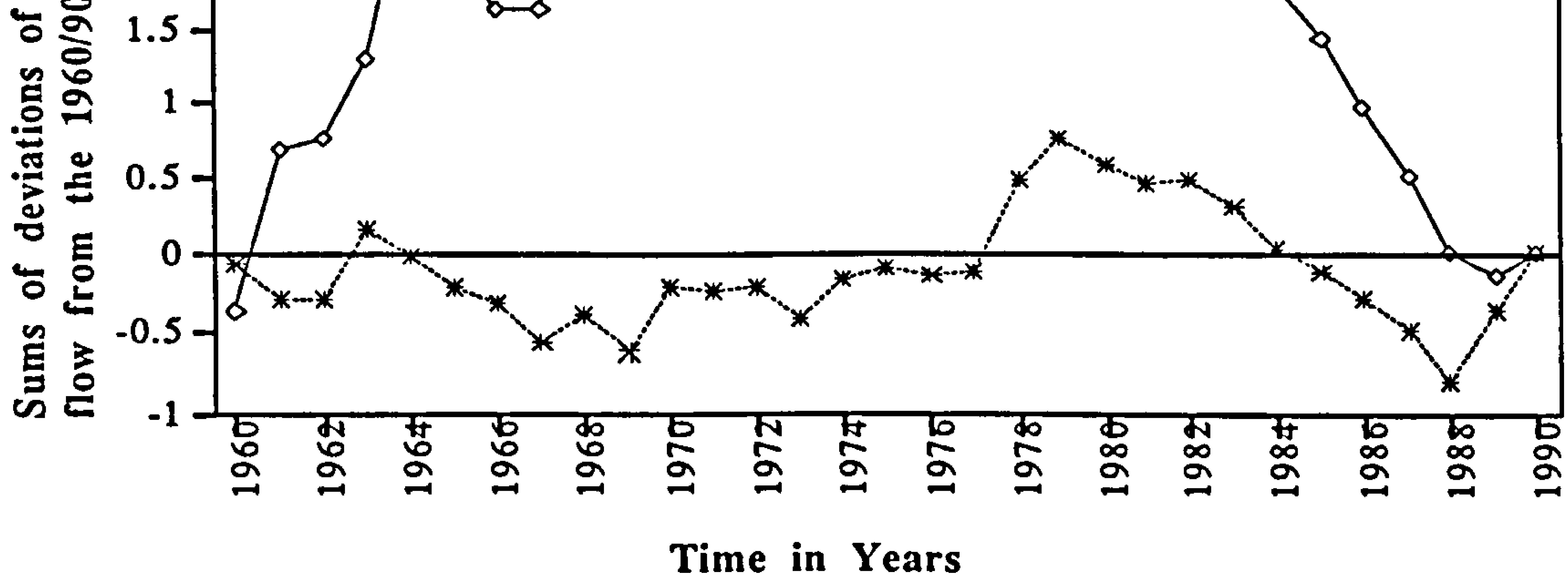


Figure 5.29. The mass curve plot of the flow residuals in the watershed

Further examination of the flow series was accomplished by using the CUSUM (Q_t) function earlier discussed. Clearly the residual mass curves of the SWSI series have departures only in one and positive side of the y = 0 line, which indicates that the flow series have probably a trend. The SWSII flow series maintains a negative departure in the 1960s and 1970s, and crosses the y=0 line in 1979 and returns to no change after 1988. It is difficult to decide the nature of the pattern of this flow. Perhaps, there is no significant trend, because the curve crosses the main line too often. However, this rather uniform and straight curve does not say much on the series, and therefore, the series cannot be adequately described at this point. Nevertheless, it suffices to conclude that either the nature of land use in the watershed influenced the flow or the data series are inconsistent. Other tests are used to explore hidden patterns and trends in the series.

5.6. SEASONAL STREAMFLOW REGIME

Monthly values constitute the highest resolution that can be analysed with the present data base. This resolution is relevant if extreme values are of interest for detecting the effects of river regulation (Bergström and Carisson, 1994). It is also applicable because the distribution of streamflow throughout the year, describes the seasonal flow regime. Since rainfall showed a distinct month-to-month variation, the seasonal flow regime should assume a similar pattern. If the response to rainfall is non existent, then the cause-effect relation is further sought by analysing the relationship of the flows and land use change, which this study postulates to contribute to changes in hydrological regime in the watershed. To start the analysis, it was found necessary to examine the time series of the seasonal flow in Figure 5.30.

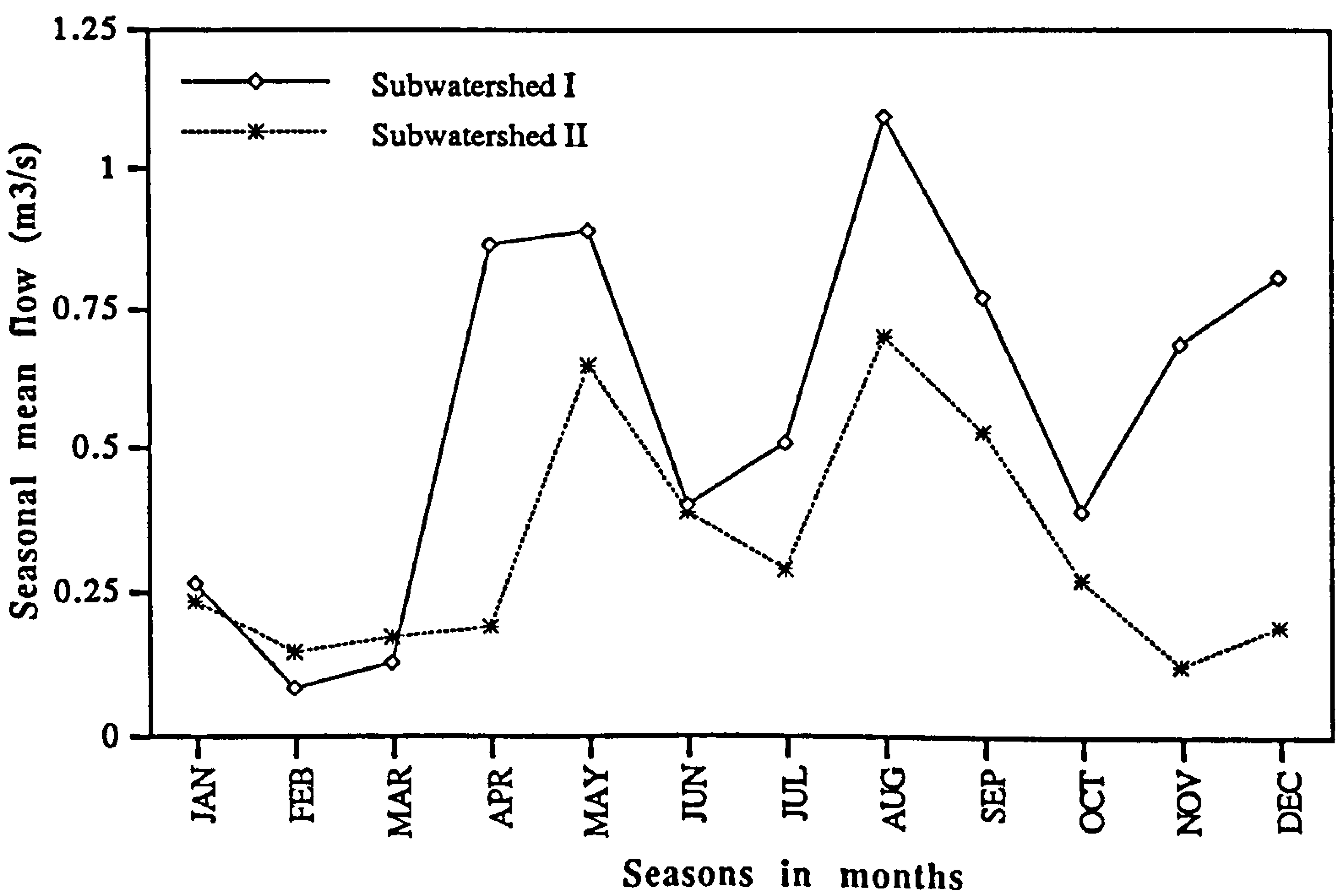


Figure 5.30. Time series of the seasonal mean flows

The seasonal behaviour of the flows shown in Figure 5.30 is shown to follow the pattern earlier seen in the seasonal rainfall pattern of having peaks in April, August, and November. The series in SWSI is larger than that of SWSII at all the seasons, as they correspond to the rainfall peak seasons, rather than in SWSII where the flow peaks seem to have one month lag in May, responds very well in August, and does not respond at all in November. This response mechanism becomes clear when the relative runoff factors or hydraulicity are considered. During the driest season (February to March), however, the flow regime in SWSII is higher than the upper SWSI, an indication that SWSI may in fact be not holding moisture for a long time, or its recharge source is depleted of moisture faster during these dry spells.

The distribution of the streamflow is also described by the ratio of the mean monthly flow and the mean annual flow (Q_{mi}/Q_y), for the two data series. The objective is to establish how much of the ith mean monthly flow contributes to the annual flow. This relationship is sometimes referred to as the "hydraulicity "of the river regime or the relative runoff factor (Parde', 1955). The ith monthly flow (Q_{mi}) is expressed as a percentage of the annual mean flow (Q_y) as:

$$K_{mi} = \frac{Q_{mi} * 100}{Q_y} \qquad (5.10)$$

where, K_{mi} is the long-term monthly relative factor or hydraulicity (dimensionless),

Q_{mi} is the ith month/season flow; 30-year monthly mean flow (m^3s^{-1}),

Q_y is the long-term annual mean flow (1960-1990 mean in m^3s^{-1}).

Using this representation, the seasonal variations of K_{mi} were produced and used to detect the seasonality or regime in the flow series. A plot of these relative flow factors against time is presented in Figure 5.31. The ratios in most of the time are consistent and correspond to the seasonal rainfall regime. Seasonal peaks occur in April, August, and to a lesser extent in November that correspond to the long-term seasonal behaviour of the rainfall in the watershed. It is seen that SWSII contributes its annual flow during the wet months of April and August while SWSI seem to contribute flow throughout the months except in February and March.

The seasonal trends of flows in SWSI shown in Figure 5.32 was further examined on a partial basis; (1961-1970); (1971-1980); and (1981-1990). It is clear that the 1960s has higher monthly volumes, followed by 1970s, and least in the 1980s. The 1960's pattern again corresponded to the rainfall regime in April, and August, but not in November. The 1970s had a one month lag from the rainfall peaks of April and August, but they display a continuous flow throughout the decade, probably from volumes retained in the watershed from the 1960s' surpluses, or some physical changes in the watershed contributed to this scenario. The 1980s is the least flow periods. The cause is yet to be established. What is interesting is that, the watershed responded immediately to the rainfall as depicted by peaks in April, and August, and a one month lag in December. This may indicate that the subwatershed did not retain much more rainfall than it released into the streams.

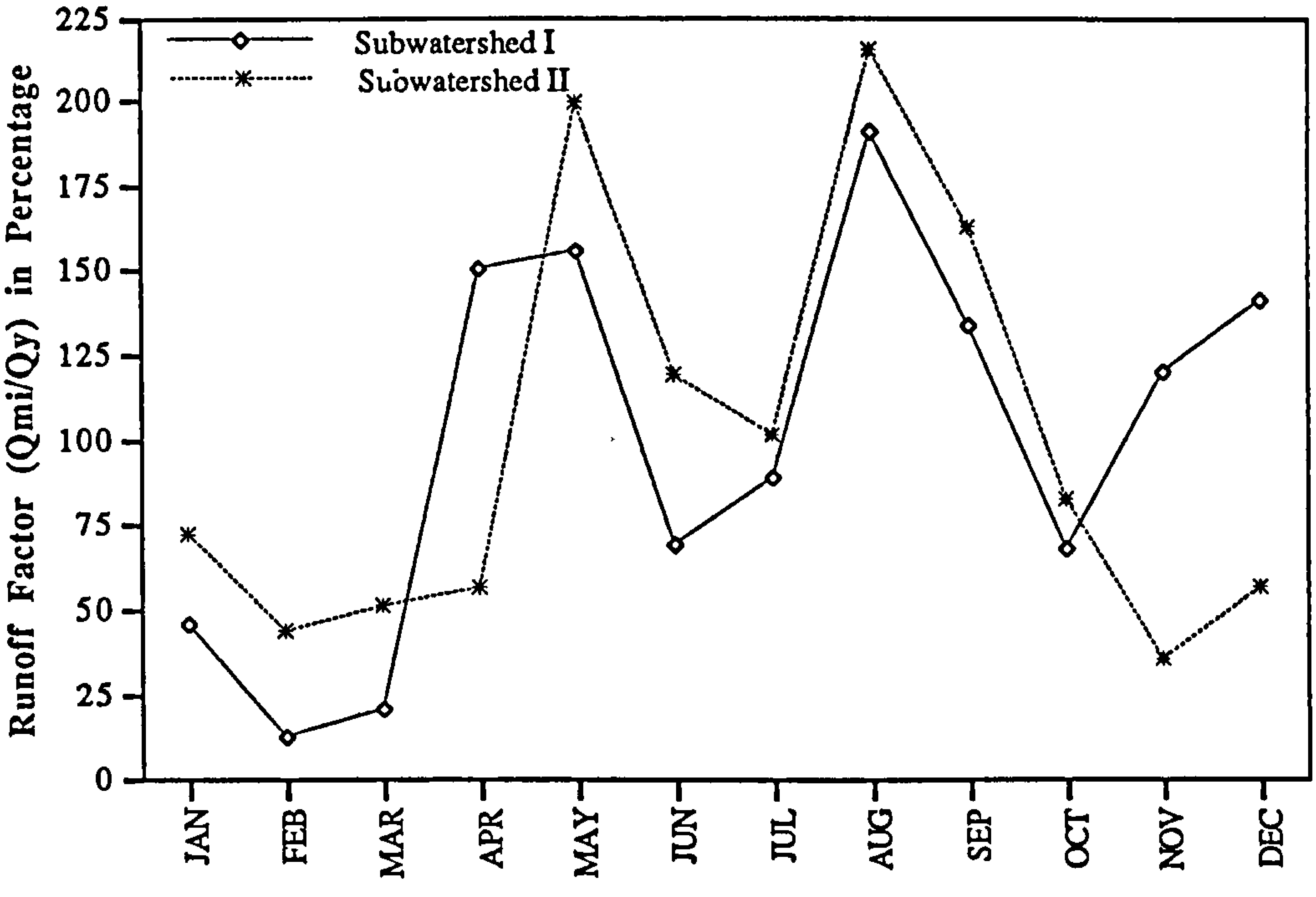


Figure 5.31. Seasonal variation of the runoff relative factor in the watershed

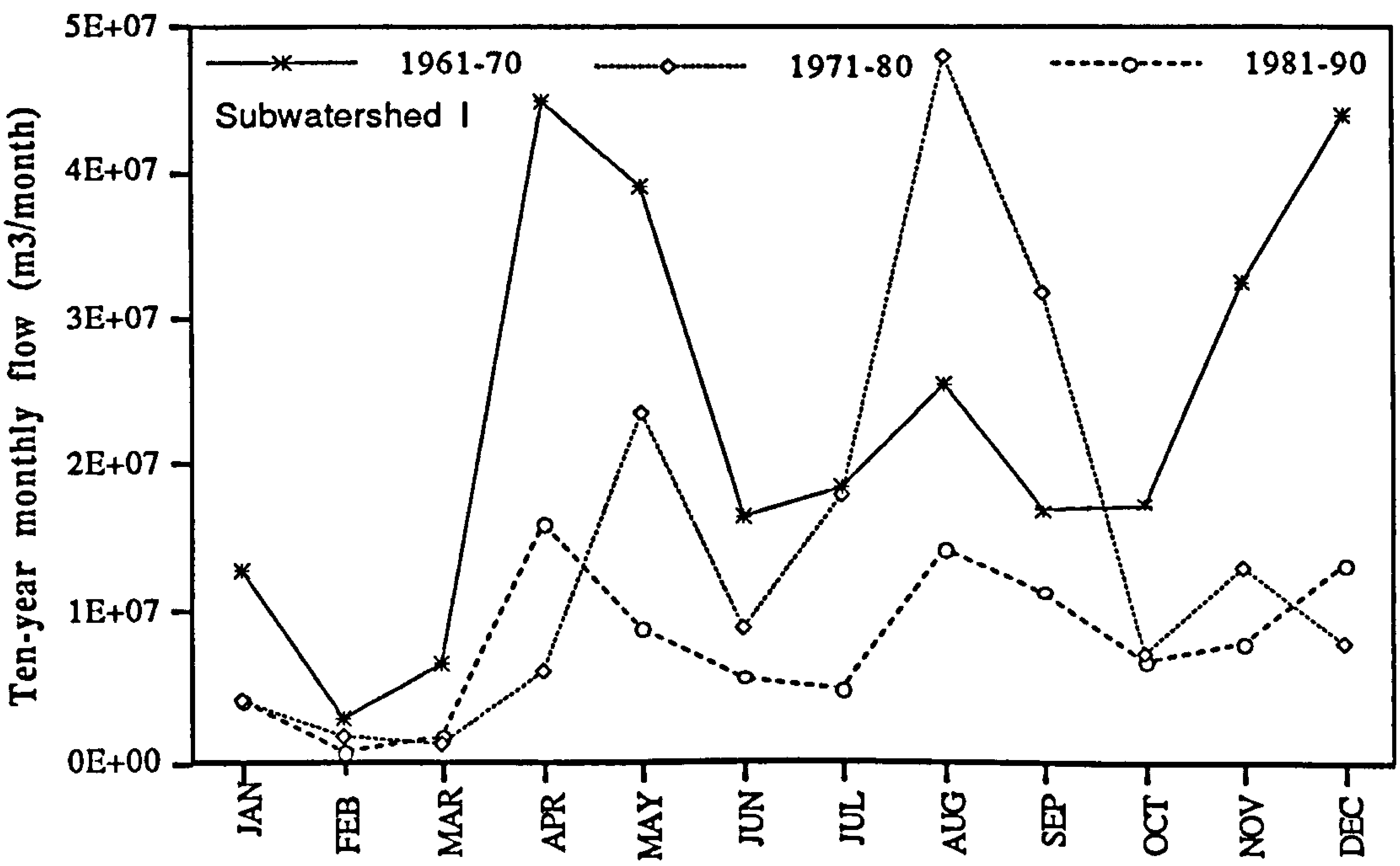


Figure 5.32. Trends of the mean monthly flows in subwatershed I

Similarly the lower subwatershed II was examined on the same partial period basis and the result shown in Figure 5.33. The 1960s again dominate the high flow regime with peaks in April, and August. The 1970s, however, had a lower flow regime during the first six months before peaking in August and maintained a higher pattern than the other periods. The 1980s had initially a higher flow from March to May, but begun to assume the lowest flow from July and decreasing towards the end of the year. This may indicate that, the small accumulation of flow in the early months of the year, was not enough to sustain the flow through out the year and generally, the flow regime during the period is considered to be relatively low.

To see the evolving flow regime clearly, the cumulative sums of the mean monthly volumes were computed and presented in Figures 5.34 and 5.35 for SWSI and SWSII respectively. The evolving opinion that 1960s experienced higher flows than the rest of the decades under consideration is once again confirmed in Figure 5.34 where there is a clear larger volumes of flow, decreasing slowly into the 1970s and experiencing a drastic drop or decrease in the 1980s. This is evidently shown by the 1981-1990 curve being displaced far to the right and tending to a flatter slope, an indication of a very low flow, which can only be described as a hydrologic drought and lasting a decade.

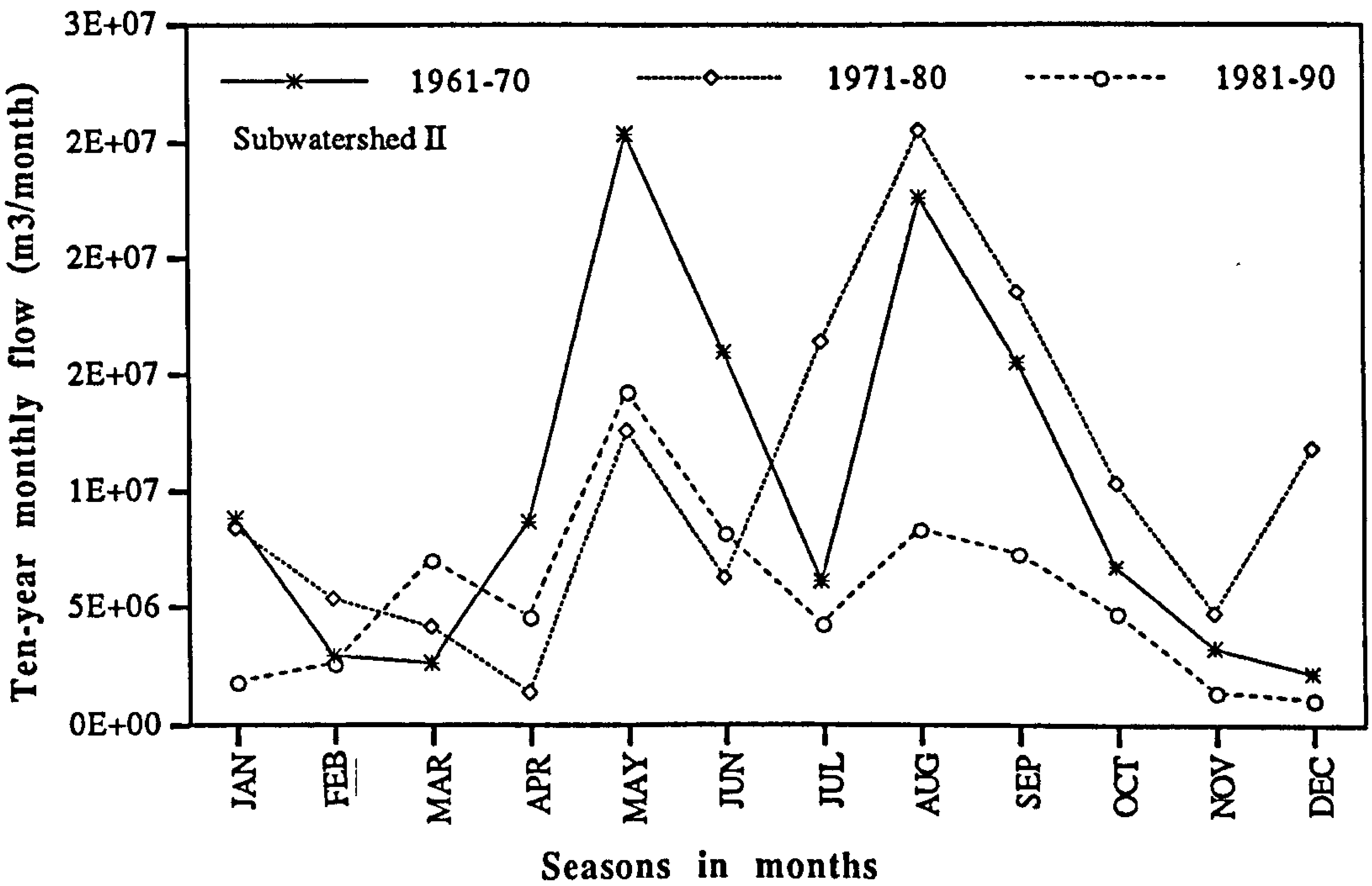


Figure 5.33. Trends of the mean monthly flows in subwatershed II

A similar pattern can also be seen in the SWSII series between 1970 and 1990 as shown in Figure 5.35. The only difference is that the deflection of the 1980's curve to the right occurred later in June, unlike in the SWSI where it occurred immediately after the usually rainy season. The difference of flow during the 1960s and 1970s is more significant, because of the wider gap between the 1970s, and 1980s' curves, which shift to the right indicating lower volumes. The slopes of the 1980s curve however, seem to recover, because it is steeper, a condition that can only be attributed to a changed watershed characteristic that seem to encourage moisture retention for stream recharge.

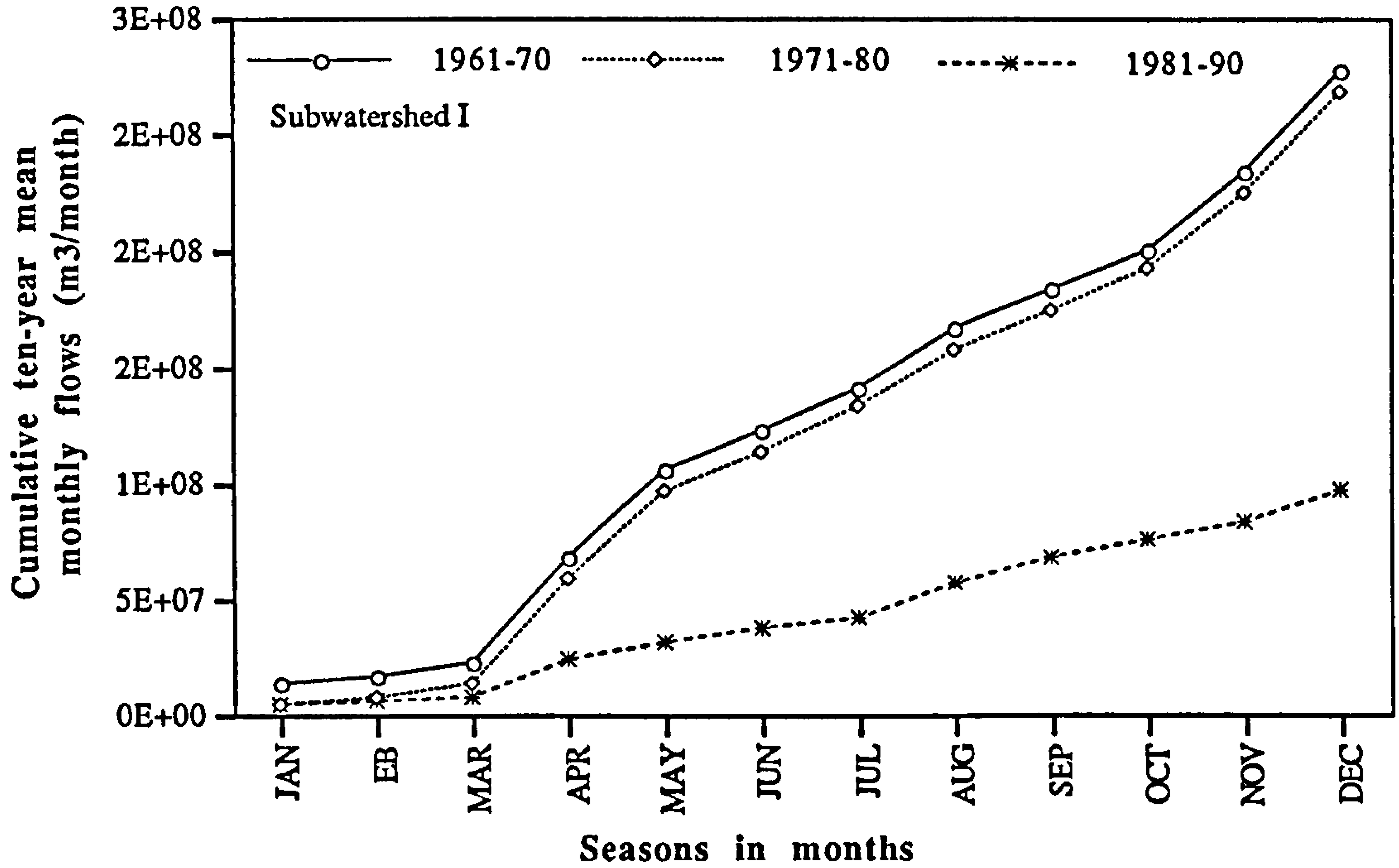


Figure 5.34. Sums of the ten-year mean month flow in subwatershed I

The analysis of the two subwatersheds revealed existence of hydrological differences and similarities between SWSI and SWSII. The differences are postulated to originate from changes in land use, relief and climatic factors. However, changes in land use cause and reliability of the data can be assumed to have caused the spatial differences. The series have a similar temporal, as trends in the subperiod flow analyses arrived to the same conclusion of decreasing trends in flows from the 1960s to the 1980s.

What causes the differences in water volumes from the two subwatersheds is presently unknown, because in an ideal homogeneous watershed, volumes of water should be increase in magnitude into SWSII, since flows from SWSI pass through SWSII. On this basis, the analysis from here henceforth, assumes that the results from SWSI alon should serve to reveal the causes of the changed temporal streamflow patterns and hence meet the objectives of the study. Thus, SWSII is removed and introduced only at some sections when a comparison between the two subwatersheds is required.

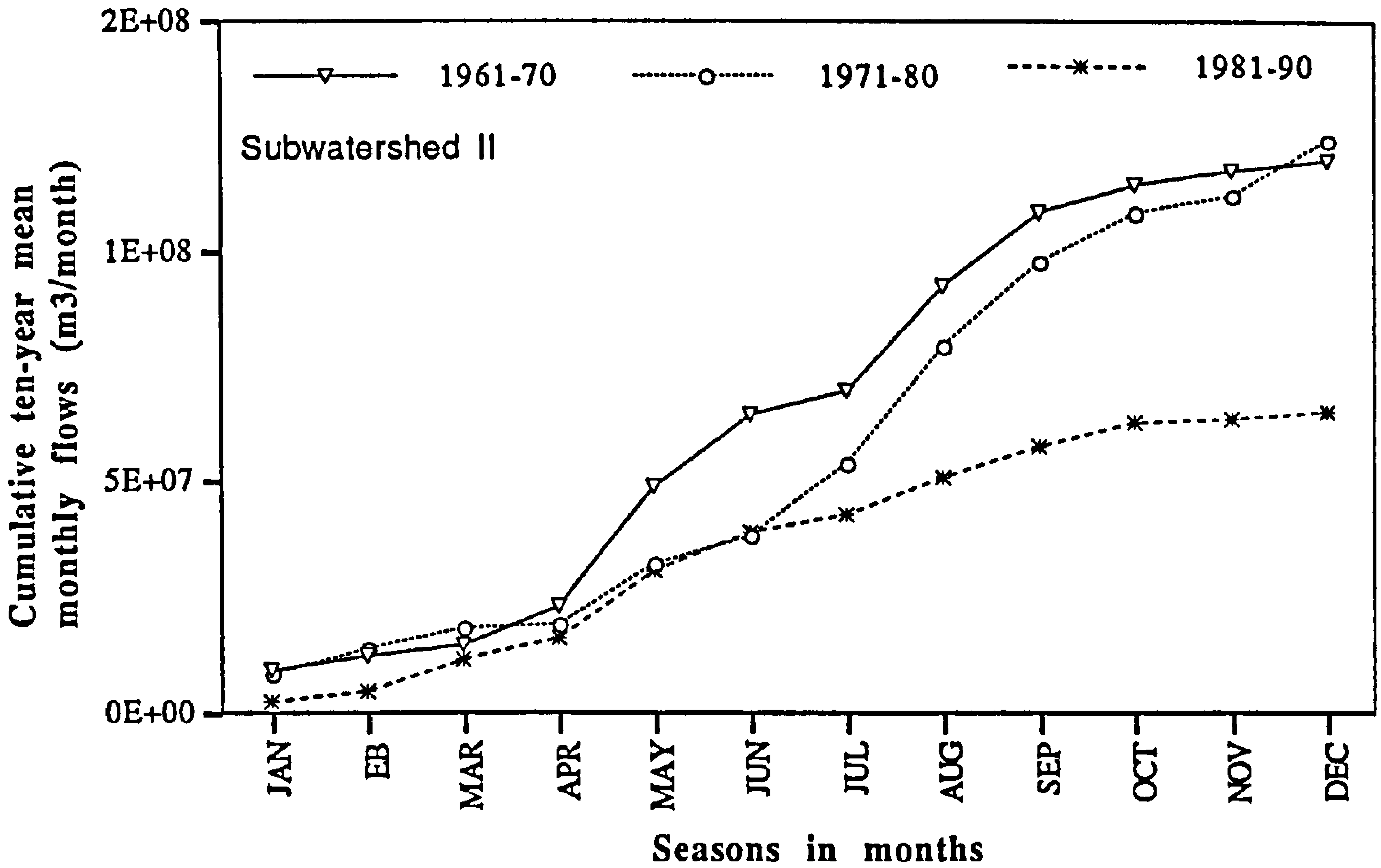


Figure 5.35. Sums of the ten-year mean monthly flow in subwatershed II

5.7. VARIATIONS IN CATCHMENT WATER BALANCE

This subsection describes the variations of the catchment water balance so as to ascertain the evolving drying regime. The results will also establish the role of human activities in aggravating the reduction in moisture storage capacity of the subwatershed. Changes in land use and its management affect streamflow generation especially in small watersheds. Therefore, the knowledge of the water balance components and processes provide the key to understanding, forecasting, and controlling these changes.

Evapotranspiration is the main component that is affected mostly by any changes in land use. In humid regions, moisture is readily available to satisfy potential evapotranspiration (i.e. PET = ETa) and the mean annual rainfall is regulated by climate alone(Szesztay, 1979). In the semi-arid and subhumid-tropics however, evapotranspiration does not reach its potential value, and in addition, the actual evaporation is greatly affected by the vagaries of climate and human-induced water use and unavailability. The estimation of the water balance in these regions therefore, are approximate in most cases.

5.7.1. Estimation of Evapotranspiration

Evapotranspiration is composed of evaporation and transpiration and are governed by different physical processes. Unlike rainfall and streamflow, evapotranspiration is indirectly measured. Consequently, a suitable estimation method was chosen. Among the direct methods, the use of lysimeters considered as the best device for determining evapotranspiration (Kova'cs, 1976a) is expensive and labour intensive. The calculation from energy balance on the other hand sometimes introduces some uncertainties. Factors influencing evapotranspiration were extensively discussed in Chapter III.

The Penman (1948) equation in FAO (1990) modified Penman-Monteith version is the preferred method to estimate reference ET. Because of inadequate data and lack of a continuous albedo value for a time series purpose, other methods were also considered. A simpler method under these circumstances is that developed by Christiansen (1968), and later Christiansen and Hargreaves (1969) which estimates PET from the USWB class A pan evaporation and is reproduced from equation 3.8:

$$PET = 0.755\, E_{pan}\, C_{T2}\, C_{W2}\, C_{H2}\, C_{S2} \qquad (5.11)$$

where the coefficients have been defined.

In addition, a simple and reliable temperature based procedure for calculation of crop ETo developed by Hargreaves (1983), and Hargreaves et al.,(1985) was also selected. McVicker (1982) evaluated 12 equations for estimating crop reference ETo and ranked those equations using only solar radiation and mean temperatures first and recommended for use with watershed models. Hargreaves (1983) also compared and evaluated the various equations and concurred with McVicker's selection. The final equation (Hargreaves et al., (1985) which improved the estimation by calibrating it with actual data from 82 stations in Africa, 62 in Brazil and in Davis, California is:

$$ETo = 0.0022 * R_A * (T_m + 17.8) * TD^{0.5} \qquad (5.12)$$

where R_A is the extraterrestrial radiation in equivalent (mm/day), T_m is mean air temperature (^oC), TD is the mean maximum minus mean minimum temperature (^oC), and the coefficient, 0.0022 is applicable in Africa.

The original Christiansen equation referred only to PET, but the PET, may be regarded as grass reference PET, because data from the grass surfaces were used as calibration input to the method. Jensen et al., (1990) and FAO (1990) further ranked ETo estimation methods, with the Penman-monteith and related modified versions taking the first six positions and Hargreaves et al., (1985) were ranked eleven after the FAO 24, (1992) and Jensen-Haise equations. On grouping all Penman related equations however, it was found that this simple easy-to-use method, takes the fourth position out of 20 methods, overestimating PET by 8% of actual lysimeter measurements (FAO, 1990)

The selection of these depended upon several factors. These included the accuracy required and the availability of the climatic data. Jensen et al., (1990) recommended the use of the combination or the energy balance-aerodynamic equations to estimate PET because they are based on the physical laws and rational relationships. Penman (1948, 1963) and FAO (1990) Penman-Monteith version are singled out as providing reliable estimates for grass PET under both the semi-humid and humid conditions.

Where, only the maximum and minimum temperature data are available and ex terrestrial radiation tables are available, the Hargreaves (1985) method was recommended for up to mean monthly estimates. Pan evaporation data, were found to be reliable if the evaporation site was well maintained and data collection was consistent, and hence the use of the Christiansen and Hargreaves (1969) still provided a good estimate of the PET (Jensen et al; 1990).

Since the watershed area under investigation had limited continuous meteorological data from 1977 to 1992, the FAO(1990) modified Penman-Monteith, Hargreaves et al., (1985) and Kenya pan A evaporation approaches were selected to estimate monthly ETo values for the period. Using these approaches, the estimated ETo, was multiplied by a correction factor and vegetation cover coefficients to obtain the estimated maximum evapotranspiration (ET_m). To account for the effect of the vegetation cover and other land uses in the watershed, an integrated cover coefficient (k_c) was introduced that converts the ETo into ETm:

$$ETm = k_c .ETo .f \qquad (5.13)$$

where,

k_c is a land use (vegetation cover) coefficient approximate for the ETa, f is the lysimeter adjustment factors, 1, 8 and 18% for Penman-monteith, Hargeaves et al.,(1985), and Pan A methods given in FAO(1990).

Since an average Kc for all the physical watershed characteristics and for an entire time period is difficult to obtain, an approximate estimate was derived from the average of the Kcs from tree crops grown in tropical areas with a moderate wind speed (175-425 kph), a medium relative humidity (40-70%) and limited moisture availability. From

Tables 25 and 26 in FAO-24 (1992), an average Kc value of 0.75 was selected. Using this constant, the following representations were obtained for each method:

Penman-Monteith: ETm = 0.75*ETo*100/101 = 0.7425*ETo (5.14)

Hargreaves et al : ETm = 0.75*ETo*100/108 = 0.6944*ETo (5.15)

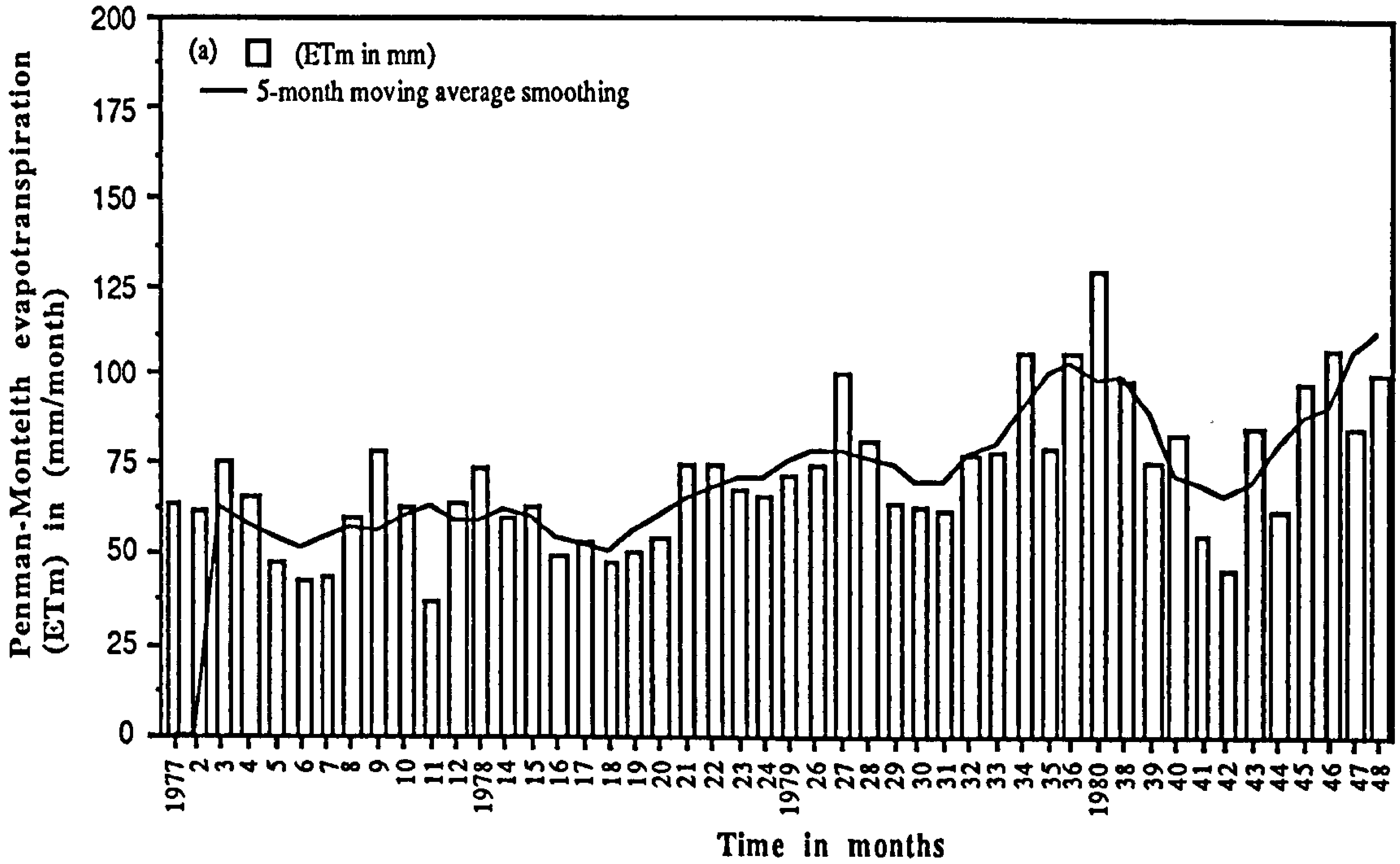
Evaporation pan: ETm = 0.75*Epan*100/118 = 0.6354*Epan (5.16)

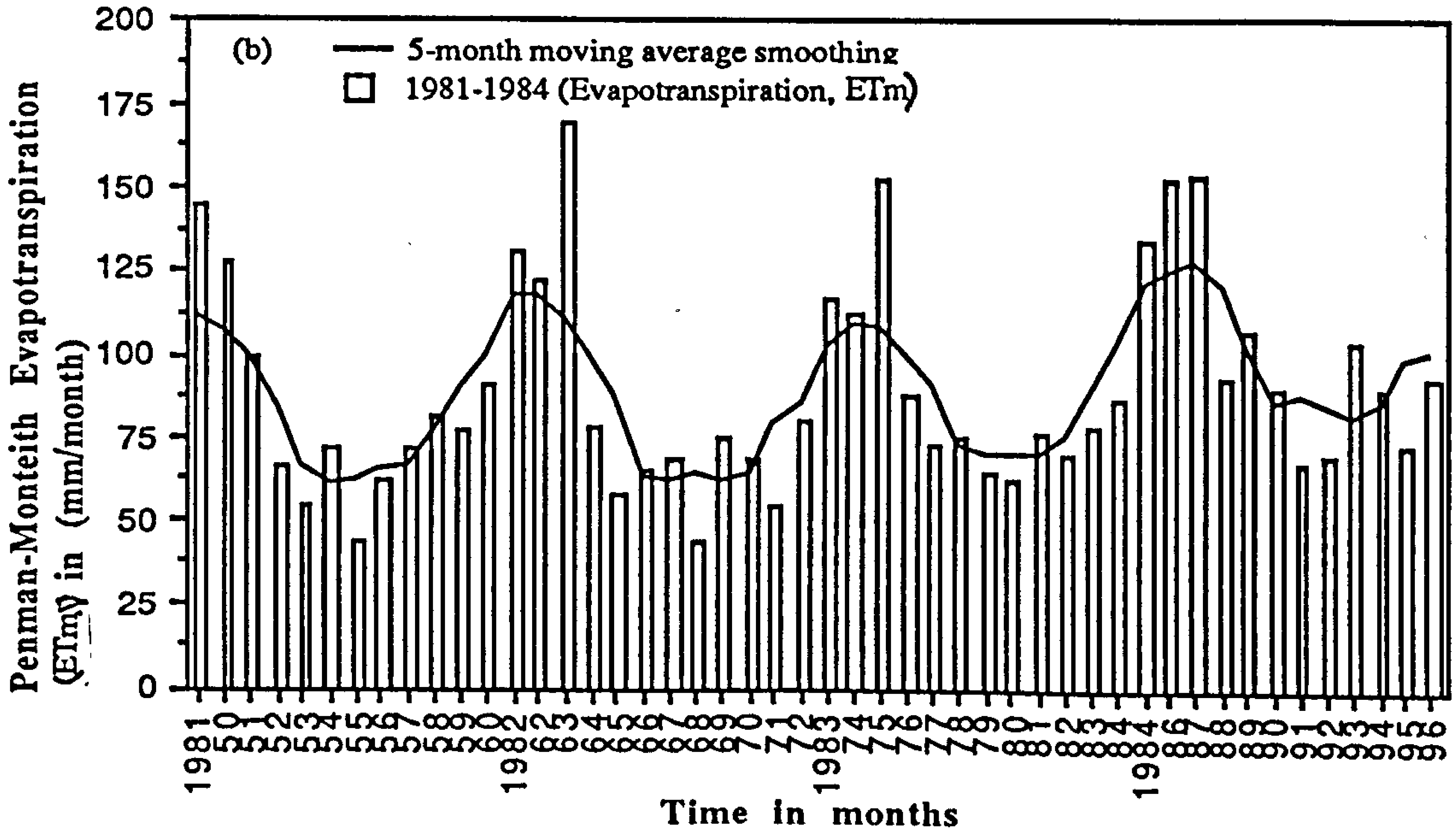
Calculations and results for the three methods using mean watershed rainfall and climatic data from Egerton station in SWI are given in Appendix A.2. Results for the preferred Penman-Monteith approach are presented in Figure 5.36.

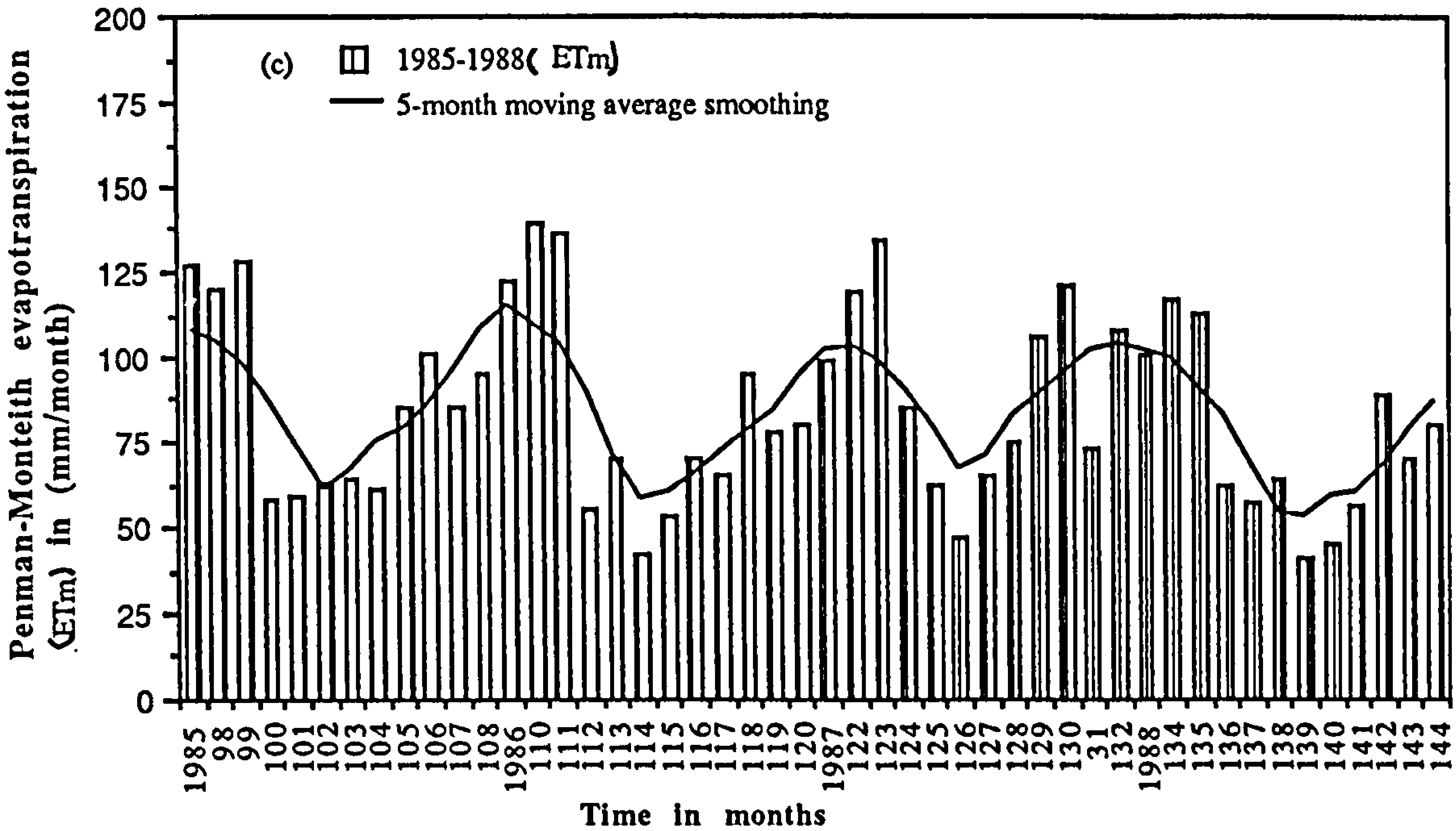
The results from these methods were compared on seasonal basis in Figure 5.37. The methods estimated the higher ETm values during the dry seasons, ranging from 110 mm in December to 117 mm in March. Penman-Monteith method overestimated the ETa values during the dry seasons and under estimated during the wet seasons when compared to the mean values from the three methods. Hargeaves et al., (1985) and Kenya Pan A methods did not differ much except during the wet season when Pan A estimation was relatively higher. The two methods overestimated the ETm during the wet seasons and under estimated in the dry seasons relative to the mean value. Since the Penman-Monteith was established as the best method (Blackie and Eeles, 1985, Jensen, 1990), it was used in this analysis. However, during the period between 1960 and 1976 when climatic data were inadequate for Penman-Monteith approach, the Kenya Pan A was adopted.

5.7.2 Trends in Water Balance Components

In evaluating the streamflow generation in human-modified watersheds, a useful approach is to inspect the annual trends of the water balance components. The first step of the assessment was to provide a picture of the three components. Thus, the annual mean watershed rainfall (R), streamflow (Q), and estimated areal reference evapotranspiration (ET_m) were used to compute the magnitudes of the annual watershed recharge (R-Q), aridity (R-ET_m) and change in watershed soil moisture storage Δ(R-Q-ETm) with time. Seasonal and annual time series plots of the historical record for the flow (Q), rainfall (R) and the estimated evapotranspiration (ETm) are given in Figures 5.38 for SWSI.


(a) (ETm in mm)
5-month moving average smoothing
Penman-Monteith evapotranspiration (ETm) in (mm/month)
Time in months


(b) 5-month moving average smoothing
1981-1984 (Evapotranspiration, ETm)
Penman-Monteith Evapotranspiration (ETm) in (mm/month)
Time in months



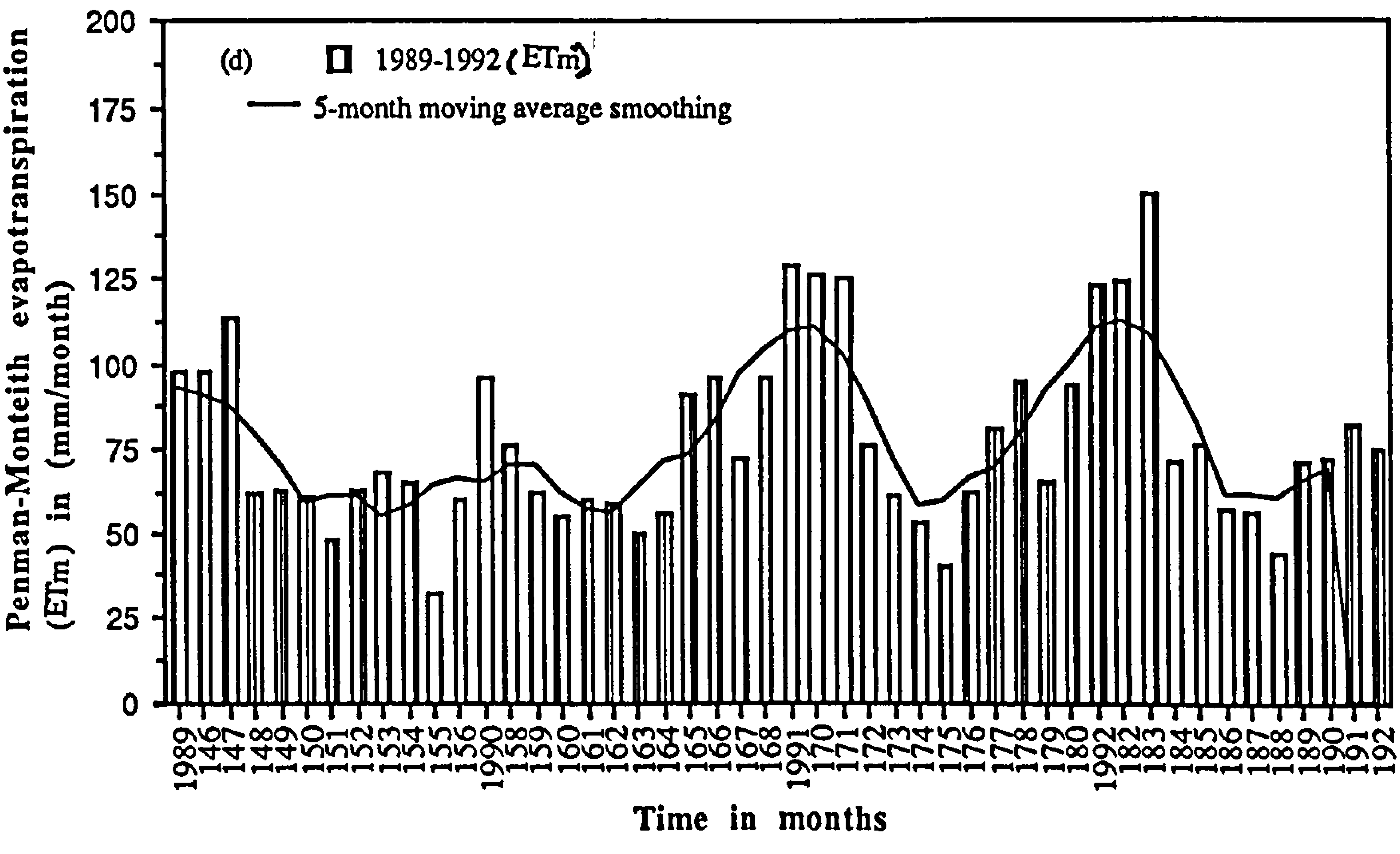


Figure 5.36 (a-d). Time series of the estimated ETm in the upper reaches of the watershed using the FAO (1990) Penman-Monteith method.

The time series of the three water balance components are shown in Figures 5.38. Subjectively the components responded to changes in each other. Rainfall however, is the most noticeable controlling component. During meteorological drought years in

1965/66, 1973/75 and 1983/86, rainfall regime influenced the water balance regime. The ETm responded favourably well to the rises and falls in the rainfall. The flow regime was affected during dry years and particularly in the 1980s, when it reached its lowest levels in 30 years.

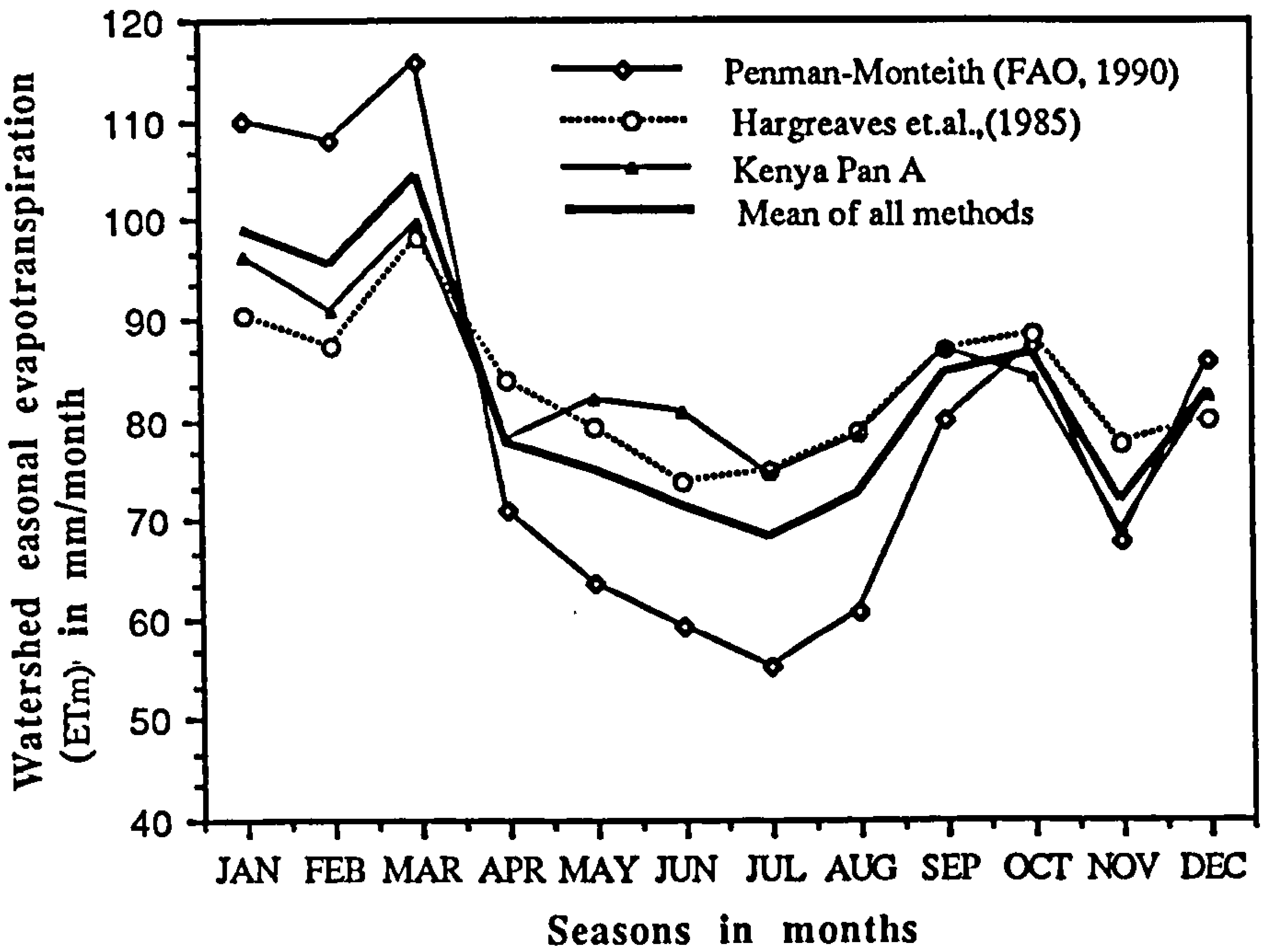


Figure 5.37. Seasonally estimated evapotranspiration from the FAO modified Penman-Monteith, Hargeaves et al.,(1985) and Kenya Pan A methods

To gain a better insight of the changes in water balance components, Kleme's (1983) recommended the use of a 'down search approach' "(...a systematic search, moving step by step on the basis of the results of the proceeding one...this can lead to significant findings if each step is evaluated separately and the hypotheses based on its results tested)". The search proceeds with the assumption that the relation between rainfall and the flow is a straightforward one, and that the rainfall-runoff-relation (RRR) can be plotted against time in order to identify points of change in its time series. Analysing the process rather than synthesising is therefore adopted.

This section examined the various leading works on the evaluation of the water balance components by Kleme's (1983), Morton (1983). On the basis of these leading works, Kleme's (1983) approach of systematic search was adopted. Consequently, the RRR was examined first by plotting the annual watershed recharge (R-Q) against time to search for evidence of rainfall losses in the watershed. The objective was to establish the amount of water supposedly retained in the watershed, assuming an ideally homogenous watershed term.

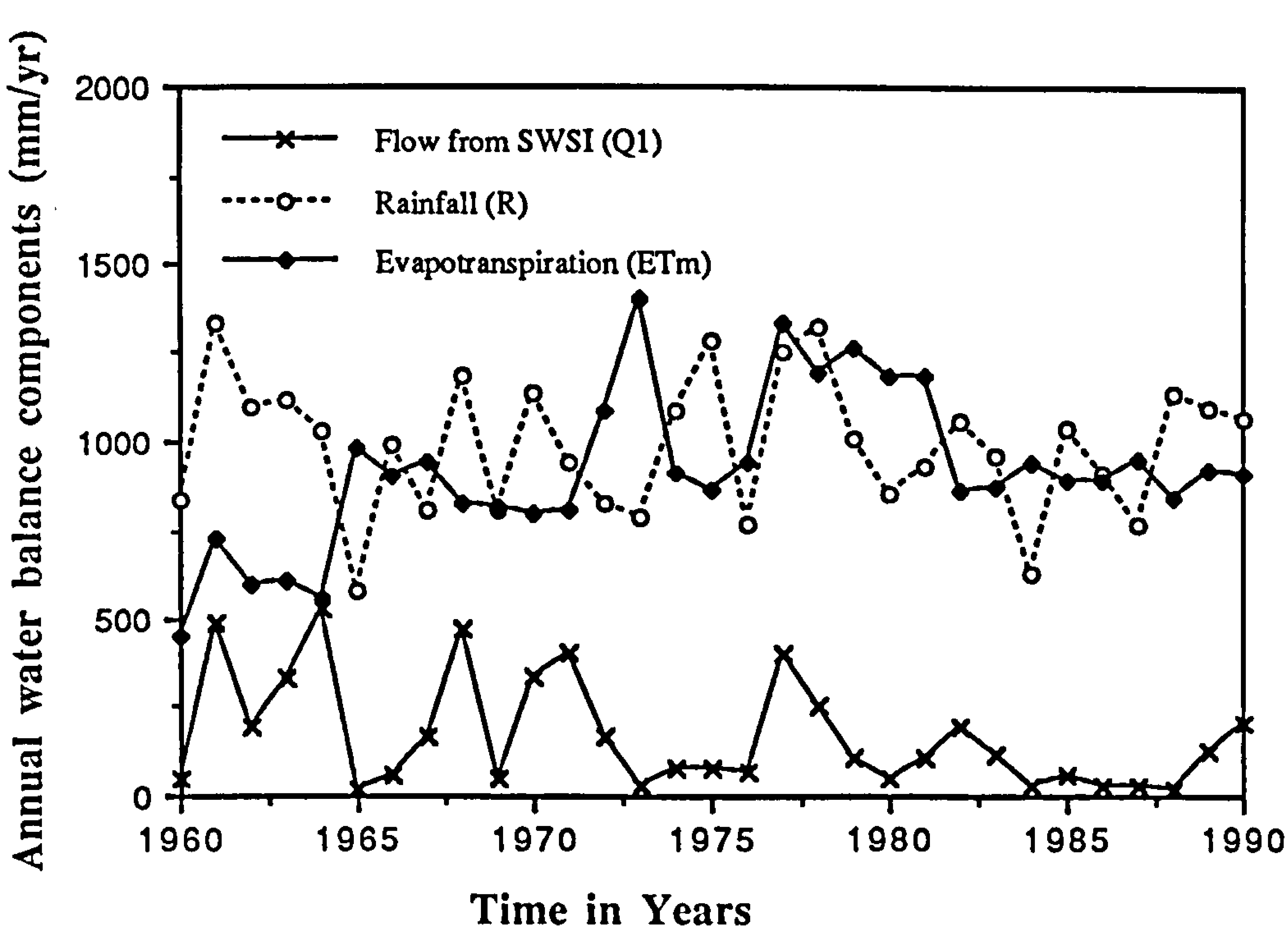


Figure 5.38. The annual totals of rainfall (R), runoff (Q), and estimated evapotranspiration (ETm) in subwatershed I

The next step was to introduce the estimated ETm and compare with the recharge in the ratio (R-Q)/ETm). The aim here was to search for its relationship and thus establish the amount of water that could have been lost through other sinks namely, deep percolation, abstraction and inter-basin transfer. The ratio is plotted against time in Figure 5.39. The larger the ratio, the more water is retained, and hence a possibility of a continuous recharge of the river. Conversely, a smaller ratio indicates that the recharge is not meeting the evaporative demand, and hence the watershed is drying up. A continuously drier conditions eventually lead to hydrologic drought in the river basin.

From the physical nature of the watershed, if the outflow of the incoming water is delayed, the water must be temporarily stored in the hydrologic system, and if it is removed it should be seen. Therefore an analysis of the watershed storage changes is necessary. The water balance equation in mm/Δt was resolved with the remainder term ΔSUM_i that was subdivided step by step by reconstructing changes of total watershed storage for an ^{i}th year as:

$$\Delta SUM1 = R_i - Q_i$$
$$\Delta SUM2 = R_i - Q_i - ETm_i$$

$\Delta SUM3 = R_i - Q_i - ETm_i - GWS$

where GWS is the ground water storage. By equating this water balance to zero, it is possible that GWS can be estimated by summing the storage for a long time, usually, a year is considered adequate (Ward and Robinson, 1988; Dingman., 1994) thus:

$$\Delta SUM_i = \{R_i - Q_i - ETm_i\} \tag{5.16}$$

where ΔSUM_i is the change in storage in the ^{i}th year (mm/yr) and the end-of-year storage of year i is:

$$SUM_i = s_o + \sum_{t=0}^{i=t=N} \Delta SUM_t \tag{5.17}$$

The value of the initial storage S_0 is unknown, but it was set arbitrarily set to zero. The watershed was then regarded as a semi-finite reservoir of the bottomless type (Kleme's, 1983; Morton, 1983, and Jain, 1994). This systematic search allowed for a further investigation of a relationship between the time (t), and the watershed storage SUM_i presented in Figure 5.40 and sums of the cumulative SUMi in Figure 5.41.

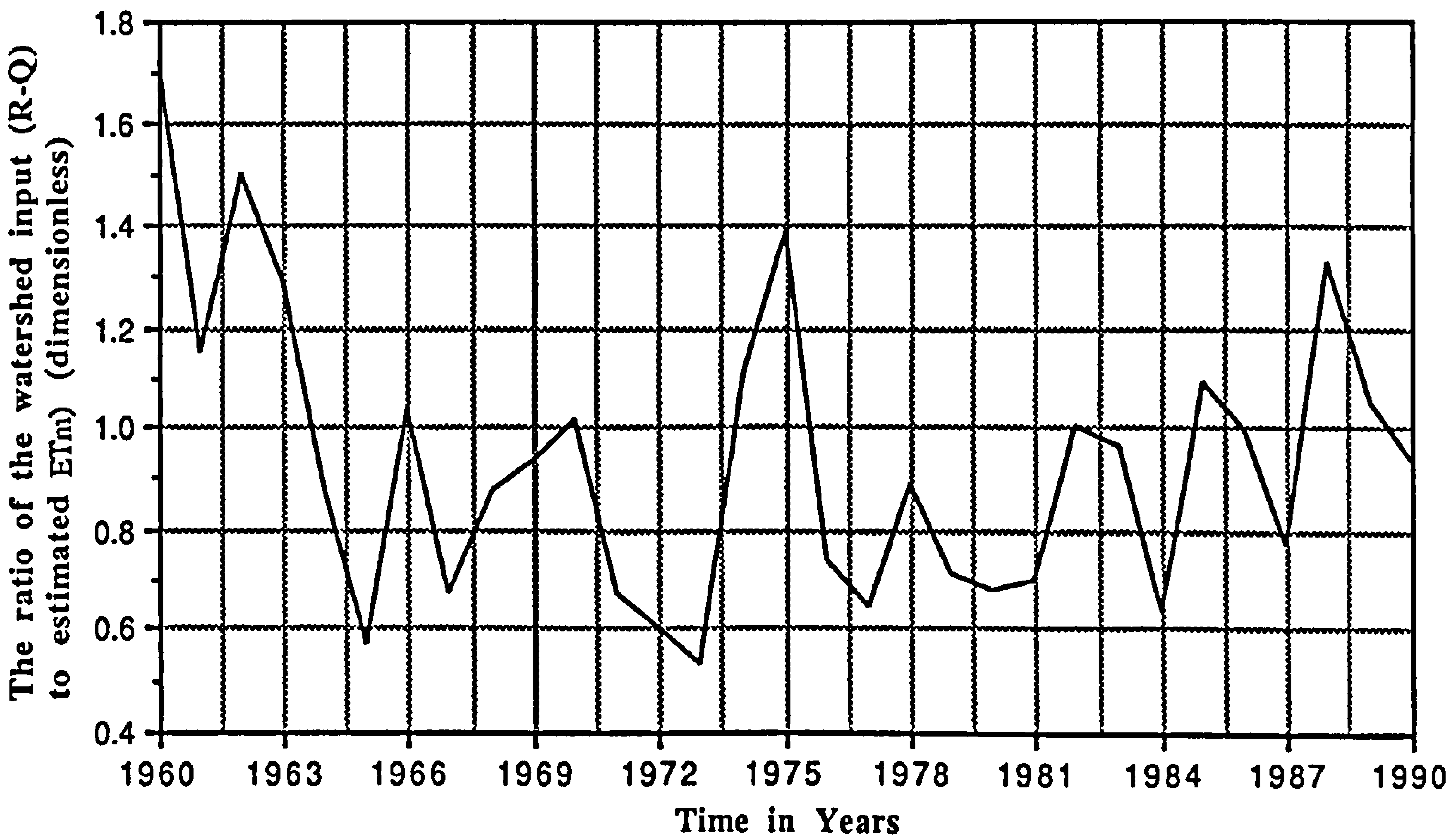


Figure 5.39. The time series of the ratio (R-Q)/ETm in the watershed (1960-1990).

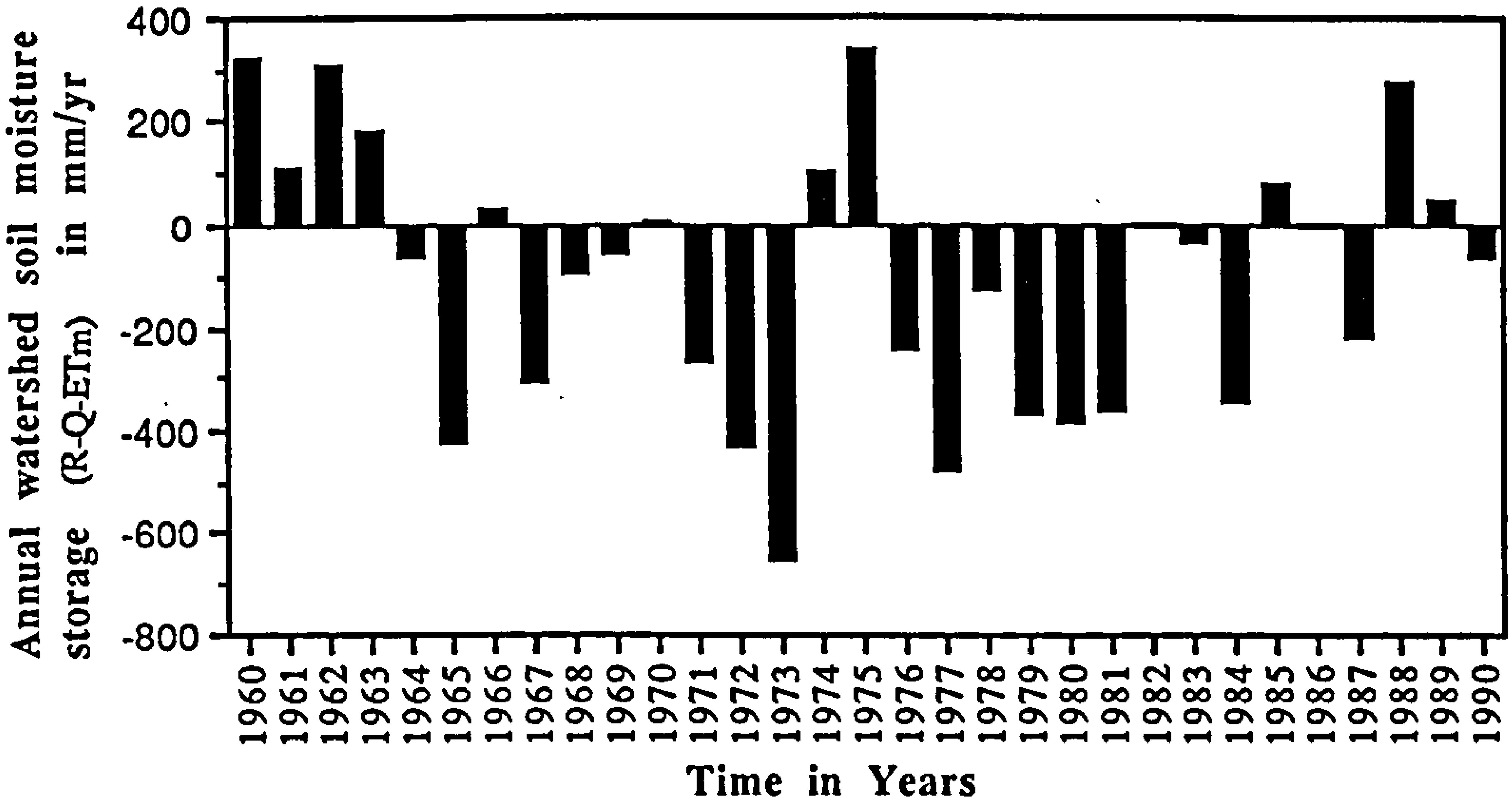


Figure 5.40. The time series of SUMi in subwatershed I

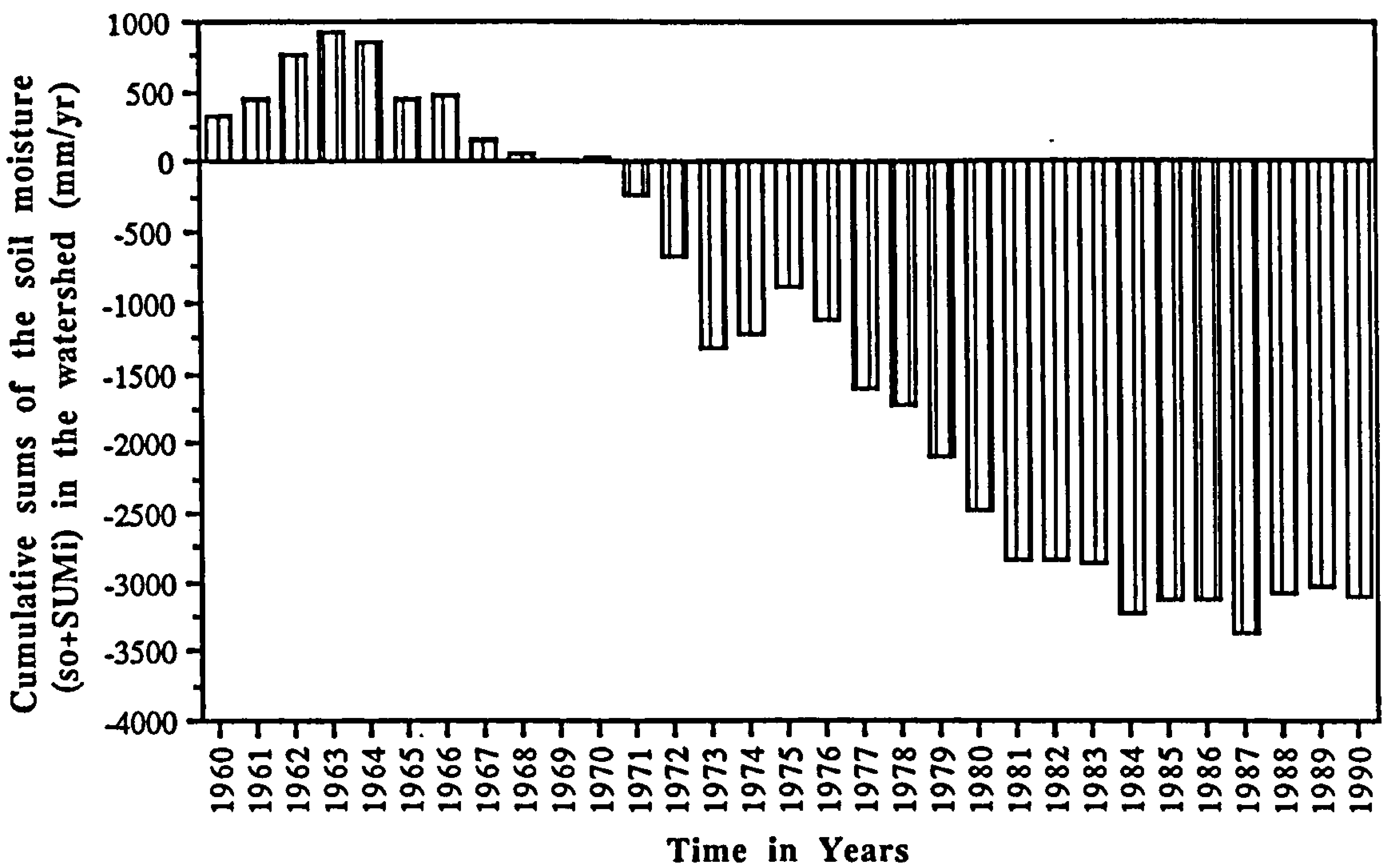


Figure 5.41. The mass plots of the SUMi in subwatershed I

To understand further the nature and extent of the observed watershed moisture changes a difference between the mass curves from the two subwatersheds was examined by plotting their mass flow curves (SWSII- (SWSI) against time from the results in Figure 5.29. The time series of this computation is given in Figure 5.42 that gives negative values and indicating that not all the water from SWSI passes through SWII. The reason could be that the water is being abstracted and used before reaching RGS 2FC09 or that an inter basin groundwater transfer is taking place. There is a possibility of water being abstracted since the river passes through the peri-urban and urban areas of Egerton University and Njoro township, which if true reflect the extent and magnitude in which human activities are influencing the river flow regime.

Table 5.3. Annual loss of water between subwatershed I and II.

YEAR	SWSII (m^3yr^{-1}	SWSI (m^3yr^{-1})	GAINS=SWS(II-I), m^3yr^{-1}
1960	7032528	5392656	1639872
1961	2932848	51309072	-48376224
1962	9650016	20246112	-10596096
1963	24818832	34405776	-9586944
1964	3847392	55282608	-51435216
1965	3847392	1986768	1860624
1966	6401808	5644944	756864
1967	1860624	17817840	-15957216
1968	15231888	48596976	-33365088
1969	2585952	4667328	-2081376
1970	23178960	34405776	-11226816
1971	8451648	42226704	-33775056
1972	10596096	16997904	-6401808
1973	3563568	3279744	283824
1974	17691696	8199360	9492336
1975	12803616	8104752	4698864
1976	8293968	7253280	1040688
1977	10249200	41595984	-31346784
1978	28950048	26742528	2207520
1979	18890064	10911456	7978608
1980	4446576	5455728	-1009152
1981	5865696	10753776	-4888080
1982	10343808	19962288	-9618480
1983	3878928	12519792	-8640864
1984	1166832	2901312	-1734480
1985	5487264	6401808	-914544
1986	4099680	3248208	851472
1985	3090528	3090528	0
1988	851472	1608336	-756864
1989	22958208	13024368	9933840
1990	21791376	22705920	-914544

The trend of unaccounted water loss between the two subwatersheds changed after 1971 (Table 5.3 Figure 5.42). The results show that while the unaccounted loss still continued, there was a gradual increase of water passing through the lower watershed or that SWSI was losing more water through other means. No explanation for this phenomenon can be deduced from this study and data series. Future studies should concentrate on the rate of water abstraction directly from the river and from the aquifers in both the subwatersheds. This will then help establish the causal mechanism of the observed event. In addition, automatic river gauging and groundwater assessment should be initiated to eliminate the possibility of human error in data collection and processing.

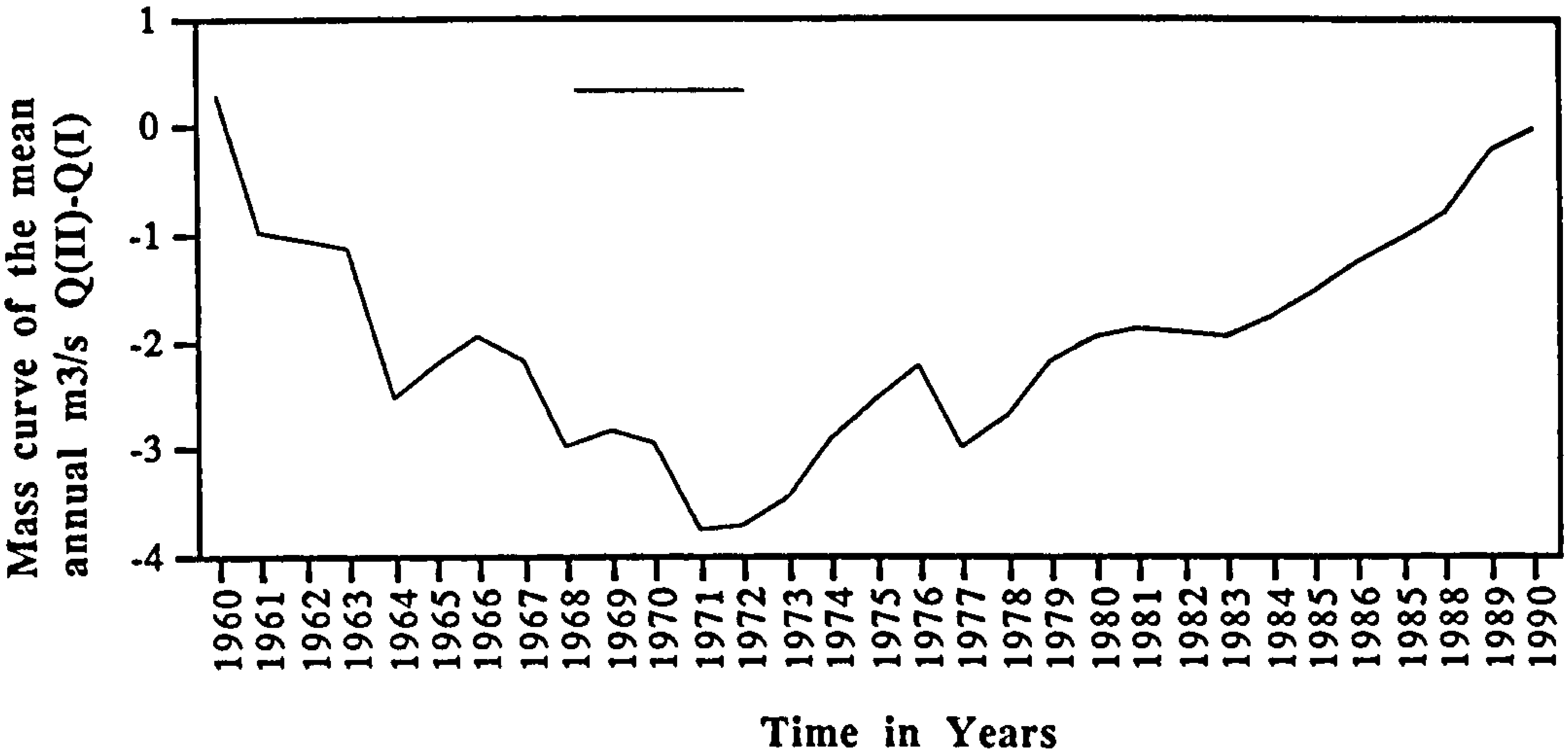


Figure 5.42. The difference in the mass curves from the two subwatersheds.

Similarly the spatial change in moisture storage in relation to inputs and outputs were evaluated by comparing the moisture storage residual mass curves for SWSI and SWSII. The residual mass curve from SWSI was represented as:

$$SUM(I) = s_o + \sum_{i=0}^{t}[\{R_i - Q_i - ETm_i\}] \tag{5.18}$$

and the residual mass curve of the runoff with respect to the mean net storage in subwatershed II as:

$$SUM(II) = s_o + \sum_{i=0}^{t}[\{R_i - Q_i - ETm_i\}] \tag{5.19}$$

These two residual mass curves were examined in the same time scale (t_0= 1960 to t_{30} = 1990). In order to estimate the amount of water lost due to unaccounted means , the residual mass curve in SWSII was subtracted from that of subwatershed I as:

$$LOSS(II - I)_i = SUM_i(II) - SUM_i(I) \tag{5.20}$$

and plotting this loss against time produced the result in Figure 5.43 shows clearly that there has been an increase in water loss. This lost amount could be attributed to several factors stated earlier as geological, abstraction, deep percolation, and inter-basin transfer. The curve however, can be split into three slopes; 1960/65, 1966/72 and a rather flatter slope between 1973 to 1988. The three periods correspond to the earlier observations of higher flows (steeper slopes) in the 1960s, and decreased flows (tending to flatter slopes) in the 1970s and 1980s.

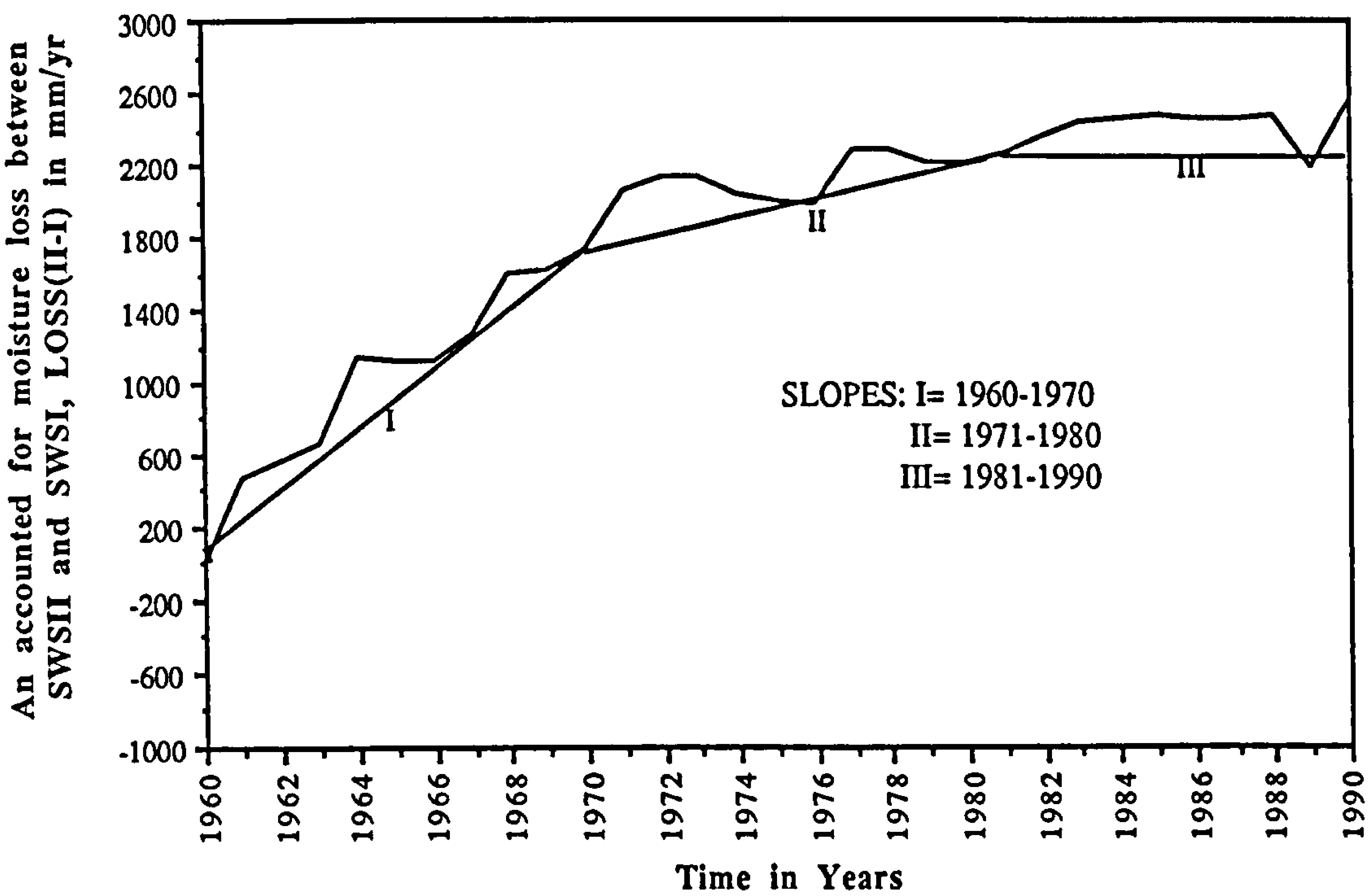


Figure 5.43. Mass curve plot of the unaccounted loss of water in the watershed.

The seasonal behaviour of the water balance components was also assessed so as to understand the seasonal regime of the river. The results of this assessment in Figure 5.44 for SWSI display two clear rainfall and streamflow peaks in April and August. A well-defined low patter in R and Q exist during the dry seasons that corresponded to the seasons of peak evapotranspiration.

To establish the drying and wetting pattern of the watershed a step by step comparison of the water inputs into SWSI is presented in Figure 5.45. With rainfall minus ETm (R-ETm), the deviation curve is positive, indicating conditions of moisture storage and hence a wetter watershed. On subtracting both the Q and ETm (R-ETm-Q), there is a clear drying regime developing in the subwatershed equivalent to roughly 100 mm per month, although the system rebounces back to a wet status in April when it peaks at a +70 mm and maintains a deficit status for the rest of the seasons. Such a pattern implies that after May of each year, the moisture recharge rate is reduced, which will most likely affect the streamflow generation mechanism and hence its regime.

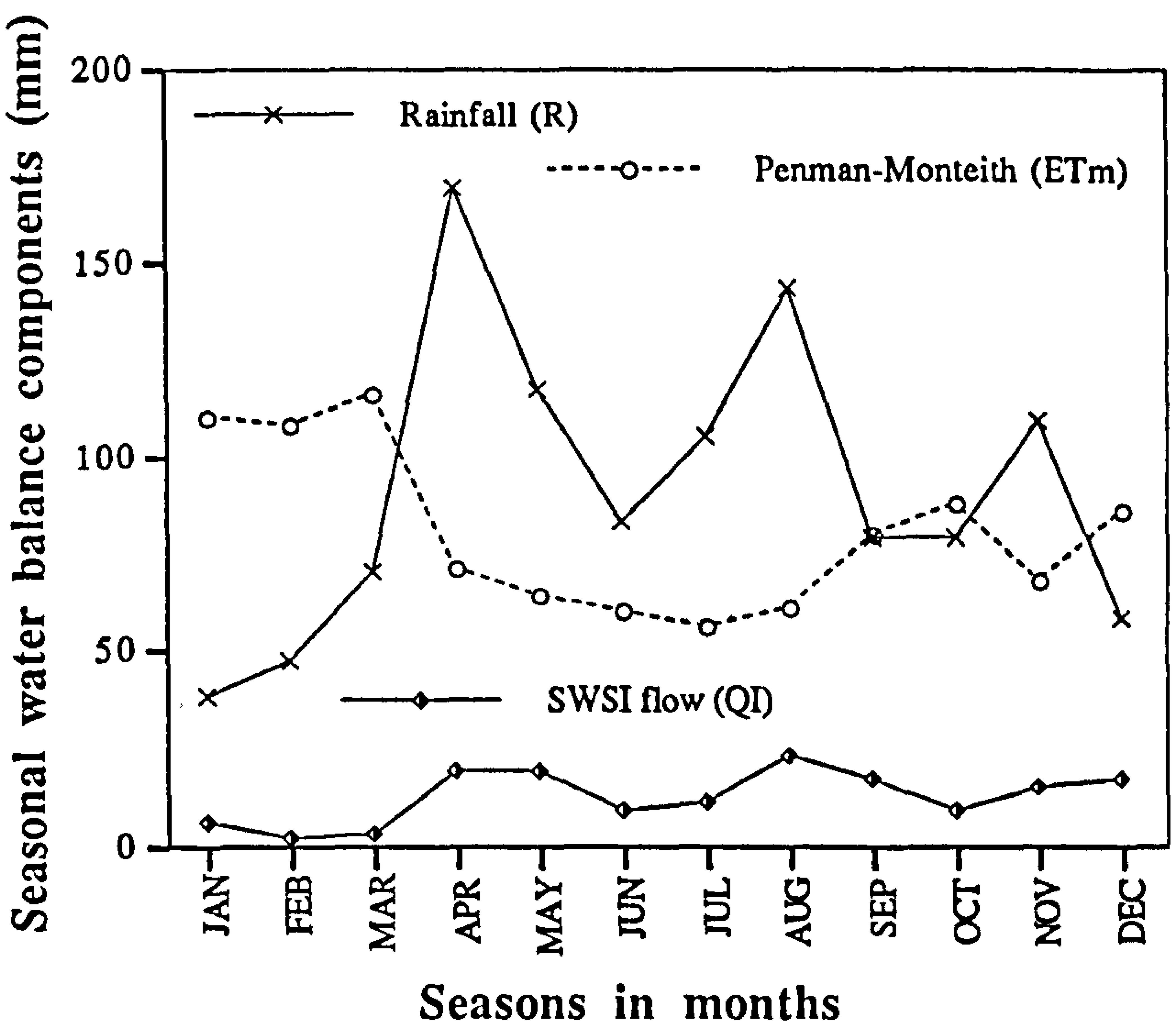


Figure 5.44. Seasonal water balance using Penman-Montieth estimated ETm

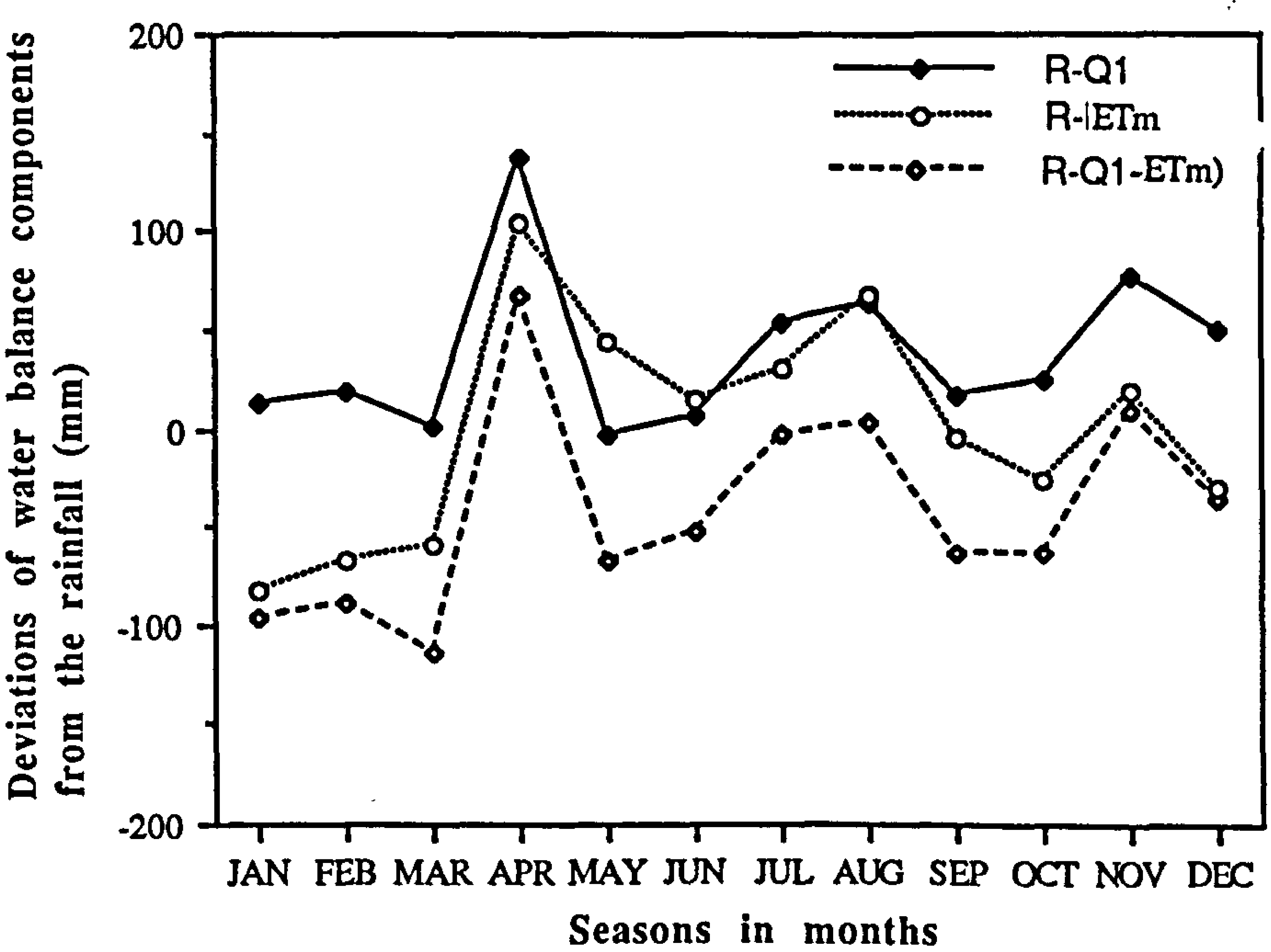


Figure 5.45. Seasonal water balance components from rainfall in subwatershed I

Spatial variations in the mean streamflow of a river are controlled mainly by the spatial distribution of rainfall, evapotranspiration, topography, temporal variations and changes in land use. This section therefore examines the relations between the rainfall- and runoff over time so as to establish the contribution of land use to the variation of the relationship. The ten-year mean flow presented as depth of flow over SWSI (125 km^2) is 182.62 for the 1960s, 113.69 for the 1970s, and 63.38 mm yr^{-1} in the 1980s.

The hydrological behaviour of a watershed is determined by examining its rainfall inputs, as modified by the gradients of the soil moisture storage, evapotranspiration (Swift, 1988) and by human-induced factors. The RRR of the flow series was therefore assessed by plotting the annual rainfall-runoff against time shown in Figure 5.46 that shows that the flow increased with increasing rainfall from a close of 150 mm, when the annual rainfall reached 1400 mm/yr. It seems that when the rainfall is <400 mm/yr, the flow reaches its lowest levels and in dry seasons, it ceases to flow.

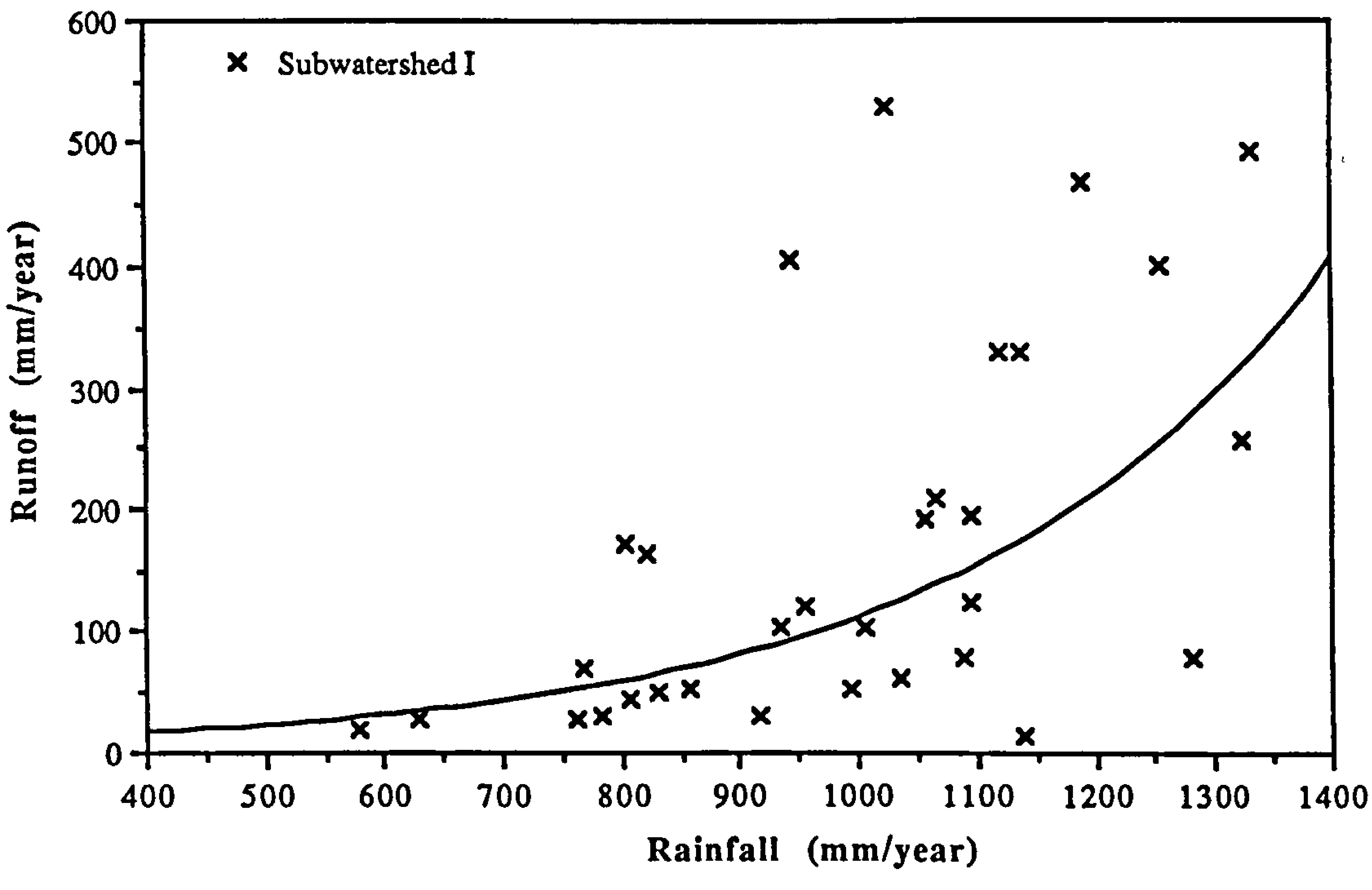


Figure 5.46. Rainfall-runoff relation in subwatershed I between 1960 and 1990.

To support this observation, trends of the rainfall-runoff partial series were produced to establish the time period during which their relationship adversely changed in magnitude or shifted away from its natural norm. This was accomplished by plotting the RRR for the subperiods (1960-1970); (1971-1980), and (1981-1990) in Figure 5.47.

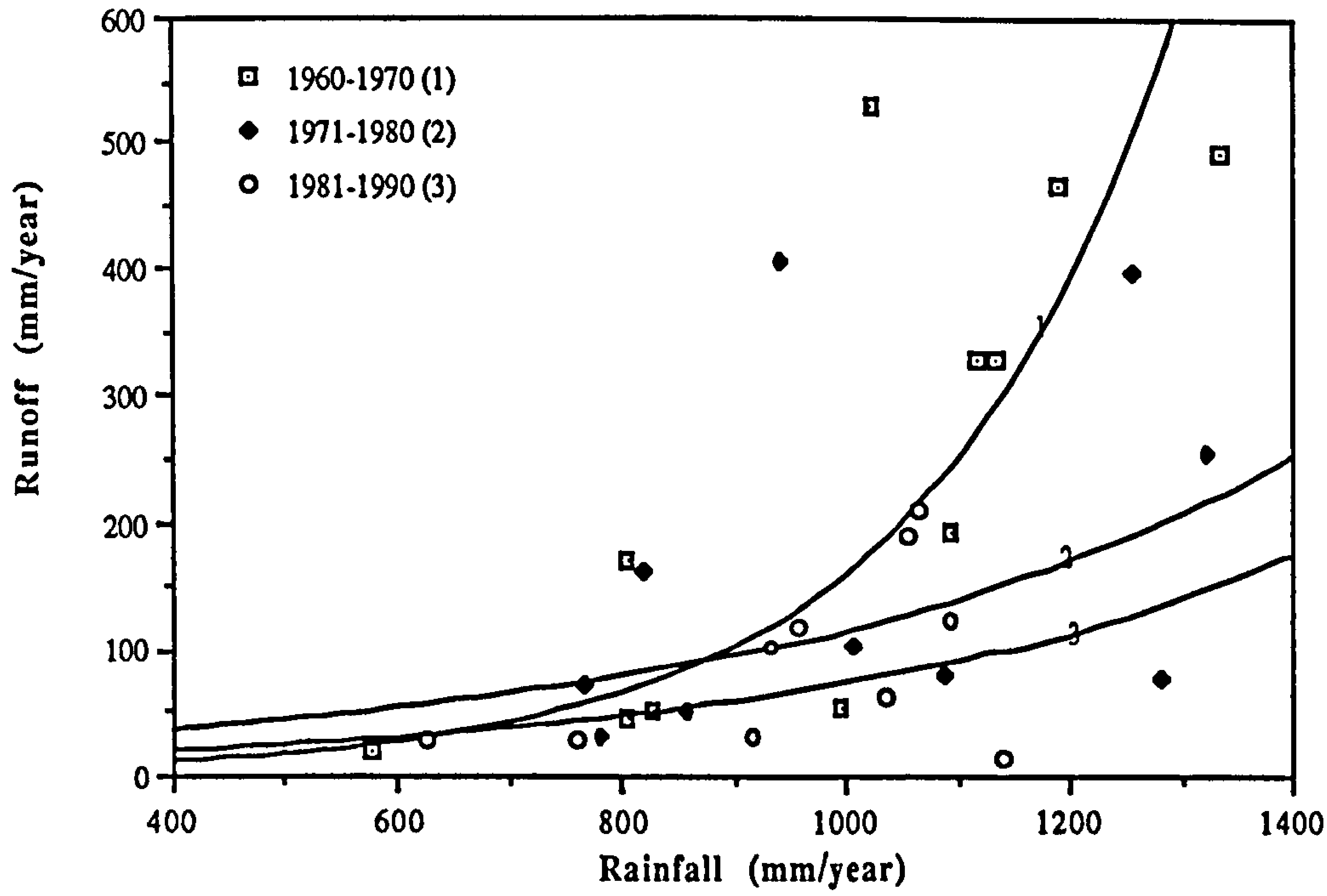


Figure 5.47. Trends of the rainfall-runoff-relations (RRR) in subwatershed I

It can be seen that the RRR in subwatershed I had a pattern of higher flows in the 1960s, then 1970s, and 1980s has the lowest flow-rainfall relationship. During the 1960s, flow >600 mm/yr was realised from rainfall amounts of only 1200 mm/yr, while in the 1970s, the subwatershed required more than 1400 mm/yr of rainfall to generate only 250 mm/yr runoff and 150 mm/yr in the 1980s. This confirms that the watershed assumed dry-up condition, so that any rain falling recharges the watershed first before issuing out as streamflow. The unavailability of moisture is attributed to the excessive human abstractions (is not possible to estimate) due to increased population and urbanisation (chapter IV) or an existence of an inter basin groundwater transfer during the 1970s and 1980s.

5.8.1. Variation of the Mean Annual Runoff Coefficients

A range of hydrological indices is used to illustrate and quantify the spatial and temporal variations of rainfall and runoff. The indices include the runoff coefficients ($\propto$), runoff relative factors (hydraulicity or runoff index), rainfall index, tests of means and variances and the coefficient of variation (Cv).

The runoff coefficient is defined as the ratio of the mean runoff to the mean rainfall (Q_m/R_m) for a chosen time scale. It is the amount (in percentage) of rainfall that contribute to the net runoff at the watershed outlet. The index was therefore used to describe the variation of the RRR in the watershed and quantify the subjective conclusions drawn in Figures 5.44 to 5.47. The results also enabled an interpretation of the causal mechanism to the observed variation and phenomena.

Streamflow, being an integrated output from the watershed, not all the amount of rain falling in an area is expected to appear at the river's outlet, instead, some is held to meet the ground water storage demand and some lost through evapotranspiration. Yet another amount is abstracted to meet human and animal water needs in the watershed. The runoff coefficient being an integrating index, should therefore reveal this change in the water balance. Using the index, it was possible to generate the time series of the mean annual runoff coefficients for SWSI given in Figure 5.48.

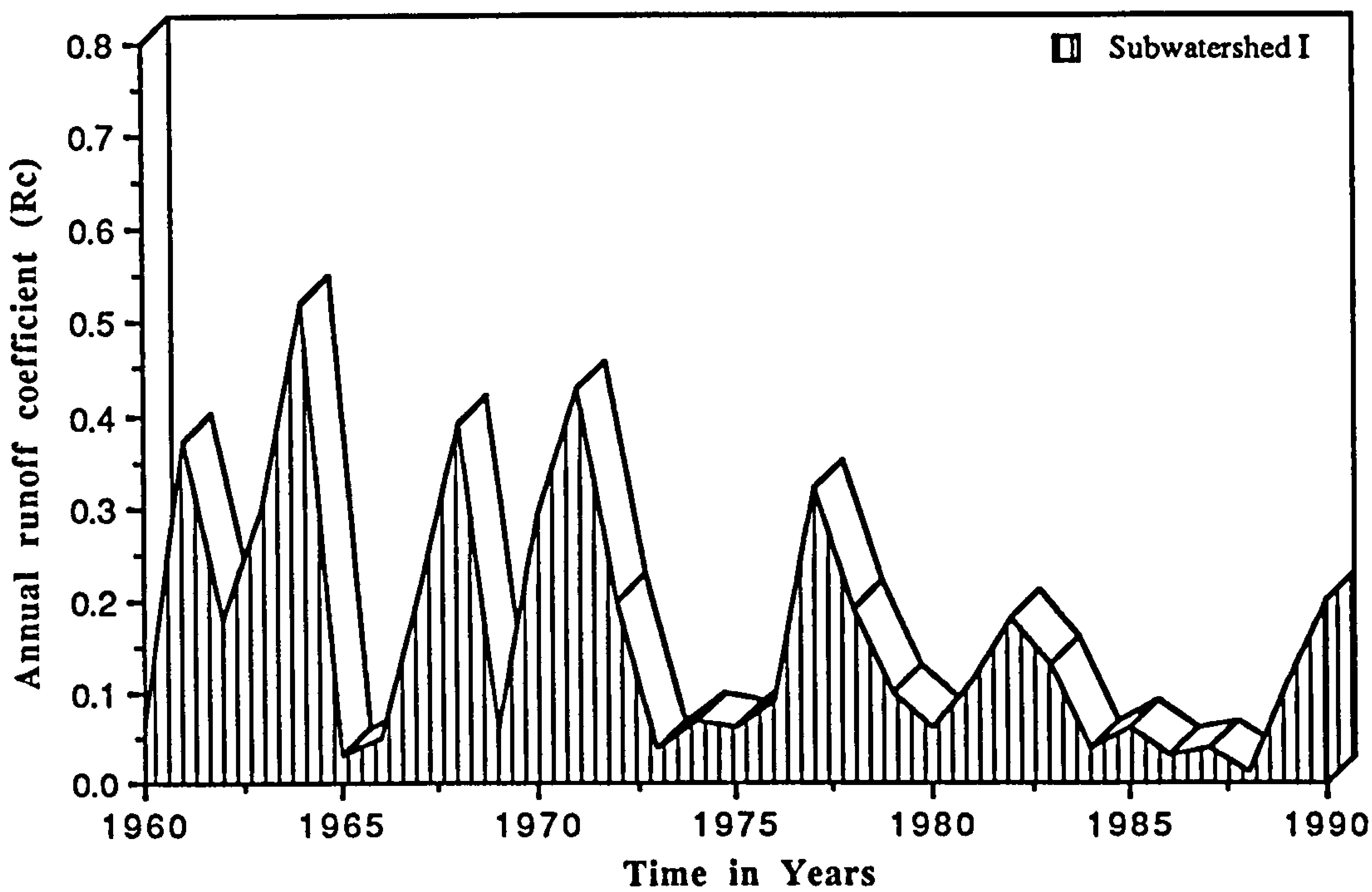


Figure 5.48. Mean annual runoff coefficients for the period 1960-1990 in SWSI

The representation shows that the annual mean runoff coefficient of the subwatershed varied from year-to-year in response to the rainfall variations and as influenced by the watershed characteristics. The conversion of rainfall into flow initially was higher from an average of 40 % in the 1960s and steeply decreasing to about 20% in the 1970s before plummeting to about 10% in the 1980s. The difference in relative conversion of rainfall may be attributed to combined changes in land use and increased water use rate by the increased population in the area. In the 1960s, the subwatershed was mostly forested and saturated so that even a small amount of rainfall was released continuously as streamflow flow. During the 1980s, the subwatershed seemed to consume any rain falling on it and therefore required a longer wetting time before contributing to the streams. Overall, the ten-year mean runoff coefficient support the decreasing trend pattern from 21.6% in the 1960s, 9.8 % in the 1970s and 8 % in the 1980s.

5.8.2. Rainfall-Runoff Residual Mass Curves

Residual flow diagrams present hydrological snapshot of a watershed for a particular flow condition. Each significant influence on a river is quantified and an overall picture of the total flow at any point in the system can be presented (Pirt, 1987). The residual mass curves (cumulative sums of the deviations from mean values) against time in Figure 5.49 were also used to examine the behaviour of the flow and rainfall at any point in time under the 1960-1990 watershed characteristics and to support findings in section 8.2. The two subwatersheds are examined for comparative purposes only.

The subwatersheds overlap each other so that water flowing through SWSI is naturally expected to contribute to the total flow output in the lower SWII. The total flow at SWSII is thus expected to increase with increasing contributing area, but results from analysing their differences in the Figure 5.42 produced negative values. This can only be attributed to the changed land use in the subwatershed from originally forestry and agriculturally to mainly agriculture and urban (Figure 4.6) with abstractions of water from the river or geological factors that can not be established at the moment.

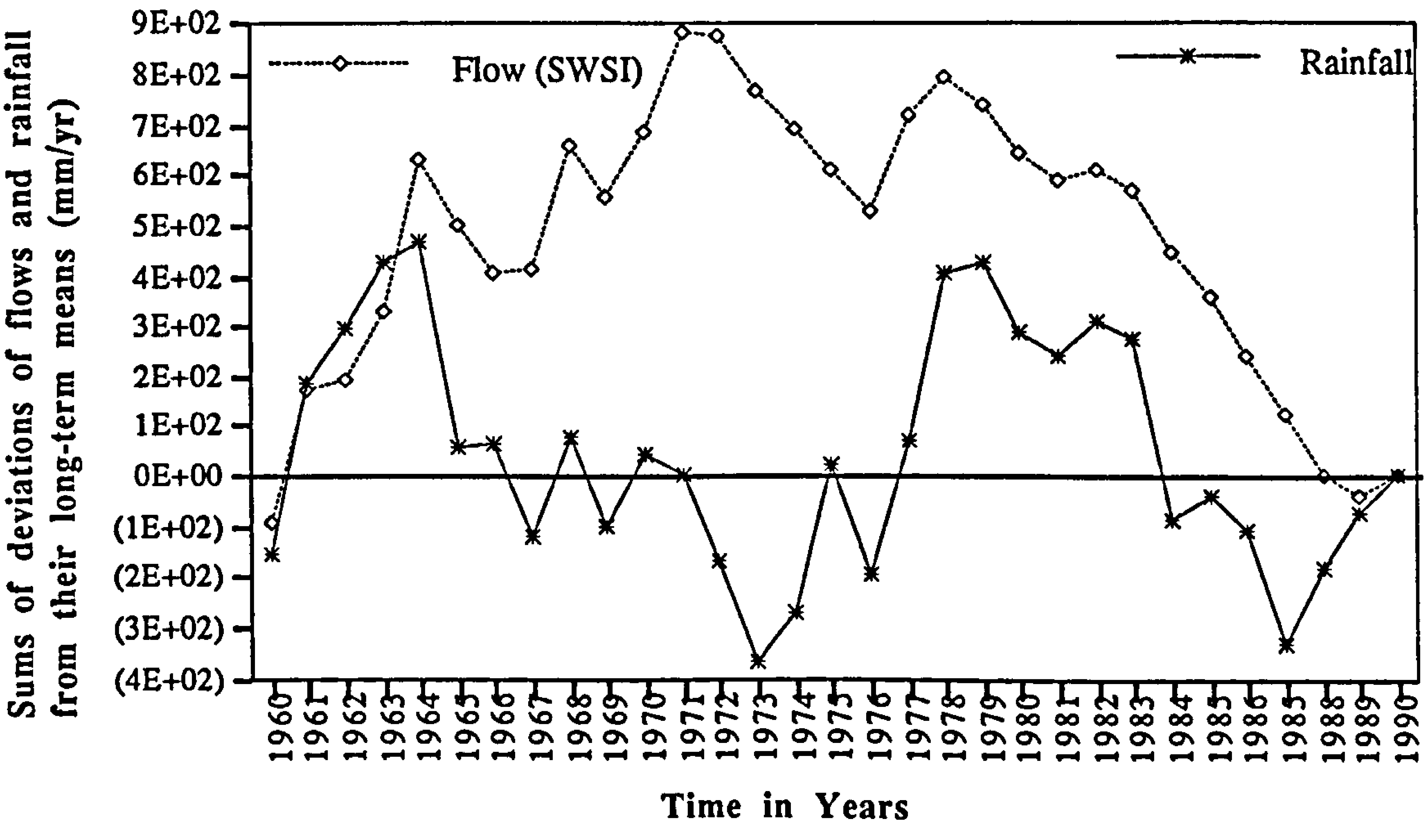


Figure 5.49. Residual mass curve patterns of rainfall-runoff in the subwatersheds

5.8.3. Seasonal Rainfall-Runoff-Relations

The mean seasonal rainfall-runoff variations in the watershed were examined to establish a long-term "cause-effect" pattern of the observed changes. The seasonal rainfall and runoff regime in Figure 5.50 present a slow response picture of the runoff to the seasonal rainfall. Being a seasonal relation, the RR-relation did not as expect reveal additional information on the long-term changes. However, since the flow-rainfall regime previously determined showed evidence of decreasing trends, the seasonal mean runoff coefficients should provide a clearer picture.

The value of the multiseasonal mean runoff coefficient was thus examined at intervals of ten years that has been used in the entire analysis. This would reveal the subperiodic variations, during which an appreciable change in the RRR occurred so as to infer the causes of the observed phenomenon. It will then explain the dynamics of the observed phenomenon i.e. from what point in time did the increase/decrease in RRR observed in the new regime become more obvious?. Does the change correspond to any changes that took place in the watershed at that particular time? or there was a time lag or no evidence of the changes. On this basis, it would be possible to attribute any observed change to other factors. As aforementioned, the search for the evidence was a step by step elimination of the causal mechanism.

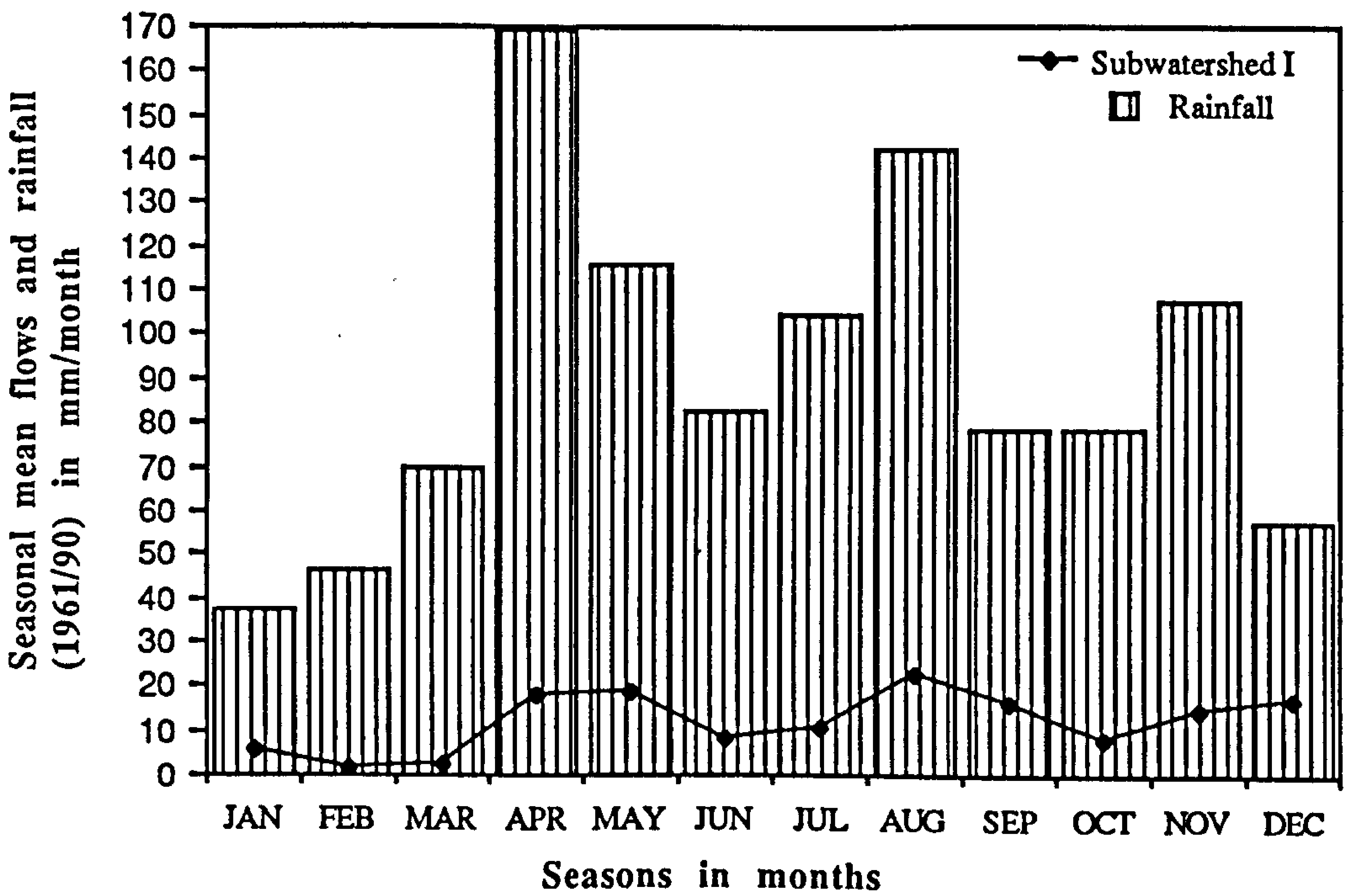


Figure 5.50. Seasonal variations of the flows and rainfall in subwatershed I

The seasonal trend of the mean runoff coefficients ($\propto$) for SWSI is presented in Figures 5.51. The conversion of rainfall into runoff is higher in the 1960s, followed by the 1970s' subperiods and the 1980s has the lowest conversion rates. This confirms the earlier determined decreasing trends in runoff despite rainfall remaining on its long-term mean value.

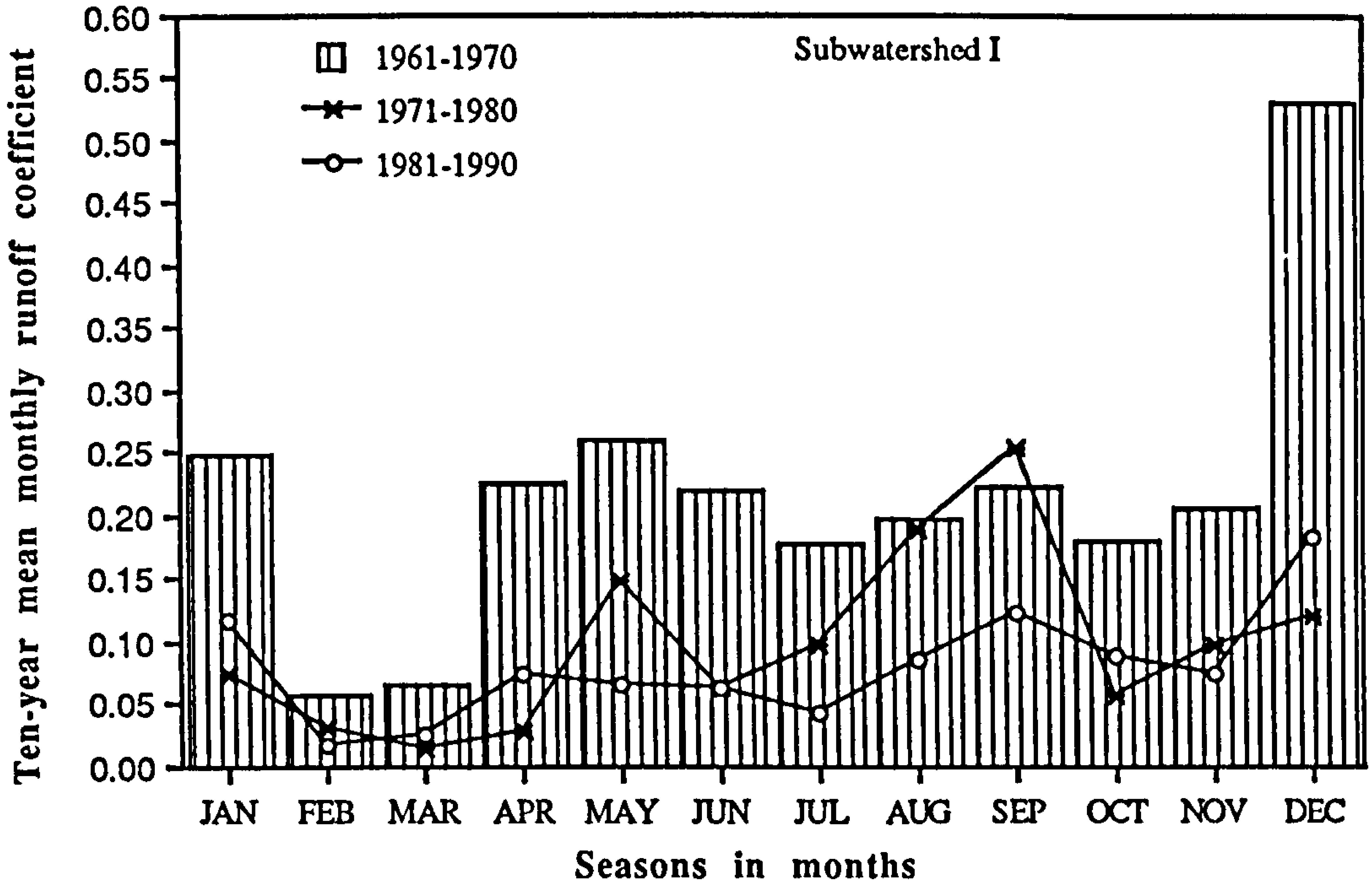


Figure 5.51. Trends in mean runoff coefficients for subwatershed I

This difference in the rainfall conversion may be attributed to the land use changes' patterns (a 20% increase in agriculture, 30% reduction in forestry and 10.4% increase in urbanisation, discussed section 4.6.2), and possible by an excessive human population (section 4.7) which may have increased water abstraction and use rate from the river basin. This generalisation and conclusions however need to be explored further in a detailed examination of the changes in the watershed characteristics and water balance components. An attempt is made to qualify and quantify the observations, and inferences in this section are dealt with in Chapters Six and Seven.

5.9. CONCLUDING REMARKS

5.9.1 Rainfall characteristics

- The mean annual rainfall did not significantly change, despite exhibiting decreasing trends from the 1960s to the 1980s. Mean annual rainfall ranged from 839.49mm/yr in the 1960s, 1074.72 in the 1970s and 810.75 mm/yr in the 1980s. The distribution of wet and dry years using the rainfall index was 1.0 in the 1960s (a wet decade), 0.99 in the 1970s (a slightly dry decade) and 0.97 (drought decade) in the 1980s. The seasonal rainfall regime in the entire watershed assumed a somewhat trimodal distribution with peaks in April, August, and November.

- The meteorologic drought occurrence assumed an increasing trend from the 1960s. One major drought in the 1960s (1965), four in the 1970s (1971, 1972, 1973 and 1976), and six in the 1980s (1980, 1981, 1983, 1984, 1986 and 1987) depicting a 10-12 year recurrence interval. The existence of droughts however is not a new phenomenon because previous findings in Ogallo ,(1978, 1978, 1981, 1982, 1984), Farmer and Wigley (1985), Lema (1986, 1990) and Smout et al., (1993) reported droughts of somewhat similar patterns in the the East African region for the last 100-years. The only difference is that the intervals and number of droughts appeared to have increased during the last three decades.

- Considering a RI > 1.20 to be a heavy rainfall year, it was possible to isolate a rather weak 9-10 year occurrence interval of heavy rainfall in 1961/62, 1979/70, 1978/79 and 1989/1990, although the 30-yr record period may be considered relatively short for such an observation. Spatially, the watershed physiography and particularly its relief influenced annual rainfall distribution and amounts. Larger rainfall amounts dominated the higher elevation areas of the watershed and decreased gradually to its lowest values in the lower elevation parts on the shores of Lake Nakuru.

5.9.2. Runoff characteristics

- The flow series responded to climatic changes in addition to human advertent and inadvertent factors. In particular, the natural variability of the rainfall was considered a major controlling factor. High mean annual flow of 182.62 mm yr^{-1} in the 1960s, decreased to 113.69 mm yr^{-1} in the 1970s, and 63.38 mm yr^{-1} in the 1980s in response to the rainfall variability. The reduction was depicted more by the very low runoff relative factors (Kis) for these periods. The 1980s may be considered a decade with exceptionally long duration of low flows.

- There was a clear difference between the flow series from the more forested SWSI and urbanised SWSII. High flow patterns was recorded in SWSI. SWSII seemed more susceptible to drought occurrences. During the unusually high rainfall years of 1978/79 and 1989/90, SWSII responded with larger peak flows than SWSI, indicating that the increased urbanised area in the watershed contributed to this pattern. Seasonal flow regime results confirmed the existence of a three-pattern (1960s, 70s, and 1980s) character in the flows series.

- Subperiodic analyses to quantify the rate of change of the mass flow curve slopes, revealed a clear decreasing pattern. The mass curve slopes dropped from a steep value of 2.53 in the 1960s suddently to 1.54in 1970s and to its lowest value of 0.71 in the 1980s in SWSI. SWII on the other hand, showed increased slope from 0.87 in the 1960s to 1.35 in the 1970s before falling to its lowest of 0.54 in the 1980s. Overall, there was a 37 % drop in slope between 1960 and 1990 which clearly confirmed the decreasing flow trend.

- The rainfall-runoff relations (RRR) established the time during which the flows deviated adversely from its natural pattern. The runoff coefficient from the SWSI also established the observed decreasing rainfall conversion trend. It ranged from mean value of 21.6 % in the 1960s, 9.8% in the 1970s, and 8.1% in the 1980s. This result indicate a drying-up sequence of the watershed. The conversion of rainfall to runoff however, seemed unstable in the 1980s, perhaps, because the watershed was excessively dry and hence required longer concentration times before contributing to the streams. As a result of this changed flow regime, the water balance, moisture availability, and water supply to the watershed consequently reduced creating water deficits and hydrologic drought regime in the 1980s.

CHAPTER VI

THE EFFECTS OF WATERSHED CHANGES ON STREAMFLOWS

6.1. INTRODUCTION

The preceding review of the subject matter and discussions have established that spatial land use changes alter temporal nature of the water balance components. Urbanisation for example, increases peak runoff, encourages progressive drying up of watersheds and finally, alters the hydrological regime of the local and surrounding areas. The main effect of changed land use was examined throughout the study, and therefore, this section endeavours to establish that this historically held scientific opinion is true for a case study catchment area in Kenya.

From the onset, the study attempted to infer changes in flow regime using various hydrological indices. Subsequent to the results and discussions in Chapter V, it is possible at this stage to draw some suggestions and observations that link land use and water balance components. These are covered in this chapter and in chapter seven respectively. The analyses and discussions in the chapter are expected to meet the requirements stated in objective number 3 and hypotheses 1 and 3.

6.2. LAND USE AND HYDROLOGICAL REGIMES

6.2.1. Observed Streamflows and Measured Land Uses Areas

After chapter V, SWSII was omitted from further analysis and research concentrated on detailed evaluation of SWSI. This was because of the similarity in hydrologic character evolving from the analysis of the two subwatersheds. It was felt that the subwatershed with more reliable data will provide adequate and reliable information to meet the stated objectives. SWSI was therefore selected because of its consistent, reliable and rather homogeneous land use, which helped the study deduce changes as they evolve. Thus the subsection investigates the rainfall-runoff characteristics as influenced by various processes governing the hydrological cycle in SWSI only.

It is possible to demonstrate subjectively, the effects of land use changes on streamflows, by compiling the results and analyses in chapter V in a single graphical representation. This would hopefully highlight the location and magnitude of the effect of land use on streamflows in the subwatershed as presented in Figure 6.1. This plot summarises the time series of the mean annual rainfall, flow series, runoff coefficient, runoff relative factors, and land use in the subwatershed. The plots were purposely grouped to establish the temporal evolution of the influence of land use on the flows. A physical examination of the trends of these variables are used to provide additional understanding of the processes. Detailed statistical analyses of their significance are carried out to confirm the observations and findings so as to support hypothesis 1.

The land use data in Table 4.12 and analysis presented in Figures 4.6 showed that historically, there was a change in land use particularly in the 1960s from conservative agriculture and forestry to urban and intensive agriculture during the 1970_s and 1980s. The graphically representations showed that during the period 1970-1990, there was a 8% increase in agricultural land, 30% decrease in forestry and 10.4% increase in urban area (including roads, towns and open grounds).

During the same period, the hydrological variables of rainfall, flow and runoff coefficient were examined to search for a corresponding change in flow. From Figure 6.1, a subjective examination of these variables reveals a decreasing trends from the 1960s to the late 1980s. Similar results were found in Figures 5.32 on the analysis of the en-year partial series of the variables. The large flows in 1960s (Figures 5.32 &5.33) was due implicitly to initial land disturbance. During the 1970s, two clear periods emerges: one of very low flow (1973-1975) resulting from meteorologic droughts (Figure 5.6 and 5.7). The mean flow during this period was below the 1960-1990 mean (< 0.538 m^3s^{-1}). The second period is the increased peak runoff (1978-1979) as a result of the 1978 heavy rainfall. After this, the flows reduced drastically in the 1980s only to recover later in 1989 and 1990. The changes are assumed to have resulted from the changed land use that combined with the reduced rainfall amounts to produce the lowest flows observed in the entire 1980s (1979-1988). The partial period's runoff relative factor in Figure 5.31 and the runoff coefficients in Figure 5.32 supports this assessment and observations.

It is however difficult to assign an exact percentage change of flow to a respective percentage change of a particular land use because, the processes themselves occur simultaneously in a discontinuous domain. What can only be suggested is that the combined effect of land use change and rainfall regime contributed the observed patterns and regime. The rainfall contribution however is contested by results from the periodic analyses and significance tests. The rainfall results in Figures 5.18, and 5.19 show indeed that, rainfall reduced in 1970s and 1980s in the upper-most parts of the watershed where SWSI is located. The decreasing trend of flow subjectively looks more or less the same during the 1970s and 1980s (Figure 6.1), unlike the clear difference shown in Figures 5.32 and 5.34. The trends in the 1970s and 1980s corresponded to the period in which land use change occurred.

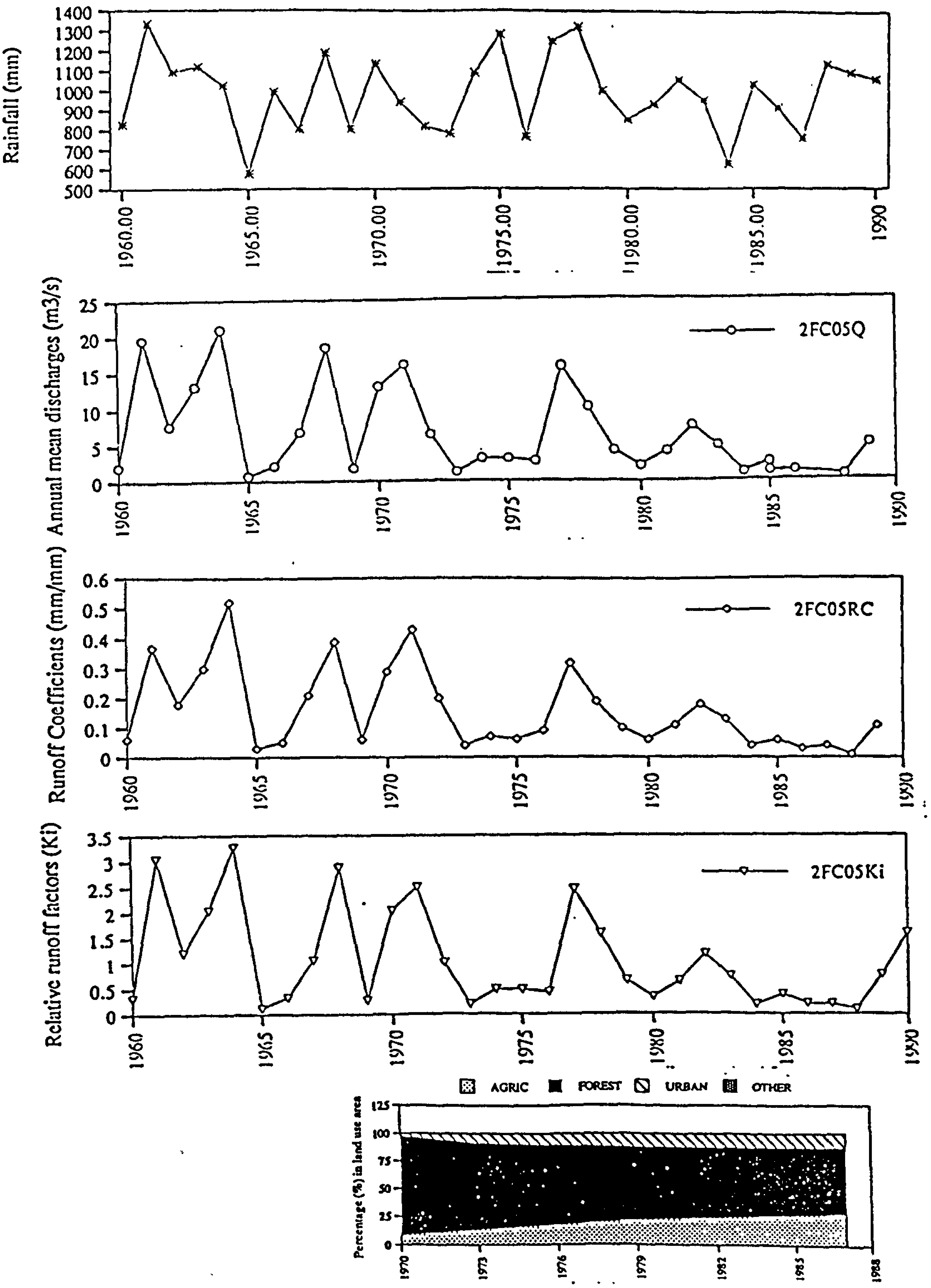


Figure 6.1. A summary of the trends of the hydrologic variables and percentages (%) in land use areas in subwatershed I

A test for stationarity and homogeneity of the hydrologic data series was done by dividing the series into a number of partial series of ten equal subperiods, and examining whether the means of the partial series are significantly different from those of the previous periods and of the entire period under study. The data series that did not pass the test were considered to be those that had experienced a periodic change, in which case, the cause can be inferred. Tests on the rainfall, flows and the runoff coefficients during the selected partial periods were carried to establish the extent and magnitude of their change using the students t-test distribution.

The students t-test provides a good test for homogeneity (UNESCO, 1987), where the null hypothesis is that the means of the individual flows are the same at all the selected partial periods, any difference can then be related to changes in land use. The t-distribution is symmetrical at default values (t_{α}=10 or 5 % at both sides of the curve). If the critical values are greater than the calculated values ($t_{\alpha} > t$), the hypothesis is rejected, and one concludes that there was no induced trend in the series and thus infer the effect of land use changed for the period under examination.

Using this test on the three hydrologic variables, it was possible to relate the historical changes in land use to the hydrologic variables. For the rainfall series, and using the 1961-1970 (1960s) as the base, for a pair of ten-year partial period (df=18), the t-values were -0.153 for 1970s, and 0.169 for 1980s. Overall, the value for the 30 year period was within the critical range of -1.734 and 1.734. This implies that, the rainfall regime did not experience a significant change during the three decades and therefore it could not have been the only and main contributor to the observed changes in the flow regime. These findings concur with the results from the analyses in chapter V and that from previous studies in Africa. For example, Sansom (1952) examined the East African rainfall over the period 1920-1949 and found a complex of increases and decreases, implying no significant trends. Studies of rainfall trends in Kenya by Ogallo (1981) and time series analysis of rainfall in East Africa (1982, 1984) found no significant change in rainfall in the region. Similar findings were reported by Rodhe and Virji (1976), Farmer (1981), Farmer and Wigley (1985) and recently in Lema (1990).

While human influence is linked to overgrazing by animals, which in turn is assumed to increase the land surface albedo, and hence subsidence and less rainfall (Otterman, 1974), there is no significant evidence to show that the same occurred in Enjoro river watershed. Evidence from the rainfall analysis however, is that moisture supply to most parts of the watershed declined gradually but not significant enough to cause the kind and magnitude of decrease in river flows observed during the 1981-1990 period.

Hence, there is no firm statistical evidence from the results to support the null hypothesis that the low flow change over the last two decades (1970-1990) in SWSI resulted from changes in rainfall trends. Previous findings and data analysis confirm that factors other than rainfall, may have caused the observed river regime. It is therefore the opinion of this study that the changed flow regime was enhanced by human-induced land use and misuse.

The t-tests on flows from the same subwatershed and using the 1960s as the base period, gave the calculated t-value of between -1.522 and 1.522 for the period 1971-1980 and between -2.579 and 2.579 for 1981-1990. The value obtained when the 1970s and 1980s are compared is between -1.319 and 1.319. These results show that the t-values of 1.522 for the 1970s is within the critical range of -1.734 and 1.734 and 1.319 between 1970s and 1980s. The 1960s and 1980s series, have significant difference with t-values falling between -2.579 and 2.579 being outside the critical range. The results suggest that, while, there is no significant difference (at 5%) between 1970s and 1980s, it exists between 1960s and 1980s. What can one concluded from this observation? Is the mean rainfall the main factor contributor to the observed difference in the flow series? In an attempt to understand this anomaly, runoff coefficients, which are a ratio of the ten-year mean period flow to the respective period mean rainfall, were subjected to the same t-test. With this, the results improved with t-values of between -1.964 and 1.964 for the 1970s and between -1.370 and 1.370 for the 1980s confirming that while there is no significant difference between the 1960s and 1970s' values, the cumulative effect of reducing rainfall is more pronounced in the 1980s.

These results can be placed into two subhypotheses: First, that there was a change in the river regime from perennial (continual flow) in the 1960s, to ephemeral (short-lived flow) in the 1980s depicted in Figures 5.47 and 5.51 respectively. Secondly, the study considers that despite the decreasing trends in the rainfall patterns, there is no conclusive evidence that attributes the observed phenomena to rainfall being the main contributor to the observed low flows in the 1980s. The discussions and analysis that follow attempts to support these hypotheses, and in the process, address hypotheses 1 and 2 respectively. The following summary can be drawn from this assessment and discussions:

•1 That the increased flows in the 1960s resulted from increased land subdivision, fragmentation and agricultural development; and the decreased trends in the 1980s shows the watershed moisture storage diminished as a result of reduced recharge

•2 The rainfall influenced flow regime during the 1960-1990 period but not significantly different from its historical pattern. Therefore the change in runoff coefficients is only attributed to other external factors such as human activities.

6.3. STREAMFLOW RESPONSE TO CHANGES IN LAND VEGETATION COVER

To gain further insight, and understanding of the hydrological processes that occurred in the watershed, the normal difference vegetation index (NDVI) was utilised. The index values derived from 1982-1990 satellite -AVHRR data base provided support to the findings from earlier analyses and observations of various hydrological indices. The NDVI value is a direct measure of the radiative response to the rigour of the surface vegetation, and so will respond indirectly to rainfall over the watershed (Liu et al., 1994), in the interception and retention of rainfall. Different thresholds of these NDVI therefore have a particular meaning about the water resources available from different types of vegetation. A relationship between rainfall, flows, evapotranspiration (ETm), and NDVI, is developed to establish the extent to which vegetation cover change influenced flow regimes at given rainfall amounts.

Does the mean annual streamflow reflect changes in NDVI? These questions were addressed and examined graphically by comparing the long-term seasonal regimes of the streamflows and NDVI values for the period 1982-1990 and presented graphically in Figure 6.4. The evolution of the NDVI values at different locations and sites in the watersheds is presented in the same plot. The sites under examination ranged from the highest elevation at Little Shuru tributary (2FC11) to the lower sites near the shore (2FC10) and bordering the Lake Nakuru National Park which cover the entire Enjoro river watershed. The plots show NDVI values decreasing with elevation, and more with the rainfall patterns. The NDVI value range from an average of 0.6 at the highest point (2300 m.a.s.l) to 0.15 at the lowest elevation site (1730m). The Rhonda area on the floor of lake Nakuru basin has been rendered bare since the late 1970s as a result of the expanding Nakuru Municipality. It is extensively eroded hence characteristized with very low NDVI values. The droughts in 1984 and 1988 for example are reflected by the low vegetation in this area with NDVI<0.20 for the whole year. The effect of the drought was enhanced by the increased urbanisation. The following observations are made on the spatial and temporal evolution of vegetation in the watershed between 1982 and 1990.

- Spatially, the seasonal variation of NDVI is >0.4 in the whole year at the upper reaches (higher elevation), and gradually decreasing to reach the lowest levels at the floor of the lake Nakuru basin, near Rhonda, an area experiencing increased urbanisation.

- Temporally, the lowest seasonal NDVI value occur between February and April, before the start of the long rains. Rains start in March but does not encourage an immediate vegetation regrowth until May, maximising by July and August, before decreasing gradually for the rest of the year.

- The flow response to the NDVI evolution is evident as they move together. The flow hydrograph however decreases and collapses faster than vegetation towards the end of the year, an indication of a poor storage characteristic of the watershed (Figure 6.2). What is interesting however, is that the estimated evapotranspiration (Figure 6.3) also decreases gently and reaches its lowest level during the high rainfall season of June-August because of low temperatures. This is also the period of a widespread and high vegetation cover. It was expected that ETp should have been higher during this season because of moisture availability and good vegetation cover to complete the transpiration process. It may however, be presumed that the averaged vegetation cover change encouraged sufficient interception, and because of the low temperatures, evapotranspiration is reduced. A second suggestion is that, since the watershed was initially dry and with the presence of a vegetation cover, there was now an ample time and environment for water to infiltrate and percolate into the groundwater storage to recharge the river continuously. This observation and suggestions are further tested when NDVI is regressed with streamflows.

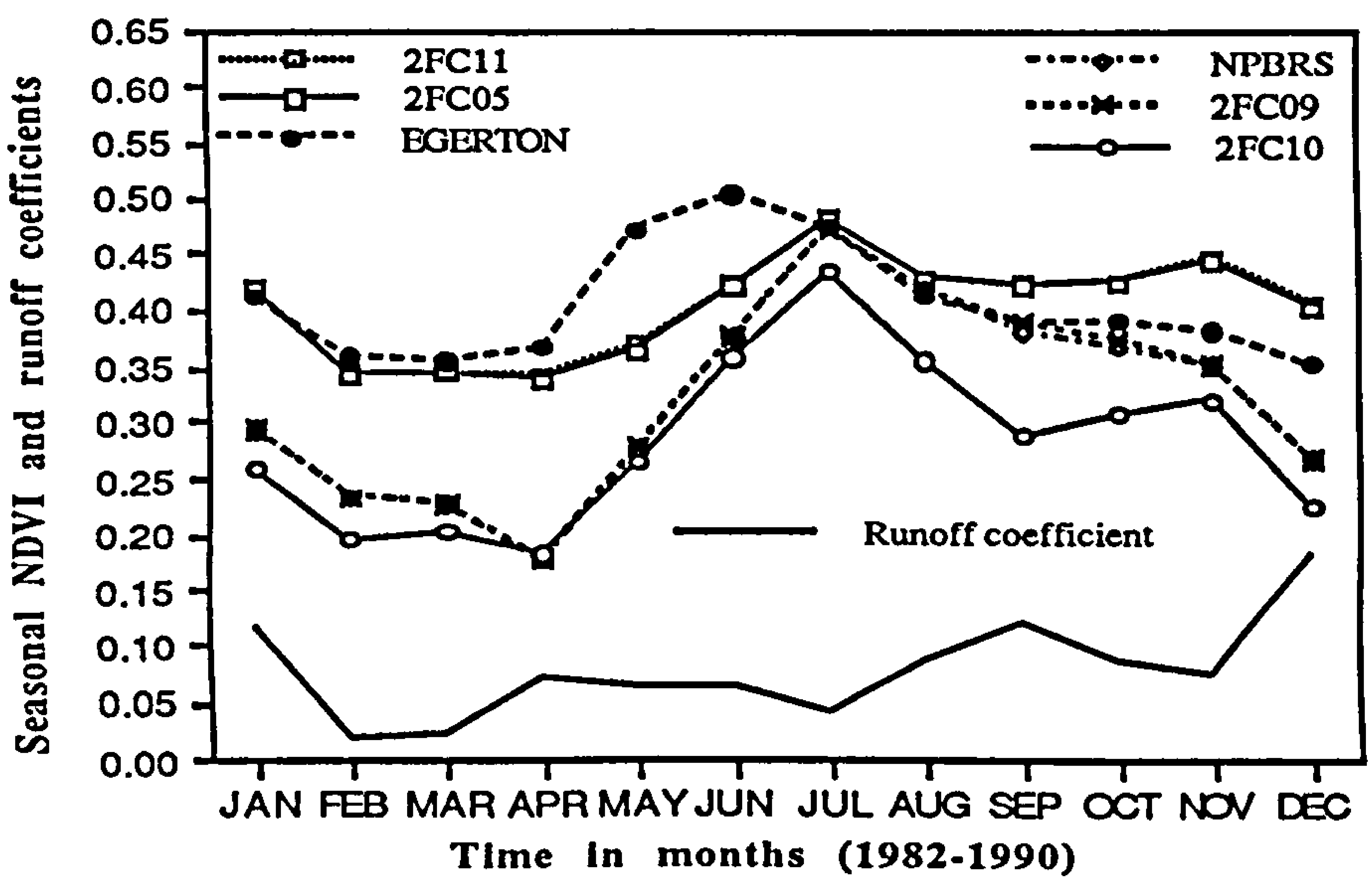


Figure 6.2. Seasonal variation of NDVI and Runoff Coefficients in subwatershed I

6.4. MONITORING OF STREAMFLOWS USING NDVI VALUES

The principal effects of vegetation cover in hydrologic systems are in the reception and disposal of all forms of precipitation through evaporation and contribution to streamflow. Alteration of the vegetation cover would therefore modify this phenomenon. These changes create differences in the albedo, which have important effects on the energy balance and hence on the water balance (Lockwood et al., 1981). Albedo investigations have been conducted, but as yet there is not an effective way to measure it remotely (Rango, 1994). Discussion of these issues has grown over the years, and particularly the consequences of the vegetation cover change on streamflows. Without a remotely sensed albedo, there is a need to use the satellite-derived NDVI as inputs to integrated response unit in hydrologic models.

Charney et al., (1975) for example, argued that the increase in surface albedo from a decrease in vegetation cover, would decrease net incoming radiation, and increase the radiative cooling of the air. Consequently, they postulated, the air would sink to maintain thermal equilibrium of adiabatic compression, associated with a suppressed rainfall. The reduction in rainfall in turn would affect plants that would lead further to decreased plant cover.

Idso et al., (1975) however, pointed out that, vegetated surfaces are normally cooler than bare grounds, since much of the absorbed solar energy is used to evaporate water, and concluded that, overgrazing and deforestation, might in contrast, be expected to raise temperatures, which would increase, rather than reduce, convection and precipitation. Model studies and experimental work by Lockwood et al., (1981) on the other hand showed that clearance of vegetation cover led to a wetter soil moisture regime and higher runoff, under a constant controlled climatic conditions.

These differing views continued into the 1980s when the use of remote sensing to monitor change in ground surface was emerging. Because of its use of reflectance of the land surface, the NDVI approach opened up opportunities for assessing these opinions. With the amount of reflected energy accounted for by this index, it was possible to determine the spatial and temporal changes in evapotranspiration and hence watershed water balances (Seevers and Ottman., 1994, Rango., 1994). This subsection therefore attempts to use the NDVI values to examine the evolution of streamflow (Q) regimes in the subwatershed that supposedly underwent extreme changes in land use and vegetation cover.

6.4.1. The Normalized Vegetation Index

Remotely sensed NDVI data can characterise vegetation cover changes because they are related to plant canopy characteristics. The relationship between the spectral reflectance and vegetative characteristics of the plants is important. Thus, the National Oceanic Atmospheric Administration (NOAA), Advanced Very High Resolution Radiometer (AVHRR) derived satellite Global Area Coverage(GAC) transformed the NDVI index in the form:

$$NDVI = \frac{[\text{Radiance (infrared band) - Radiance (red band)}]}{[\text{Radiance (infrared band) + Radiance (red band)}]} = \frac{[\{IR-R\}]}{[\{IR+R\}]} \qquad (6.1)$$

Recent works (Wiegand and Richardson, 1990; Prince et al., 1991 and Liu et al., 1994) found NDVI to be a good indicator of changes in vegetation cover. Low values of NDVI represented surfaces with less vegetation or suffering from unfavourable growth. A long term evolution of low values therefore, indicate a major vegetation cover change either through deforestation, fires and/or an extensive period of meteorologic drought. NDVI values thus can be used to estimate evapotranspiration and hence indirectly changes in Q. For example Cirhlar et al., (1991) estimated a regional potential evapotranspiration (ETp) and rainfall (P) with a good degree of accuracy with NDVI (correlation, r =0.6 and 0.8 for ETp and rainfall respectively). Thus an analysis of the temporal and spatial variations of the NDVI would provide a useful tool for monitoring climatic droughts, and human-induced changes in watersheds.

Seevers and Ottman (1994) used the index to estimate ETp and showed that when they are properly calibrated, they provide information directly related to evapotranspiration. Since changes in NDVI influence changes in ETp, it can be argued that, NDVI has an indirect relationship with Q. If such a relationship can be accurately calibrated it can also be used to monitor changes in Q due to land use changes in ungauged watersheds. The analysis of the temporal and spatial variations of these NDVI values thus would provide a useful tool for monitoring climatic and human-induced hydrologic droughts.

A conceptual examination of the Q, P, ETp, and NDVI was thus undertaken. If a relationship between P, ETp, NDVI and Q can be derived, it will be of great value in extending streamflow data and to monitor changes in flows due to changes in land use. The variables were first subjected to a correlation analysis to ascertain how each of the independent variables would affect the flow regime. The hypothesis that Q is regulated by inputs (P), outputs (ETp), and vegetation cover (NDVI) was assessed by grouping the response units as:

$$Q = f\{P, ETp, NDVI\} \qquad (6\text{-}2)$$

Assuming the aridity coefficient defined as ETp/P (Szesztay, 1979) and NDVI can describe effectively the watershed physical characteristics, equation 2 is reduced to Q= f{aridity coefficient, NDVI). A stepwise regression can then be used to develop a prediction equation relating the dependent variable (Q) to the chosen independent variables of aridity coefficient (Å) and NDVI.

The stepwise regression was used to select variables and calibrate regression coefficients. It was used to select independent variables that could be included in the final regression equation. The approach avoids irrational coefficients since the criteria used to select the independent variables eliminates independent variables with high correlation (McCuen and Snyder, 1985). The forward stepwise regression with deletion used here adds variables that maximise the total variation. At each step, those variables, that are already included in the equation are checked to ensure that they are still statistically significant. Variables in equation 6.2 were transformed logarithmically :

$$\ln Q_t = \beta_0 + \beta_1 \ln Å_t + \beta_2 \ln NDVI_t \qquad (6.3)$$

where β_0 is the constant, β_i(i=1, 2) are the partial regression coefficients, NDVI is the normalized difference vegetation index, Å is the $_i$th six-month mean aridity coefficient, Q_t is the $_i$th six-month mean flow (mm/month). Increased aridity leads to decreased NDVI. Hence the ratio of the two coefficients should describe to an extent the physical integrative nature of the watershed which is regarded as "an integrative watershed coefficient (Ic)". Alternatively, the product of the two coefficients should also provide the multiplier effect and hence represent the watershed effectively. The new coefficients given in equation 6.4 were fitted individually into eqn. 6.3 to give the integrated watershed model in equation 6.5.

$$Ic1 = \{Å/NDVI\} \quad \text{(i)}$$
$$Ic2 = \{Å * NDVI) \quad \text{(ii)} \qquad (6.4)$$

and therefore,

$$\ln Q_t = \beta_0 + \beta_1\{\ln Å_t\} + \beta_2\{\ln NDVIt\} + \beta_3\{\ln Ic_t\} \qquad (6.5)$$

6.4.2. Model Validation

Six-monthly means of the hydrological data from the watershed were fitted into these equations to establish the partial coefficients of the variables. Both the total and partial F-test were used to determine whether or not the flow was significantly related to all or either of the independent variables. The F-test null hypothesis that $\beta_1 = \beta_2 = \beta_3 = 0$ was accepted if computed F is less than the critical F{α= 0.05 and 0.01} and degrees of freedom(d,f), {q, n-q-1} where q is the number of the independent variables and n is the number of observations of the flows. If the null hypothesis is rejected (for $F<F_\alpha$}, then one or more of the independent variables is statistically related to the flow.

The National Oceanic Atmospheric Administration (NOAA), Advanced Very High Resolution Radiometer (AVHRR), global area coverage(GAC), NDVI data for the period between January, 1982 and December, 1990 were obtained from the Regional Centre for Remote Sensing and Mapping in Nairobi, Kenya. A 3-day maximum value composite procedures(Liu and Massambani, 1994) were used to eliminate the cloud contamination and atmospheric alterations. The P, Q, and climatic data were obtained from the Kenya's Ministry of Water Development, and Meteorological Department respectively. The ET_p was estimated as reference evapotranspiration using the FAO (1990) modified Penman-Monteith method

6.4.3. Discussions

The six-month mean values of the water balance components were used to estimate the stepwise regression coefficients. A correlation analysis established the association between the independent variables in Table 6.1. The correlation matrix show Q to be negatively correlated to Å, and positively to NDVI. The integrated coefficients (Ic1) and Ic2) have negative correlation (r1= -0.603 and r2= -0.267). A students t-test (df= 14 at 5% level) on the two regression coefficients, r1 and r2 indicate r2 to be more significant. On the basis of this, the Ic2 was selected to represent (Ic) in the proposed model. The multiple regression of Q and the variables indicate Ic1 to have a lower effect (R^2= 0.58) and Ic2 increased the overall fit (R^2=0.67).The time series plots of the variables selected for detailed analysis are presented in Figures 6.3 and 6.4 depict a linear trend. The predicted flows in Figure 6.3 is underestimated in all the months. However, a sensitivity analysis using different data sets, will need to be done to establish the efficiency of the model.

Table 6.1. Correlation matrix of all the variables (logarithmically transformed)

Variable	Q	NDVI	Å	Ic1	Ic2
Q	1.000	0.810	-0.475	-0.603	-0.267
NDVI	0.810	1.000	-0.556	-0.719	-0.287
Å	-0.475	-0.556	1.000	0.977	0.960
Ic1	-0.603	-0.719	0.977	1.000	-
Ic2	-0.267	-0.287	0.960	-	1.000

The constant (β_o), βi(s), and other statistical tests are given in Table 6.2 NDVI has the largest correlation value of 0.81 and is therefore selected as the first variable to enter into the equation. Its total F statistic is 29.5 and for a degrees of freedom (d.f) of (1,16), the critical $F\alpha$ (1%) is 8.54. Since $F>F\alpha$, the equation is statistically significant. The first partial F to remove is the 29.52 which is the same as the total F, hence the variable remains in the equation.

Table 6.2. Statistical tests of the stepwise regression of Q and the other variables

Variable	R	R^2	SEE	F	ß0	ß1	ß2	ß3	Se1	Se2	Se3
NDVI	0.81	0.65	0.62	29.52	7.34	6.16	-	-	1.1	-	-
Å	0.81	0.65	0.64	13.9	7.18	5.99	-0.1	-	1.4	0.43	-
Ic	0.82	0.67	0.64	9.55	7.38	21.1	14.8	-14.9	1.35	0.96	-0.97

where R is the multiple coefficient of determination, SEE is the standard error of estimation, Sei(s) the standard error of coefficients, $ß_o$ is the constant, ßi(s) are the partial regression coefficients and F is the total F-test.

The second, largest partial F to enter is 13.91 for Å. For d.f of (2,15) and a 1% level of significance, Fα is 6.36. This is significant and enters into the equation. It however, increased SEE from 0.62 to 0.64 and Se_1 from 17.86% of $ß_1$ to 23.37%. The partial F to delete are 13.91 for Å and 29.52 for NDVI, and the critical partial F value for (1,16) d.f is 8.54, hence both variables should not be deleted. Since the Se2 for ß2 of 0.43 is about 23.54 % of ß1,the inclusion of Å decreased the accuracy of the equation from a change in ßo of 7.34 to 7.18 (a 2.18% change). The last variable to enter is the I_c with a partial F of 9.55 which is greater than the critical Fα(1%) of 5.56. The variable increased ßo to 7.38 and is also significant. It is retained in the equation. Finally, at 5% level of significance, all the three independent variables are considered necessary in the final equation but with NDVI having the most effect on the streamflows. The overall change in ßo of 0.54% is relatively small. For this data set equation 6.3 thus becomes:

$$\ln Q = 7.38 + 21.1 \ln NDVI + 14.8 \ln Å - 14.9 \ln I_c \quad (R^2 = 0.67) \quad (6.6)$$

in which NDVI explains about 65% of the effect on the streamflows.

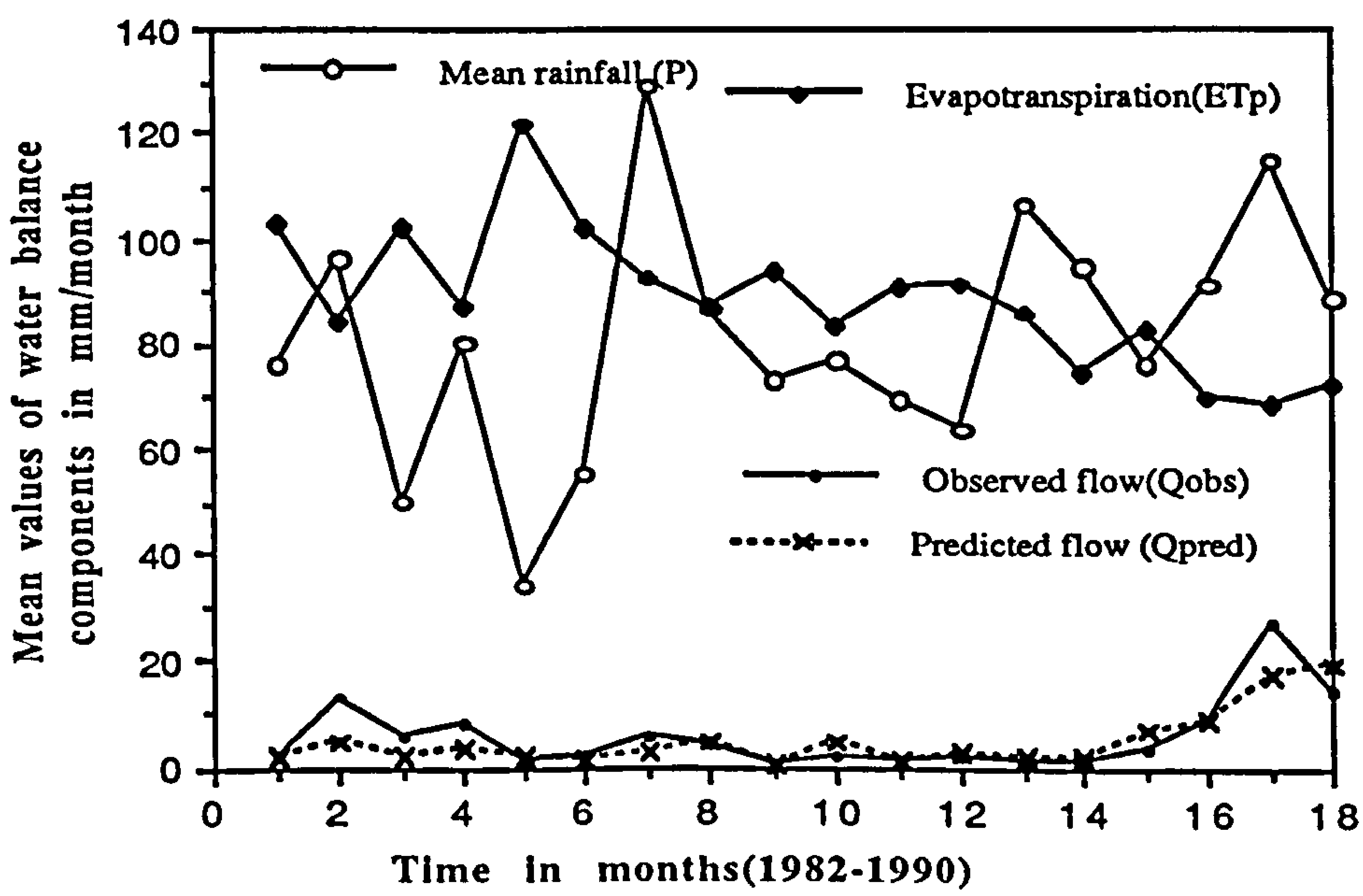


Figure 6.3. Six monthly means of the water balance components(1982-1990).

This analysis has provided an understanding of the hydrological processes in the watershed and particularly the variation of streamflows. It can now be concluded that vegetation cover changes directly influence streamflows. The high correlation value of the two variables confirms the existence of this relationship. The analysis has also shown that the decreasing values of NDVI from 1982 to 1988 (the 16-sixth month) corresponded to the periods of decreasing streamflows presumed to have occurred as a result of reduced interception, rainfall amounts and moisture storage in the watershed.

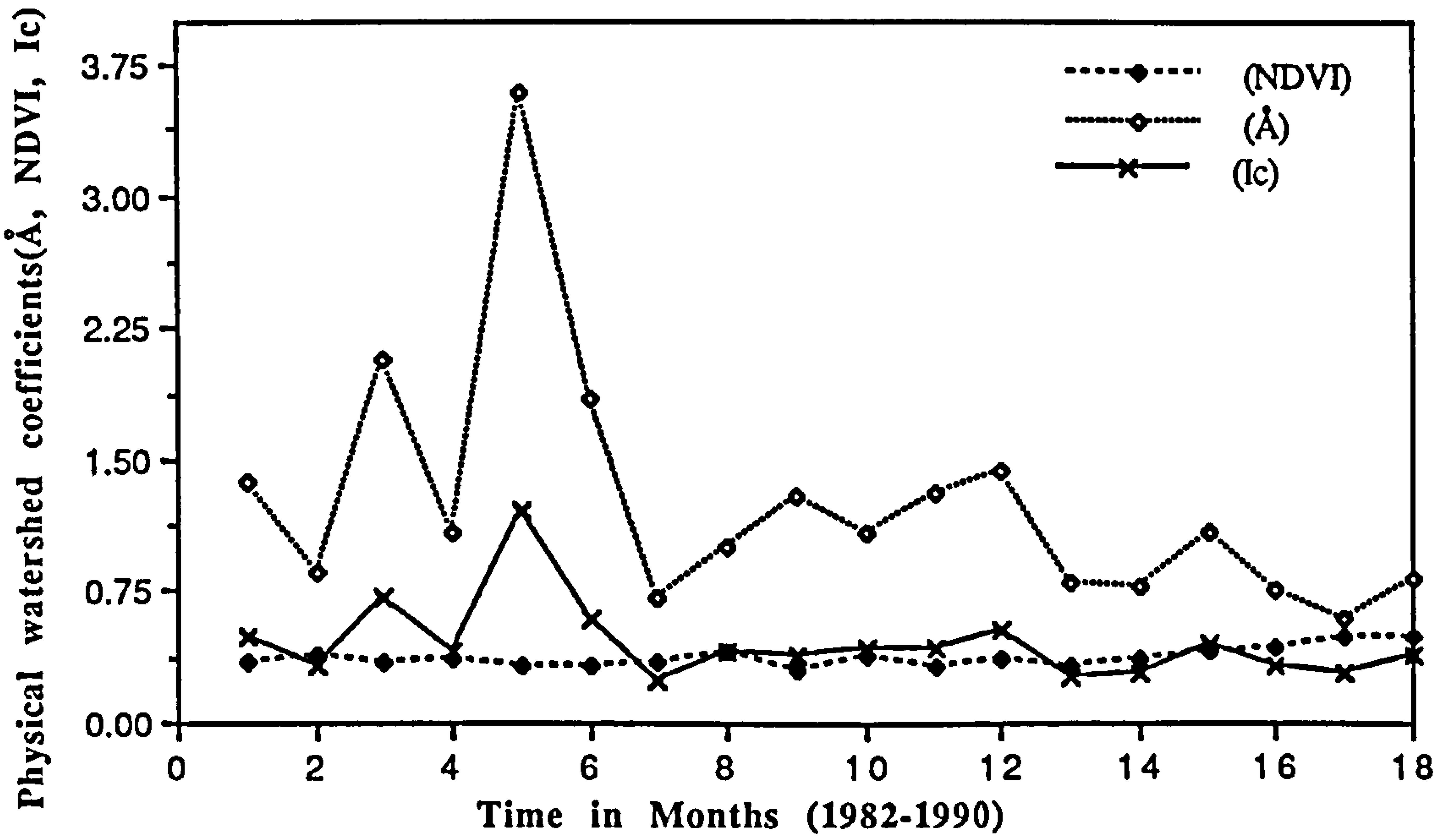


Figure 6.4. Time series of the physical watershed coefficients (A, NDVI, Ic)

In addition, the analysis suggests that moisture availability is critical to the normal growth of vegetation. The temporal and spatial evolution of the six-monthly NDVI values depended upon the amounts of rainfall. This in effect influenced the streamflow generation mechanism indirectly and reflected the changes in rainfall patterns, which indirectly affected regrowth of vegetation, and hence a reduction of the areal evapotranspiration. This feedback mechanism coupled with human actions produced long duration low flows (hydrologic droughts). Based on these results, it can safely be concluded that hypothesis 3 has to some extent been addressed and the optimisations of the physical watershed parameters using the HYRROM model in chapter VII should expound on these observations and findings.

6.5. CONCLUDING REMARKS

This chapter examined the evolution of streamflows in response to changing watershed characteristics. It became evident that subjective evaluations alone were inadequate to extract information inherent in the data so that the originally set objectives and hypotheses could be proved. As a result of this anomaly, the satellite derived vegetation index was introduced to establish the relationship existing between the watershed characteristics and the streamflows. It was felt that since the index covers the watershed areally, its effect should be regarded as integrative, and hence simulate streamflow pattern. A relationship was therefore developed between these variables which for this set of data, proved useful in relating flows and changes in land cover.

Further, the analysis demcnstrated the difficulty of reaching concrete conclusions because of the geographical nature and simultaneous occurrences of land use changes in the watershed. The 1000m drop in elevation within a 60 km horizontal distance for example, influenced the type and distribution of land vegetation cover. Secondly, the techniques used to estimate the components of the water balance components was expected to introduce errors. However, since the aim of the study was to identify periods and magnitudes in which the perceived changes in land use affected flow regime, it was possible to isolate the 1960s as a hydrologically stable period, 1970s as a period in which the effect of human activities enhanced high and low flows (increased variability) and the 1980s, as the period of extreme change in the flow pattern which developed into a hydrologic drought regime

Thirdly, a relatively simple water balance and regression equation using standard meteorological data, streamflows and vegetation cover index was proposed to estimate the availability of moisture and hence streamflows. There is a potentiality of utilising NDVI data in monitoring changes in flow regimes based on the type and nature of land use. The NDVI results are adequate to monitor hydrologic drought occurrences and changes in hydrologic regimes in small ungauged rivers. The physical nature of the equation, can be improved in the future by incorporating the NDVI derived ET_p (Seevers and Ottman (1994).

However, the findings are presently inconclusive, until errors introduced during the estimation of ETp are incorporated and tested with quality data from several rivers. The fixed value of albedo (0.23) chosen and used in the estimation ETp did not take into account the temporal evolution of the watershed reflectance, hence the likely introduction of errors of unknown magnitude. It is recommended that future estimation of ETp should make use of the remote sensing techniques to account for its temporal variation. Chapter VII that follows attempts to extract more information from the data set by optimisation the watershed physical parameters at the identified periods of changed land use and relating them to the corresponding streamflow response.

CHAPTER VII

RAINFALL-RUNOFF MODEL SIMULATIONS

7.1. MODEL SIMULATIONS IN PERSPECTIVE

The extent to which land use changes influence hydrologic regime is a perpetual inquiry which model developers have of recent times attempted to solve. As part of a continuing process, this chapter applied the Institute of Hydrology HYRROM model to simulate hydrological processes that were then used to evaluate effects of different land use changes on stream flows. The section therefore attempts to accomplish hypothesis number four and subsequently address objective five of the study.

7.1.1. Importance of Modelling

A "conceptual model" is a term used to describe models that rely on a simple arrangement of a relatively small number of interlinked conceptual elements (Jain, 1994), each representing a segment of the land phase of the hydrological cycle. The model consists of a set of rules that govern the moisture flow from one part of the watershed to other. They simplify the complexity of the real world by selectively exaggerating the fundamental aspects of the watershed hydrologic system at the expense of incidental detail. They must however, remain simple enough to be understood and used, yet complex enough to be representative of the system being studied (Anderson and Burt, 1985).

Conceptual models were initially developed to model small homogeneous areas, however, they have now been applied successfully to watersheds with varying topography and vegetation, and in areas of several square kilometres. This flexibility is readily available, but the main difficulty in using them, is the ability to choose the right type of the model to solve a particular problem. This is because of the need for the models to have some relationships to the physical reality and to provide a possibility of a multiple application, in the understanding of the natural processes and ecosystems. Unfortunately, conceptual models have the limitations of gross simplification of the very natural systems. Hence these limitations and concerns must be incorporated in the selection of the models.

7.1.2. Estimating Effects of Land Use on Streamflows Using Models

The selection of the model used in this study was based on the objective of providing a prognosis and description of the performance of the hydrologic system. The selection and methodology emphasised the dependence of the modelling exercise upon a clear definition of the problem to be solved, and upon the data base available to describe the physical system (Anderson and Burt, 1985). Objective five aimed to examine the possible adaptation and calibration of existing stochastic-conceptual models that can simulate hydrological processes and hence allow the evaluation of the different effects of land uses on streamflows. Specifically, the model requirements were:

- a physically based and lumped model with parameters and variables that could represent land use factors which may be estimated for the watershed under study.
- a model whose water balance parameters could be optimised to simulate changes in evapotranspiration associated with different levels of normalised difference vegetation indices (NDVI).
- a model which allows for an identification of periods of different runoff regimes and an optimisation of parameters for each of these regimes separately.

On the basis of these initial and basic requirements, the Institute of Hydrology, 'conceptual' HYRROM model was identified (IH, 1989). After, further consultation and authorisation, the model was selected. The model was thus used to assess the impact of watershed changes (land surface management) on flows and to establish critically sensitive parameters in the water balance (Sellers and Lockhood, 1981; Richter, Schultz, 1987, and Clarke, 1994). The use of this existing model was preferred in view of the difficulties and high cost of field controlled experiments.

These lumped models were originally reported in Nash and Sutcliffe (1970), and in Mandeville et al. (1970). They were later modified by, Douglas (1974), Eeles (1978), and Blackie (1979) and Blackie and Eeles (1985). HYRROM in particular simulated successfully flow regimes in watersheds in the United Kingdom, (Severn, Wye, Cambridge and Eynsham) and in Kenya (Kimakia and Kericho). The watersheds ranged in area from 37 to 1600 km^2 and in annual rainfall from 500 mm to 2500 mm. Their estimation gave percentage (%) error in flow prediction of -2.38% to 0.46% in Kenya (16years data), and -2.30 to 2.00 % in the UK watersheds which were excellent results (Blackie and Eeles, 1985).

The Enjoro river watershed is situated at approximate mid-distance between Kericho and Kimakia sites in Kenya. It is located at an approximate distance of about 100 km away from either of these sites. Kimakia and Kericho watersheds lie at about 0.5^o south and north respectively from the equator and at altitudes between 2000 and 2800 m.a.s.l. Enjoro river watershed is situated along the equator (0^o) at an average upper reach altitude of between 2100 and 2800 m.a.s.l. Since the HYRROM model simulated successfully flows in these sites, it provided an obvious choice to optimise the Enjoro river watershed parameters during periods of excessive land use changes. Consequently, the model was applied to flows from SWSI at the RGS number 2FC05.

7.2. THE INSTITUTE OF HYDROLOGY LUMPED MODEL

7.2.1. Description of the Model

The HYRROM model presented in Figure 7.1 describes the paths in which rainfall on a watershed naturally would follow. The model assumes rainfall to pass through the watershed and after a while , it appears as integrated flow at its outlet. It is integrated because, of its passage through various watershed physical characteristics, and experiencing varying climatic conditions. The outlet flow (Q), termed the streamflow, therefore should reflect the changing characteristics in the watershed.

The model uses the concept of "stores' to keep account of the passage of water through the system (IH., 1989). During its development, Blackie and Eeles (1985) used four 'stores': the surface, or the routing store; the interception store; the soil store; and the ground water store. Each 'store' influences the timing and magnitude of the outflow at the outlet of the watershed. The interception and soil stores also lose water by evaporation or transpiration. Other models have used 'states' instead 'of stores' with correspondingly good results (Jain, 1994). The routing concept thus does not control the efficiency of a model, rather, the depth covered in detail. The HYRROM model contains 15 parameters to be estimated and represents the watershed behaviour thus:

The incoming rainfall RAIN enters the interception storage until its contents CS reaches the storage capacity SS. Any rainfall excess, ERAIN, reaching the soil surface, is divided into two components, the 'surface runoff' and that which infiltrates into the soil moisture store. The volume of the surface runoff, ROFF, is determined by the expression ROFF = ROP x ERAIN, where ROP ('runoff proportion') is a function of the soil storage to reduce the soil moisture deficit, DC, and the rainfall intensity. The runoff proportion, ROP is represented as:

$$ROP = RC\,[e^{(-RS \times DC)} + e^{(RR \times ERAIN)} - 1] \qquad (7.1)$$

where,

RC, RS and RR are parameters to be estimated. The remaining component of the excess rainfall, ERAIN' = ERAIN - ROFF, infiltrates into the soil moisture storage to reduce the soil moisture deficit, DC. If the soil moisture storage is less than the field capacity (FC), meaning that the DC is positive, there is no drainage into the groundwater storage. If the DC is negative, water in the soil moisture storage drains into the ground water storage at a rate given by:

$$GPR = -A \times DC \qquad (7.2)$$

where A is a soil storage parameter to be optimised.
The interception storage is depleted by evaporation at a rate, ES given by, ES = FS x EO, where EO is Penman potential evaporation for the daily interval. But ES cannot exceed the storage content, CS, so that when the interception storage becomes empty (i.e. when ES = FS x EO > CS), the residual evaporative demand, EEO = EO - CS/FS is abstracted from the soil moisture storage. This storage is depleted by transpiration at a rate, EC computed as:

$$EC = FCP \times FC \times EEO \qquad (7.3)$$

where, FC is a function of the deficit, DC, given by FCP = 1 for DC < DCS and

$$FCP = \frac{(DCT-DC)}{(DCT-DCS)} \quad \text{when} \quad DCT > DC > DCS \qquad (7.4)$$

where DCS and DCT represent, respectively, the soil moisture deficits at which transpiration begins to be constrained and finally ceases. Thus, total evaporation, relative to Penman EO and to soil moisture storage is determined by the hourly/daily parameters; FS, FC, DCS and DCT.

The surface runoff store is treated as a non-linear reservoir, the volume contribution to flow as:

$$RO = RK \times RSTOR^{RX} \qquad (7.5)$$

where,
RSTOR is the reservoir content at the start of the interval. This in turn is delayed by RDEL in intervals.

The groundwater store contributes to base flow as a non-linear reservoir. In each interval, the volume output, GRO, from the store content, GS is given as:

$$GRO = (GS/GSU)^{GSP} \qquad (7.6)$$

where GSU and GSP are parameters to be optimised. The watershed output is delayed by GDEL time parameter (days). The total runoff in the time interval, t (days) is :

$$FLOW(t) = RO\,(t-RDEL) + GRO(t-GDEL) \qquad (7.7)$$

The 15 parameters whose values are estimated are: SS and FS for the interception, RC, RS, RR, RK, RX and RDEL, for the surface storage; FC, DCS, DCT, and A for the soil moisture storage; and GDEL, GSP and GSU for the groundwater storage. The permissible range of application for the nine selected parameters used in this study are given in Table 7.1. The optimisations were performed assuming dry initial soil moisture storages, so that the interception storage, CS can be assumed to be zero and DC is positive and the contents of the surface runoff, RSTOR are close to zero. GSU is computed from the initial observed flow, assumed to consist only of base flow under these conditions. This leaves the initial value of soil moisture deficit to be estimated from the analysis of the water balance components.

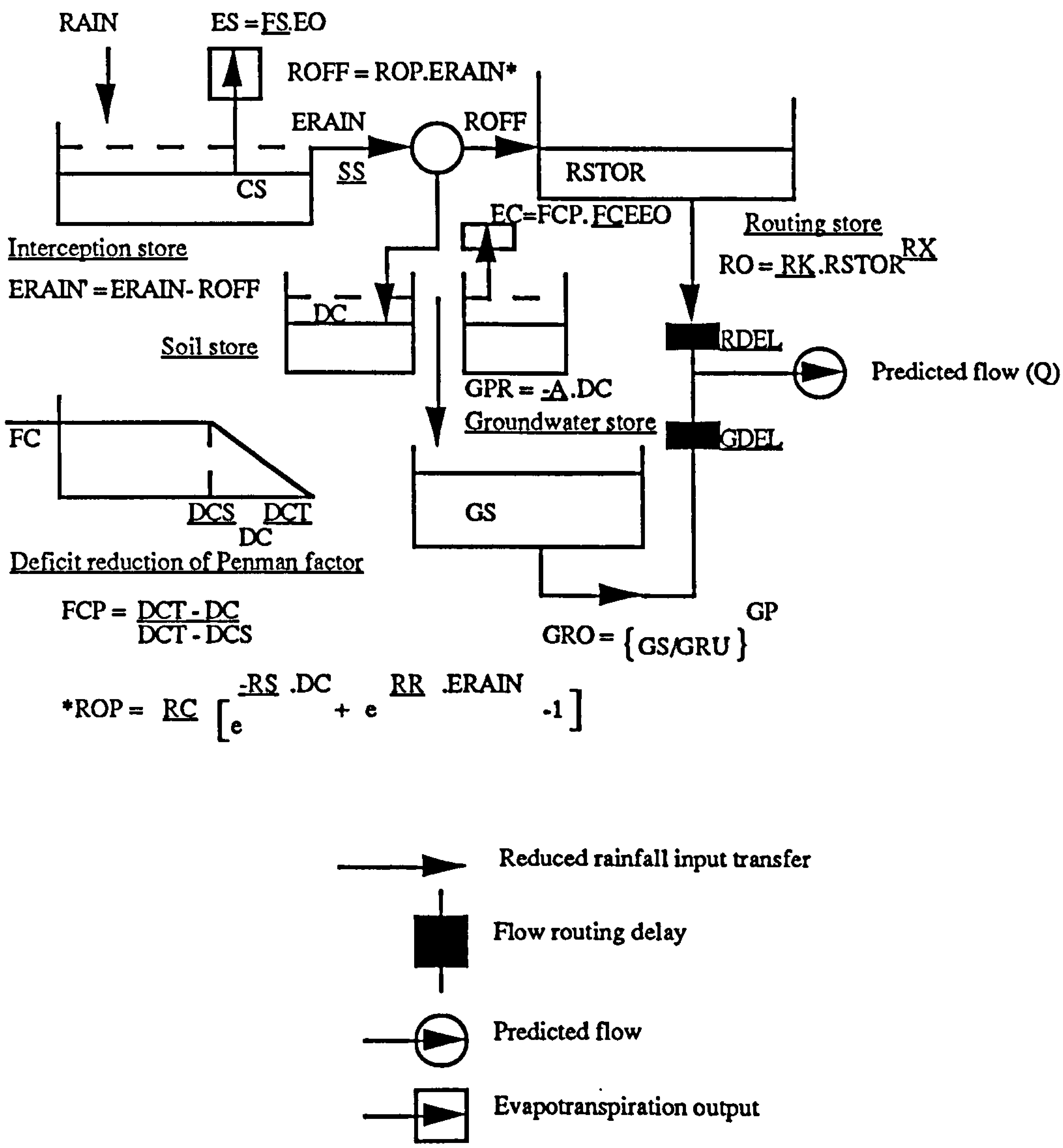


Figure 7.1. The Institute of Hydrology, lumped conceptual rainfall-runoff model (HYRROM, following Blackie and Eeles, 1985; with licence from Institute of Hydrology)

Table 7.1. The nine adjustable parameters in HYRROM (after Blackie and Eeles, 1985 and Institute of Hydrology 1989).

PARAMETER	DESCRIPTION OF PARAMETER	RANGE ALLOWED
SS	Size of the vegetation interception and surface detention store (in millimetres)	0 < X < 5
RC	Surface runoff partitioning factor	0 < X < 1
RDEL	Routing store delay (in days)	X > 0
RX	Routing store index	X > 1
RK	Routing store factor	0 < X < 1
FC	Penman open water evaporation factor	0.3 < X < 1
GDEL	Groundwater store delay (in days)	X > 0
GSP	Groundwater store index	X > 1
GSU	Groundwater store factor	X > 30

7.2.2. The model parameter Optimisation

The HYRROM parameters control the amount of flow received by, and coming from each store for each stay in the simulation. By optimising these parameters, the model can be used to represent the relationship between watershed rainfall and flow (runoff). A total of nine parameters in Table 7.1 were selected for analysis of the hydrologic regime in this study. These can be varied within the permissible and recommended range of values of each parameter (Blackie and Eeles, 1985; IH, 1989; Clarke, 1994).

7.2.2.1. Parameter Fitting and Adjustments

The data used in the analysis were not obtained from an experimental watershed. What was thus required was a conceptual approximation of the parameters. Therefore, it was necessary to iteratively adjust or optimise the selected 9 parameter values until the response of these, and indeed all the functions of the model, combine to give an acceptable approximation of the observed streamflows. The fitting techniques described in Blackie and Eeles (1985) has been followed in this study.

In brief, the estimated values of the parameters and their limits are input into the model algorithm which stores and subsequently modifies them after each run through the data set. Then, the calibration period data run into the model is introduced. The observed rainfall and evaporation, R1 and E1 are internally processed to give the computed evapotranspiration and runoff. The latter is compared with the first observed flow, Q1, in an objective test of the success of the initial parameter values. The parameter modification algorithm adjusts the initial values and these steps are repeated until the achievement of the best-fit, assessed by convergence of the algorithm, which stops the optimisation process.

When convergence has been achieved, the model is applied in prediction subroutine to a second, longer, period of input data, R2 and E2. The predicted flows are compared with the observed flows for the period of Q2, and the success, or failure, of the model is gauged from the results. If departures either, in timing or magnitude found, are unacceptable in terms of the accuracy required, then, the concepts embodied in the model are re-evaluated and modified or discarded in favour of new concepts (Clarke, 1994). Should repeated attempts at parameter optimisation and model-testing fail to produce acceptable results, it could be the case that there are errors and/or bias in or between the two data sets (Blackie and Eeles, 1985). If there is a successful comparison then the model is regarded as calibrated and suitable for the objectives of the simulation.

7.2.2.2. Parameter Optimisation Techniques

In both fitting and testing the results some objective criterion is employed to assess when the agreement between observed, Q_{obs}, and predicted, Q_{pred}, is acceptable. The choice of the objective function, F, is dependent on the type of the result required. A simple sum of squares of the residuals is used in the model:

$$F = \sum_{i=1}^{N} \left\{ Q_{pred} - Q_{obs} \right\}^2 \tag{7.8}$$

When no further reduction in F can be effected by modifying the parameter values, then the optimum fit of the model-generated flow values on the observed data set has been achieved. If a fit at low flows is required, the correct function is based on the logarithmic sum of the squares (Blackie and Eeles, 1985, IH, 1989, Clarke, 1994).

The selection of a parameter optimisation algorithm depends upon the type of the model, selected model functions and their structure, the time interval and length of historic record and its quality. It also depends upon the subjective experience of the model user on 'a trial-and-error' use of field parameters that are spatially averaged over the watershed. In all, search techniques are a form of direct search in which a parameter is changed and the result tested against observed data and the previous model outputs. Two of the most commonly used direct search techniques are those put forward by Rosenbrock (1960) and the simplex method of Nelder and Meade (1965). The Rosenbrock method is very sensitive and most suitable for sets of uncorrelated hydrologic parameters such as in dynamic watersheds. Hence it has been used in the HYRROM model (Blackie and Eeles, 1985 and Clarke, 1994).

7.3. RAINFALL-RUNOFF OPTIMISATIONS

The effects of human-induced changes on the river's hydrologic regime were examined by optimising nine parameters in HYRROM. Four dry (1965, 1973, 1984, 1986) and four rainy years (1964, 1970, 1978, 1989) were considered separately in order to locate and assess extreme periods and magnitudes of perceived human impacts. The subperiods were chosen with a view to establishing the range of land use change in the 1970s, and 1980s and particularly (i.e. different impacts, forested and urbanised parts of the watershed) in 1970, 1973, 1979, and 1987 in which measured land use area values are available. It was not possible to optimise the parameters over longer periods (say at subperiods of 5-year intervals) because of discontinuous data sets, hence the calibration at selected discrete years.

The sequence of three analyses is presented in Table 7.2. In the first stage, all parameters except the interception parameters were held constant, and the interception parameters were optimised. In the second stage, all parameters except the soil store parameters, were optimised. Similarly in the third stage, all parameters except the groundwater and routing parameters were optimised. This segmentive iteration was necessary to reduce the large number of parameters and to establish the rate of parameter change

Table 7.2. Optimisation and sensitivity analysis sequence adopted.

Optimisation order	Analysis 1	Analysis 2	Analysis 3
1	FC	RX	GDEL
2	GSP	RK	FC
3	RC	RDEL	GSP
4	SS	GSU	RC

To investigate the impact of changes in land use and watershed characteristics, on the optimal model parameter values, the model parameters were derived at different subperiods (1970, 1973, 1977 and 1987, for measured land use) in addition to the selected wet and dry years. The results of the rate of parameter change are given in Tables 7.3-7.7 and a summary of the optimised parameters are presented in Appendix E.

To examine the evolving character and trend of the optimised set of parameters, the time series of the parameters, the historical land use and the rainfall regime in the subwatershed were analysed. To objectively evaluate these parameters, some years portraying similar parameter patterns were isolated. The years blocked in Table 7.3 seem to have similar patterns while the underlined ones assume different behaviour. On examining the rainfall regime during these selected years, it was found that, the blocked years were actually the years which had the lowest rainfall below the 1960-1990 mean, and the underlined years had experienced exceptionally high rainfall. Several other years could fit into this categorisation, but it was not possible to identify. Sensitivity analysis of the parameters was done to isolate hidden information. Obvious and striking years were selected for further examination to establish how much is their behaviour attributable to land use and how much to the rainfall regime or a combination of both.

There is a distinct pattern of some parameters representing the three stores in Table 7.3; the interception (SS), the surface runoff delay (RDEL) and the groundwater streamflow generation delay (GDEL). During the dry years (blocked), the SS assumed the lowest value of 0.0, the time taken by rainfall to contribute to runoff (RDEL) varied widely, and the groundwater generation delay takes the longest time as compared to the other years (GDEL ranged from 0.75 days in 1965, to 5 in 1984, an increasing time trend into the 1980s). During the wet years (underlined), the store parameters are also distinct. The interception (SS) increased to its maximum limits, (among 0.491 in 1964, 4.448 in 1970, 5.0 in 1979 and 4.952 in 1989) , the RDEL was relatively low from 0.0 in 1972, 0.001 in 1978, 0.487 in 1979, and 0.0 in 1990. The value was low during years preceding high rainfall year, while the GDEL assumed low values' 0.006 in 1964, 0.005 in 1970, 0.009 in 1978 and 0.190 in 190. A summary of parameter values from these selected years is given in Table 7.4.

Table 7.3. Time series of the optimised parameters for Subwatershed I (1964-1990)

	PARAMETERS										
YEAR	SS	RC	RDEL	RX	RK	FC	GSU	GSP	GDEL	F	%ERROR
1964	0.491	0.885	0.439	2.637	0.446	0.750	107.56	5.686	0.006	1.846	1.51
1965	**0.00**	**0.021**	**0.594**	**2.474**	**0.060**	**0.999**	**108.50**	**1.00**	**0.75**	**0.038**	**-1.30**
1966*	0.459	0.107	0.641	2.037	0.117	0.493	889.57	1.000	0.649	0.138	0.83
1967	2.616	0.160	0.299	1.750	0.127	0.596	135.27	1.000	0.250	0.943	7.84
1968	0.029	0.487	0.506	1.000	0.378	0.583	629.91	1.007	0.000	2.870	-4.59
1969	0.057	0.029	2.758	1.000	0.130	0.971	231.49	1.005	0.500	0.124	0.50
1970	4.448	0.093	0.413	1.726	0.158	0.997	104.89	3.708	0.005	0.971	-0.56
1971											
1972*	**0.272**	**0.001**	**0.000**	**10.72**	**0.000**	**0.797**	**93.028**	**27.64**	**4.82**	**0.098**	**-0.93**
1973	**0.00**	**0.045**	**0.508**	**2.626**	**0.006**	**0.998**	**9005.9**	**3.54**	**14.448**	**0.121**	**-0.14**
1974*	**0.001**	**0.000**	**1.997**	**1.000**	**0.037**	**0.999**	**385.54**	**1.00**	**2.234**	**0.422**	**0.85**
1975*	0.054	0.045	1.794	1.000	0.187	0.427	760.78	1.004	5.007	0.103	5.95
1976	**0.023**	**0.239**	**1.939**	**1.066**	**0.042**	**1.000**	**903.02**	**1.003**	**0.002**	**0.369**	**14.54**
1977											
1978	4.784	0.559	0.001	1.001	0.001	1.000	903.99	7.113	0.009	0.771	2.77
1979*	5.000	0.298	0.487	2.986	0.096	1.000	90.400	16.957	0.719	1.490	9.50
1980*	**0.014**	**0.146**	**1.360**	**1.000**	**0.024**	**0.923**	**456.81**	**1.001**	**2.147**	**0.133**	**-0.70**
1981*	4.184	0.337	0.200	2.600	0.100	0.999	90.571	16.571	0.500	1.641	18.20
1982											
1983*	**4.980**	**0.055**	**0.797**	**1.039**	**0.225**	**0.999**	**107.56**	**3.148**	**0.001**	**0.620**	**4.32**
1984	**0.001**	**0.001**	**1.997**	**1.000**	**0.037**	**0.999**	**385.54**	**1.000**	**2.234**	**0.422**	**0.85**
1985*	4.996	0.042	0.699	1.000	0.212	1.000	896.69	1.144	0.977	0.295	5.98
1986	**1.699**	**0.015**	**0.200**	**2.600**	**0.100**	**0.997**	**90.400**	**1.001**	**0.500**	**0.180**	**-0.92**
1987	0.028	0.022	0.200	1.748	0.073	0.997	473.42	139.64	1.035	0.500	1.45
1988*	**0.000**	**0.017**	**0.077**	**2.196**	**0.103**	**1.000**	**78.057**	**1.002**	**3.386**	**0.006**	**0.64**
1989*	4.952	0.082	0.497	1.001	0.254	0.999	339.72	1.009	0.898	0.515	-0.32
1990	2.166	0.016	0.000	1.000	0.045	0.818	255.42	14.789	0.190	0.985	5.90

The clarification of the observed pattern is necessary to understand how the watershed responded to changing climatic regimes and particularly to changes in land use areas in 1970, 1973, 1979 and 1987. A possible rationale would be that as the dry-years develop into droughts, the vegetation growth is retarded which in turn decreases the SS parameter value. When the rains re-occur and continue for a longer duration, there is a re growth of vegetation that increased the SS parameters during the wet years.

Similar reasoning and approach is used to assess the behaviour of the other 'stores'. The RDEL is the parameter representing the time taken by surface runoff to meet the evaporative and surface storage demands before contributing to the streamflow. On the basis of this function, it suffices to presume that during the dry years, the moisture beneath the soil surface is rapidly used to meet the evapotranspirative demand, creating drier conditions on the soil surface. Hence precipitation that falls immediately after a dry year is utilised to meet this demand, before it contributes to streamflows.

Consequently, the contribution of the surface runoff as streamflow is delayed which can be interpreted from the RDEL time delay. The RDEL parameter in most cases however, maintained an inconsistent trend. What could be the reason for this behaviour? Assuming that during these dry years, the ground surfaces were relatively rendered impervious and because of the high intensity short-duration nature of rainfall in the tropics, an instant surface sealing would occur which would create less storage space and equally low infiltration rates. Thus the time taken for the surface runoff to contribute to the streamflow is increased as seen in the data, and gradually reduced so as to reach the levels of saturated watersheds. What must be established however, is whether the impervious surface resulted from a lack of rainfall, which reduced vegetation growth or because of human interventions on the ground surface which altered the soil-plant-water continuum?.

Further examination of the other 'stores' may draw insights of what is happening. If an assumption is made of a negligible inter-basin transfer, the contribution of the groundwater storage to the streamflow is mainly governed by the antecedent soil moisture conditions, the recharge from rainfall input and a possible inter-basin recharge. It would therefore imply that the GDEL parameter will increase if the recharge rate is reduced. This then reduces the ground water tables which in turn increases the time delay and streamflow release rate.

Alternatively, the GDEL is decreased if the groundwater storage was already saturated, suggesting that water infiltrating and percolating into this 'store' is instantly released as streamflow. The result discloses that during the dry years, GDEL increased from 0.751

in 1965, 3.541 in 1973, 2.234 in 1984, and 3.386 in 1988. A generally increased pattern of the RDEL from the 1960s to the 1980s is observed, indicating a continuously lesser contribution of the groundwater to the total outflow and hence a drier scenario was evolving. During the wet years, GDEL times were relatively shorter ranging from 0.006 in 1964, 0.005 in 1970, 0.009 in 1978 and 0.190 in 1990. Longer time values only occurred in those wet years preceding a drier year. It thus confirms that the watershed was responding favourably well to the changing conditions during these periods. The response mechanism of the watershed to the rainfall however was markedly different and unstable during the drier years, especially in the 1980s, when all the parameters fluctuated between extreme largest and lowest value.

A possible explanation of the observed extremes of parameters as has been postulated in the previous chapters and sections is the effect of human interventions. Although a specific anthropogenic contribution cannot be independently isolated, it can safely be assumed that the human activities in terms of increased human population, changed land use and other climatic factors combined to create the hydrologic condition in the 1970s and 1980s. Further, a model performance analysis is performed to establish the most likely cause of these effects and to identify which of the model parameters are more sensitive to land use changes and which to the rainfall regime.

Table 7.4. Changes in parameter values as a function of rainfall regime in SWSI

Parameter	Selected driest years				Selected wettest years			
	1965	1973	1984	1987	1964	1970	1978	1990
SS	0.000	0.000	0.001	0.028	0.491	4.449	4.784	2.166
RDEL	0.594	0.508	1.997	0.200	0.439	0.413	0.001	0.000
GDEL	0.751	4.992	2.120	1.035	0.006	0.005	0.009	0.190
Obj.Function	0.038	0.127	0.422	0.500	1.846	0.971	0.771	0.985
% ERROR	-1.30	-2.97	2.960	1.450	1.510	-0.560	2.770	5.900
Rainfall index	0.580	0.790	0.640	0.770	1.040	1.150	1.340	1.140
Runoff Coeff.	0.030	0.040	0.040	0.040	0.520	0.290	0.320	0.200

where Rainfall index is the ratio of the ith year rainfall to the (1960-90) mean and the runoff coefficient is the ith year flow to the ith year rainfall

The results for the selected wet and dry years were plotted against time in Figures 7.2 and 7.3 to examine changes of the parameters in response to rainfall and runoff regime. The plots confirm earlier observations and findings on increasing and decreasing trends of parameter values in response to changing rainfall regime and possibly watershed characteristics. Results from the analysis of NDVI, RI and model 'store' parameters in Figures 7.4 to 7.7 also support these findings.

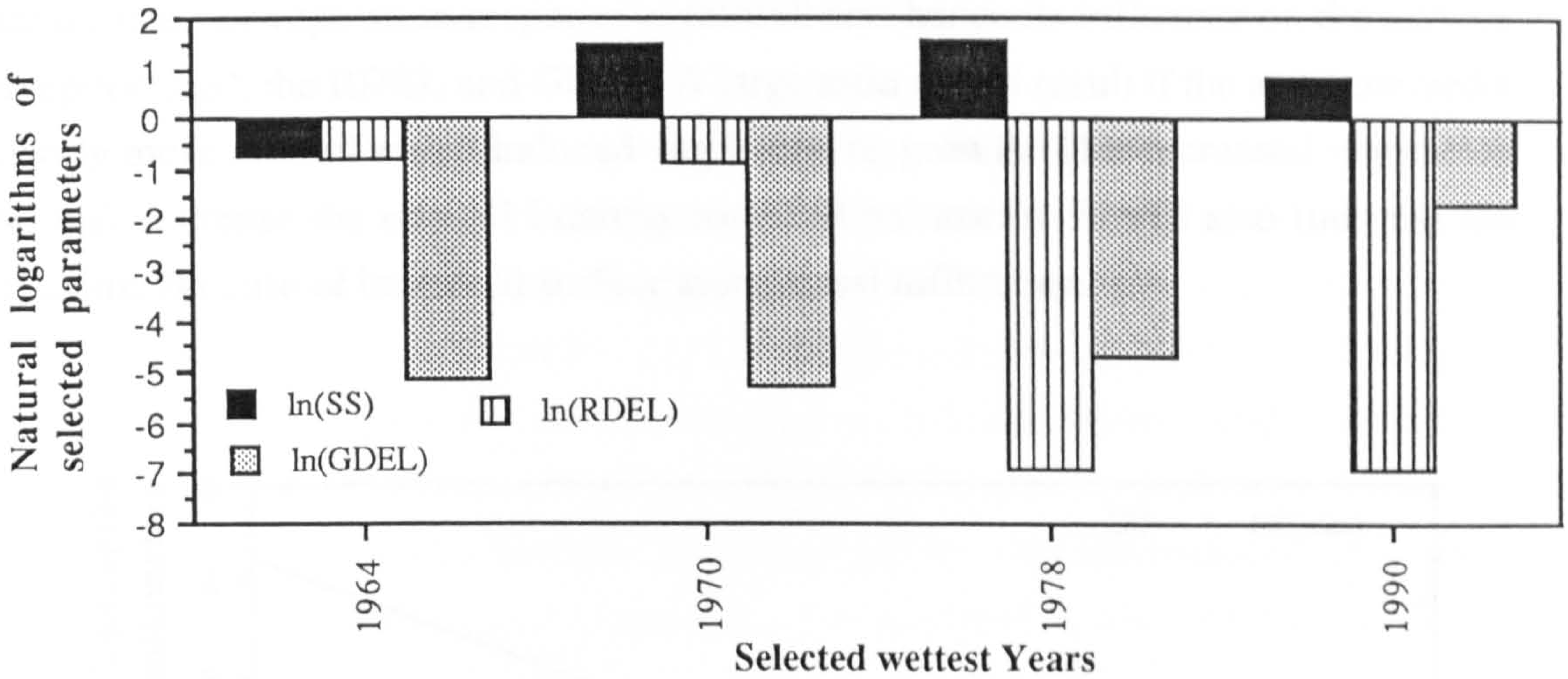


Figure 7.2. The time series plot of selected parameters during the wet years (1964, 1970, 1978, 1990)

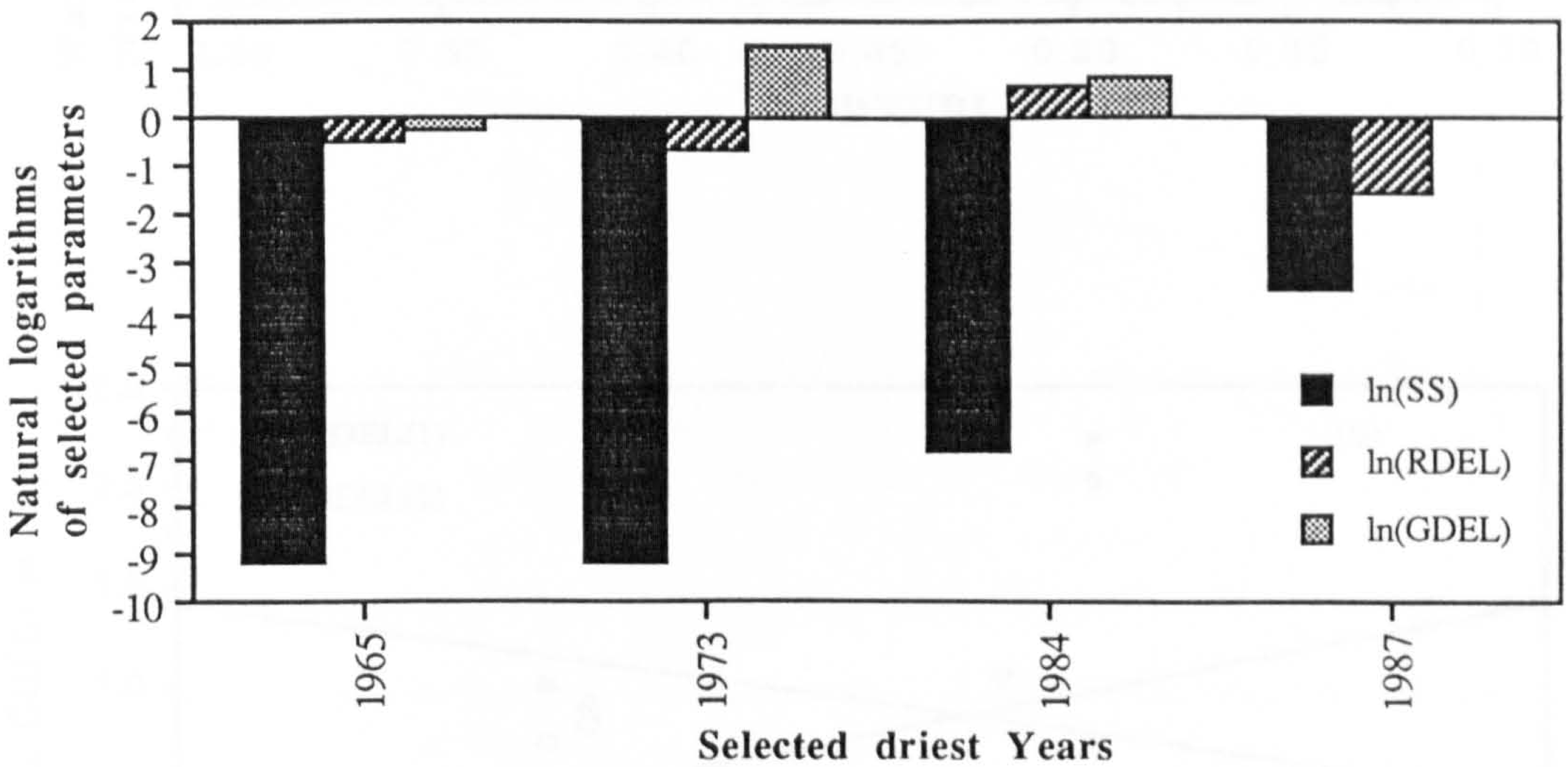
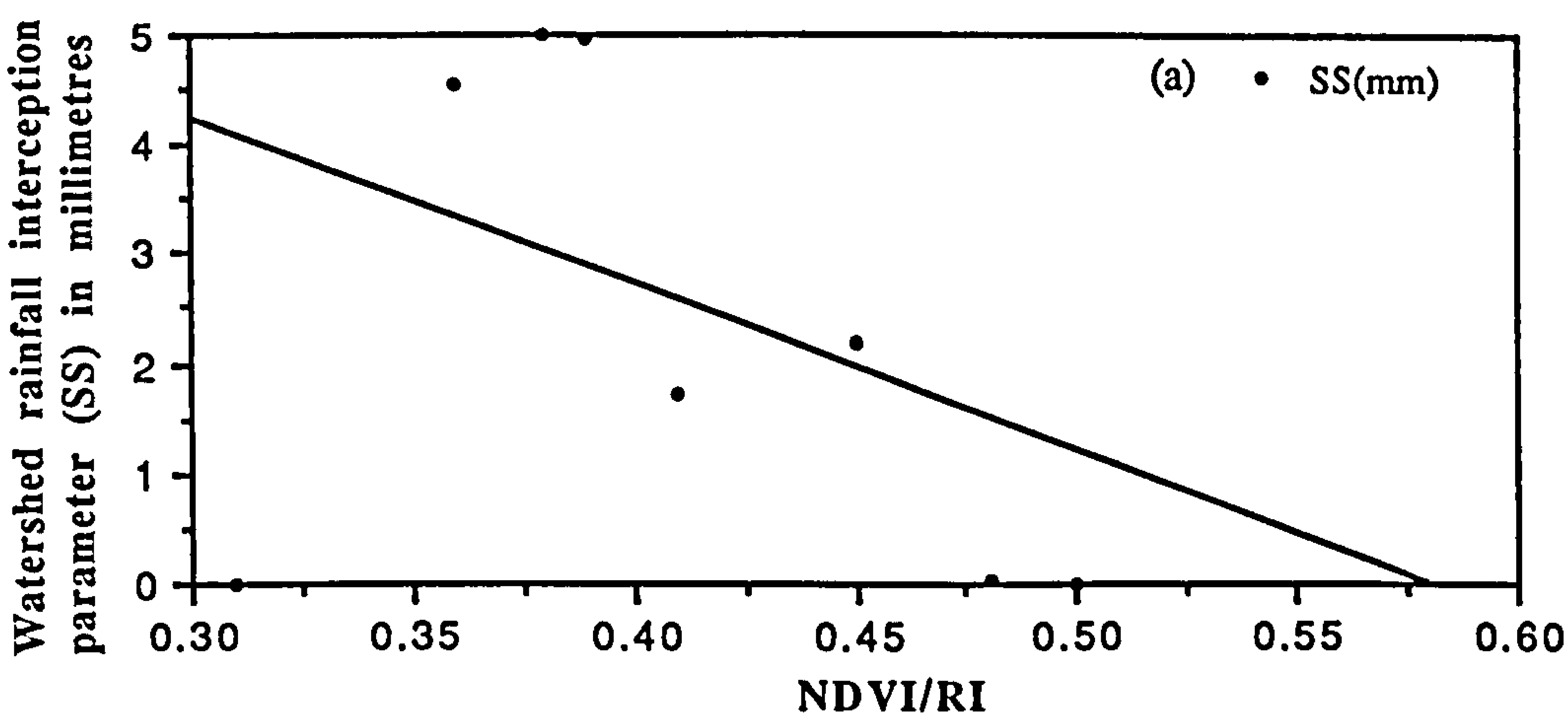


Figure 7.3. Time series plot of selected parameters during dry years (1965, 1973, 1984 and 1987)

Figure 7.4(a & b) compares the model 'store' parameters (SS, RDEL, GDEL) and normalised difference vegetation index (NDVI) presented in equation 6.1. The objective was to evaluate changes of selected parameters with respect to changes in land use. The changes in NDVI value as compared to the model parameter values were then used to infer effects of land use on streamflow since NDVI was established to relate well with streamflow in equation 6.6. To achieve this goal, it was necessary to plot the relationship between the individual 'store' parameters and NDVI values. However, since NDVI represent changes in land vegetation cover which on the other hand depends upon the rainfall for vegetation re-growth and effects of human activities on the watershed, an examination of the ratio of NDVI and rainfall index (RI) provided a better measure of vegetation response to rainfall and hence its influence on the surface interception (SS), the RDEL and GDEL. A large ratio would result if the area received a relatively more rainfall which induced vegetation re-growth. This increased vegetation cover will increase the rainfall interception (SS) parameter. It will also increase the RDEL time because of increased surface storage and infiltration.



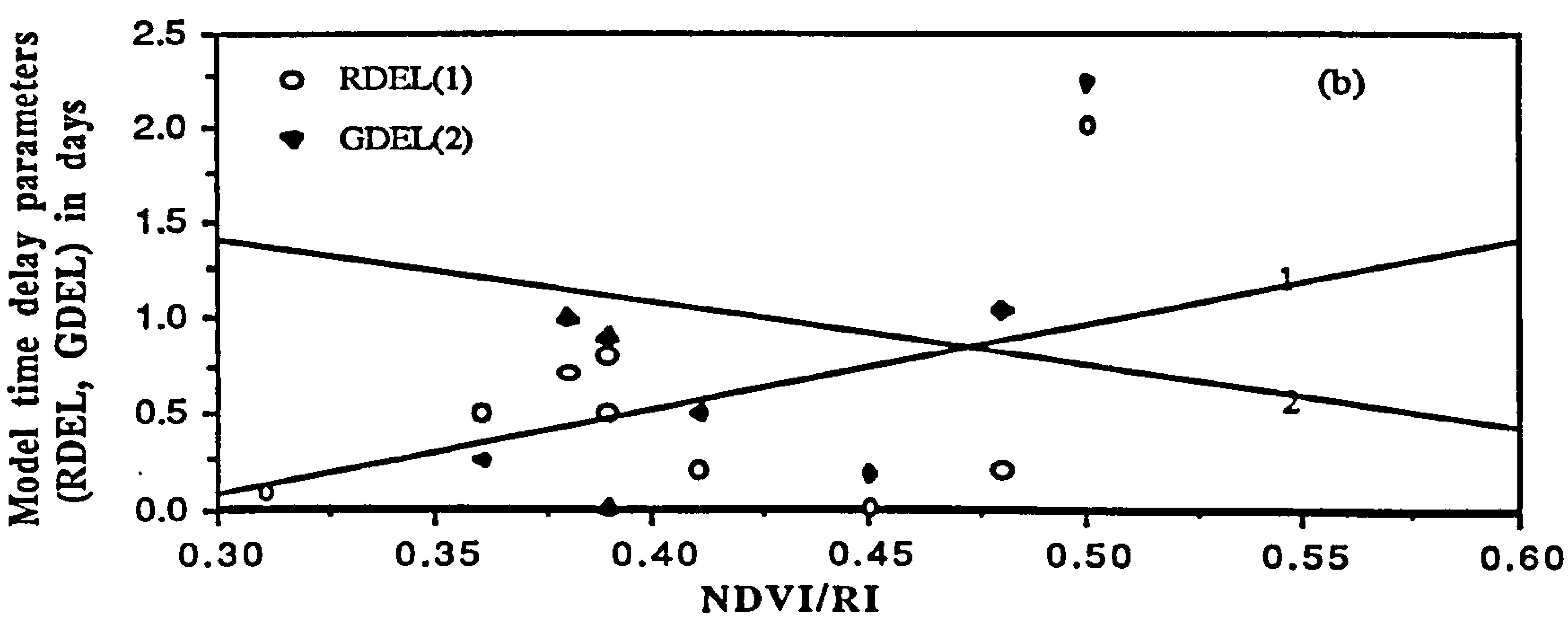


Figure 7.4(a&b). The relationship between NDVI/RI and model 'store' parameters.

After a long duration of continuous rainfall, infiltration and percolation, the GDEL will decrease as the watershed gets saturated. The opposite would occur during the dry period. The NDVI/RI ratio decreases (land degradation), and induces less surface interception, decreases RDEL and increases the GDEL parameter. But RDEL, on the other hand would change this pattern as the watershed gets saturated. Because less space will be available for additional moisture storage, shorter time is taken for surface runoff to accumulate and be released into the streams. On the basis of this analogy, it is possible to interpret the model parameter values with respect to changing land uses.

While acknowledging the existence of errors in estimating NDVI and during the model parameter optimisation, there is evidence of the parameters being sensitive to the combined changes in land use. As implied in section 6.4, the combined effect of the rainfall regime and human activities could be represented by an integrative watershed coefficient (Ic) from equation 6.4. This coefficient and the selected model parameters should also rate the performance of the HYRROM model in evaluating effects of watershed changes on streamflows. As a basis of this opinion, the relationship between the optimised model 'store' parameters and Ic were provided in a scatter and eye-fit plot in Figure 7.5(a&b).

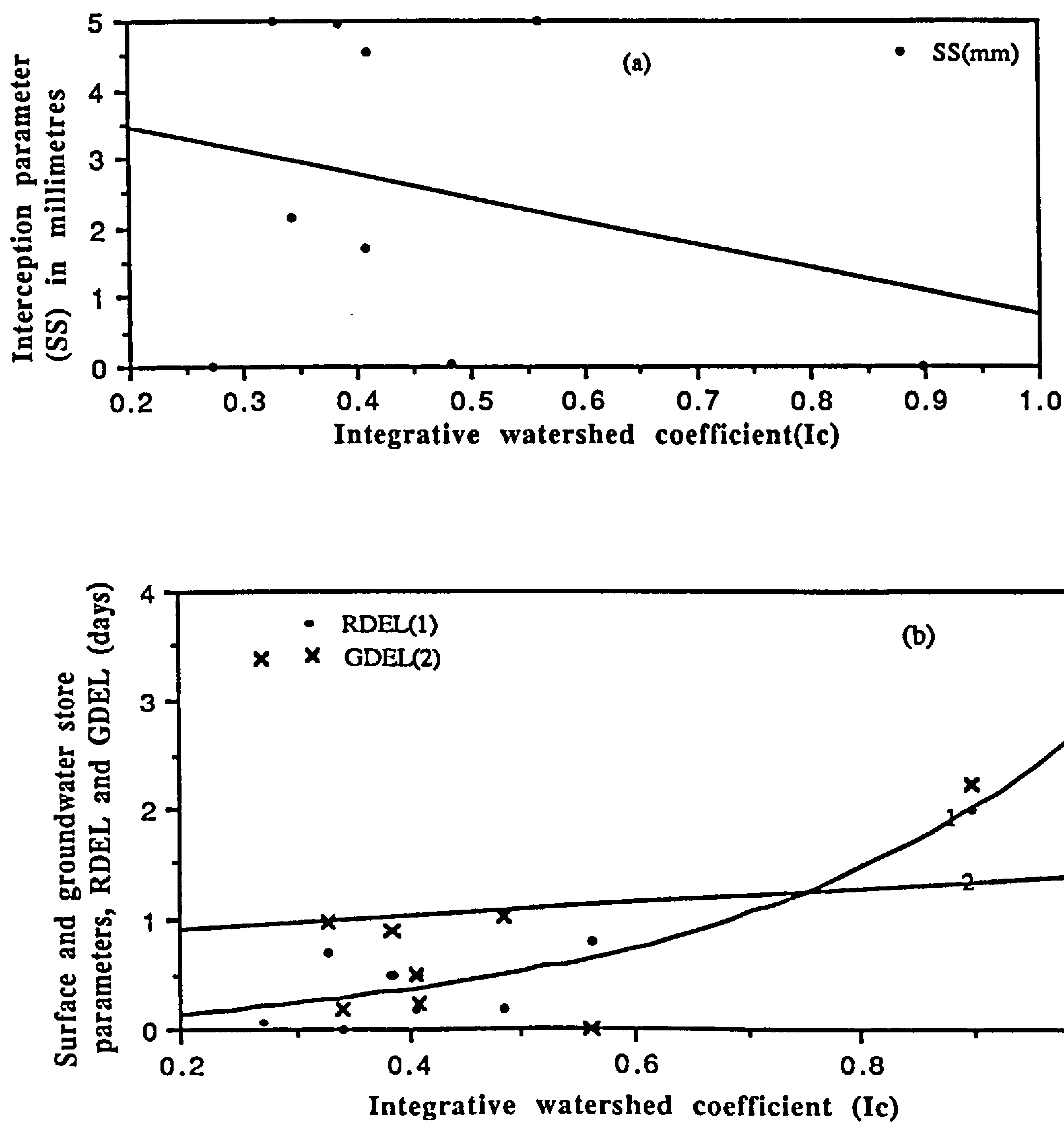


Figure 7.5(a&b). The relationship between selected model parameters and NDVI (1982-1990).

The results in Figures 7.4 and 7.5 therefore depict decreasing trends of SS with increasing NDVI/RI and Ic. Although, the correlation indicate poor fit (r= -0.397) for SS and NDVI/RI, the relationship signifies the relative importance of vegetation cover on interception. The relationship between the Ic and the RDEL is relatively stable and good ($R^2 = 0.84$) and increased with increasing Ic. The GDEL parameter seemed to slightly increase with increased Ic (although not significant r= 0.17). Therefore HYRROM model parameters can be used with NDVI, RI, and Ic to infer the effects of land cover changes on streamflow generation. In other words, the parameters could be optimised to simulate changes in water balance components associated with different levels of NDVI (land use changes).

7.4. ANALYSIS OF THE MODEL PERFORMANCE

7.4.1. The coefficient of Efficiency

The coefficient of efficiency (E) is derived from the relationship between the observed and predicted flows and the long-term mean flow value ($\overline{Q}$):

$$E = \frac{\sum_{i=1}^{N}\{Q_{obs_i} - \overline{Q}\}^2 - \sum_{i=1}^{N}\{Q_{pred_i} - \overline{Q}\}^2}{\sum_{i=1}^{N}\{Q_{obs_i} - \overline{Q}\}^2} \quad (7.9)$$

The coefficient of efficiency expresses the proportion of the variance of the observed flows that can be accounted for by the model (Nash and Sutcliffe, 1970) and provides a direct measure of the ability of the model to reproduce the observed flows with E = 1.0, indicating that the predicted flows are the same as the observed flows (Chiew and McMahon, 1994). The overall results are presented in Table 7.5 and Figures 7.6 - 7.8.

7.4.2. Analysis of Model Parameters

The relative significance of the respective parameters is examined by analysing a set of parameters describing the three stores and carrying out three types of analysis. A measure of the sensitivity of the objective function due to the changes in parameter values are calculated. The sensitivity measure used is defined as the ratio of the proportionate change in the value of the objective function resulting from the change of the four parameter value set for the period 1964-1990 as

$$SENSI = \frac{(F_n - F)}{F|PC|} \quad (7.10)$$

where,

Fn is the new value of the objective function resulting from the new set of the parameter values, F is the original value of the objective function and $|PC|$ is the absolute percentage change between the set of parameters. Therefore the sensitivities among the three analyses are:

SENSI(1) = (F2-F1)/F1*(%Δ Error 2-1)
SENSI(2) = (F3-F1)/F2*(%Δ Error 3-1)
SENSI(3) = (F3-F2)/F1*(%Δ Error 3-2)

The results of the analysis are given in Table 7.6. A SENSI = 1.0 means that a 1% change in the parameter set values would result in a 1% change in the value of the objective function (Kuczera, 1983). This representation therefore is sufficient to provide direction on the relative importance of the model parameters (Chiew and McMahon, 1990). For the purpose of this study, the approach is sufficient to extract the changes in parameter evolution with time (1964-1990).

Table 7.5. A summary of the sequential analyses and optimisation of the parameters

		Analysis 1 4 Parameters			Analysis 2 8 Parameters			Analysis 3 all Parameters		
YEAR	DAYS	F1	%Error1	E1	F2	%Error2	E2	F3	%Error3	E3
1964	365	1.889	0.65	0.999	1.846	1.510	0.999	1.846	1.51	0.999
1965	365	0.039	-16.02	0.999	0.038	-5.89	0.999	0.038	-1.30	0.999
1966	92	0.199	-45.30	0.92	0.168	-28.88	0.968	0.138	0.83	1.000
1967	365	1.144	-74.69	-654	1.079	-70.00	-588	0.943	7.84	-4.737
1968	365	3.414	-62.08	0.06	3.149	-59.08	0.144	2.870	-4.59	0.996
1969	365	0.142	-31.44	0.98	0.126	3.030	0.999	0.124	3.77	0.999
1970	366	1.001	-2.04	0.99	0.988	-0.490	0.999	0.971	-0.56	0.999
1971	365	2.691	-68.71	-0.69	2.621	-78.79	-1.228			
1972	151	0.131	43.93	0.97	0.128	40.560	0.981	0.098	-0.93	1.000
1973	365	0.149	-53.14	0.98	0.135	-25.82	0.996	0.121	-0.14	1.000
1974*										
1975	150	0.144	-53.78	0.98	0.132	-45.66	0.988	0.103	5.95	1.000
1976	366	0.469	-76.64	0.71	0.430	-61.51	0.812	0.369	14.54	0.989
1977*										
1978	365	2.278	54.64	-0.22	0.884	25.92	0.724	0.771	2.77	0.995
1979	184	1.606	9.01	-0.01	1.179	70.13	-59.74	1.490	9.15	-0.059
1980	305	0.180	-69.81	0.95	0.157	-46.00	0.979	0.133	-0.70	1.000
1981	153	1.641	18.20	0.48	1.083	26.33	-0.082	1.641	18.20	0.487
1982*	275	1.687	125.66	-13.35	1.404	47.40	-1.041			
1983	184	0.642	16.42	-0.47	0.625	5.57	0.829	0.620	4.32	0.896
1984	366	0.425	-59.70	0.99	0.423	-23.75	0.999	0.422	0.85	1.000
1985	306	0.783	-293.3	-2.29	0.309	8.36	0.997	0.295	5.98	0.998
1986	365	0.180	-0.92	1.00	0.173	16.73	0.998	0.180	-0.92	1.000
1987	365	0.115	-29.94	0.99	0.108	9.36	0.999	0.108	-1.02	1.000
1988	90	0.007	-13.00	1.00	0.007	-7.68	1.000	0.006	0.64	1.000
1989	275	0.682	14.82	0.62	0.535	-18.82	0.388	0.515	-0.32	0.999
1990	365	2.221	17.04	0.82	1.549	25.27	0.603	0.985	5.90	0.978

* omitted because of their extremely discontinuous data set

A large SENSI value would indicate that a small change in the parameter set value can affect significantly the value of the objective function, and the parameters should be optimised adequately. A very small SENSI value indicates that the parameter set are of little importance and can take any value without affecting significantly the streamflow estimates (Chiew and McMahon, 1994). The selection of the most sensitive set of parameters is based on the largest absolute SENSI value in Table 7.6.

The results show existence of variation between the parameter sets and hence, it would be advisable to optimise all the parameters. It seems however, that this approach does not explicitly isolate the most sensitive parameter. Hence for a detailed model sensitivity analysis, each parameter should be tested individually, so as to clearly identify the most sensitive parameter. Apparently this is only possible when the model is to be refined in an experimental watershed. For the purpose of this study, the three measures of objective function, percentage error in flow estimates and the coefficient of efficiency have shown that the model performance is reliable for this set of data. However, for the model to simulate the flows reasonably well, all the parameters should be optimised.

Table 7.6. Categories used to classify sensitive set of parameters

YEAR	SENSI(1)	SENSI(2)	SENSI(3)	Sensitive set
1964	-0.020	-0.020	0.000	1
1965	-0.260	-0.377	0.000	2
1966	-2.558	-14.140	-5.305	2
1967	-0.266	-14.500	-9.811	2
1968	-0.233	-9.161	-4.828	2
1969	-3.884	-4.463	-0.012	2
1970	-0.020	-0.044	0.001	2
1971	0.262	-68.710	-78.790	3
1972	0.077	11.301	9.724	2
1973	-2.567	-9.960	-2.663	2
1974				
1975	-0.677	-17.006	-11.339	2
1976	-1.258	-19.441	-10.788	2
1977				
1978	17.575	34.314	2.959	2
1979	-16.250	-0.010	-16.085	1
1980	-3.042	-18.045	-6.925	2
1981	-2.764	0.000	-4.189	3
1982	13.128	125.660	47.400	2
1983	0.287	0.415	0.010	2
1984	-0.169	-0.427	-0.058	2
1985	-182.614	-186.524	0.108	2
1986	-0.686	0.000	-0.714	3
1987	-2.392	-1.760	0.000	1
1988	0.000	-1.949	-1.189	2
1989	7.251	3.707	-0.692	1
1990	-2.490	6.199	7.053	3

7.4.3. Relationship Between Observed and Predicted Flows

The data presentation of the predicted and observed flows given in appendix E depict limited physical changes because of their condensed scales. They still however, show the decreasing trends of the flow characteristics from the 1960s to 1980s. Figures 7.6, and 7.7 were prepared to present the performance of the model in simulating the annual streamflows.

Figure 7.6 compares the observed flows and flows predicted by the model. Although the hydrographs show very small difference, because of the condensed time scale, the individual year-to-year comparisons in Figure 7.6 indicate the different flow characteristics in the various days and years. A regression of the observed and the predicted flows in Figure 7.7 gives a straight eye-fit line of $R^2 = 0.998$ when all the parameters are optimised. The ratio of the predicted and observed flows in Appendix E also supports these observations, as it is close to unity in most cases, except in 1965 and 1968 where it was greater than 1.0 and 1976 and 1981 when it was less than 0.90.

The results of the objective functions, coefficient of efficiency and the ratio of the predicted to the observed flows have enabled the study to conclude that the model is good for predicting with this set of data. The year-to-year optimisation however did not give stable parameters and attempts to simulate longer periods such as 5 or ten-year intervals may reveal changes in parameters values which can be related to changes in land use during those periods. Secondly, it was difficult to estimate reliable model parameters for the area probably because of the contrasting nature of the region in terms of tropical climate, relief and rapid changes in natural resource use. This limitations in validating model parameters has been raised by researchers in the IH, particularly by Reynard (1995) while using the general circulation models (GCM) to simulate changes in runoff for some catchments in Kenya.

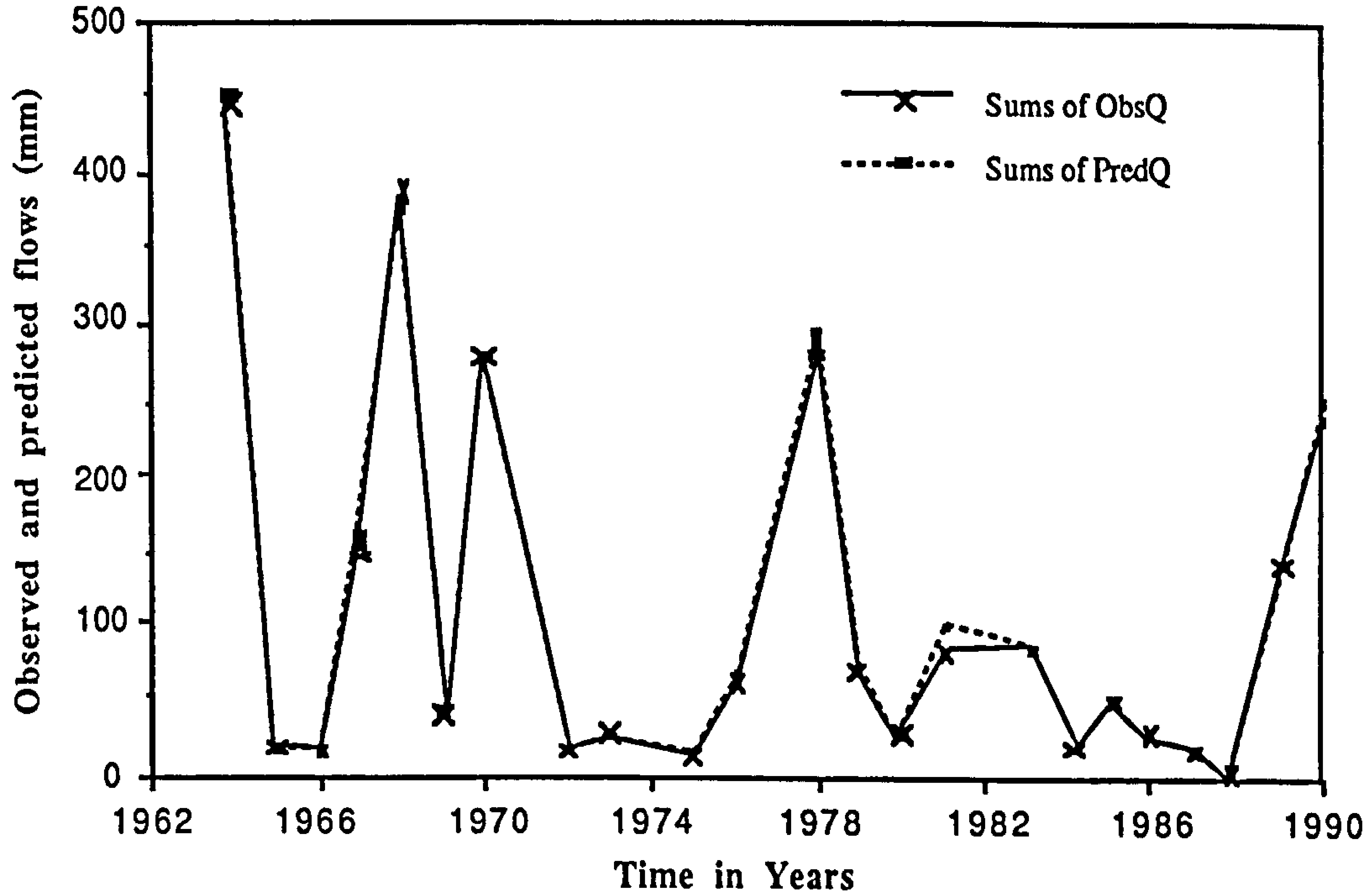


Figure 7.6. Comparison of the annual observed flows to flow predicted by HYRROM in subwatershed I.

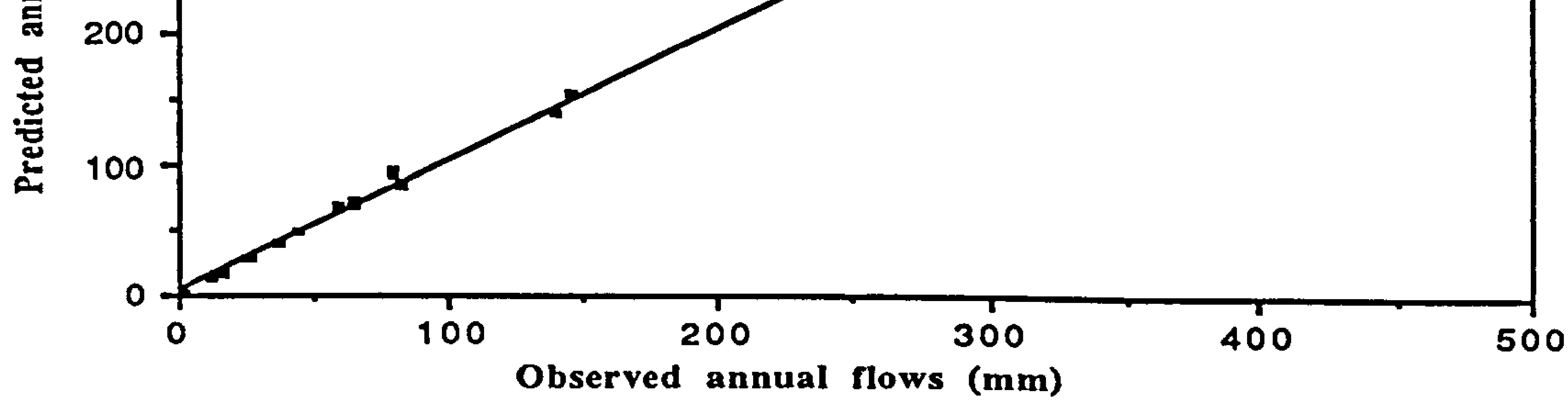


Figure 7.7. Regression of the observed and predicted flows with all parameters optimised in the HYRROM model.

7.5. CONCLUDING REMARKS

This chapter described the application of the rainfall-runoff model, HYRROM to 27 years of mean daily rainfall, flow and evapotranspiration at Enjoro river watershed in Kenya. The watershed is characterised with different climatic and physiographic patterns. Three simulations were carried out for each year, with the simulations differing in the number of days of the year. Comparisons of the daily observed and predicted streamflows indicate that HYRROM generally gives satisfactory estimates of the runoff, although it had difficulties in simulating long duration data sets. Therefore, to obtain a reliable optimised estimate from the model, a continuous good quality daily rainfall, flows, and evapotranspiration is required. The year-to-year optimisation however, did not yield stable watershed parameters, hence a calibration at longer durations would mostly likely provide a clearer picture of stable and unstable physically based parameters. These could then be used to infer changes in streamflows with respect to changes in land use.

The sensitivity analysis indicated that some parameters were more sensitive than others. The optimisation of all the nine parameters however is necessary for the model to simulate flows reasonably. In particular, the SS, RDEL and GDEL parameters were found to be more sensitive to changes in rainfall regime and land cover. These parameters were therefore assumed to relate to some of the watershed characteristics, but it was difficult to identify which in particular watershed characteristics (e.g. land vegetation cover) influenced the SS parameter, and which influenced the surface and groundwater generation mechanism. The trends of the three 'store' parameters assumed the patterns of changed land use from the 1960s to the 1980s discussed in chapters V and VI. The study tried to generalise and to infer the cause-effect of these observations, but they are yet inconclusive. It was difficult to estimate and validate long-term reliable model parameters because of geographical contrasts and tropical nature of the region. Further analysis is necessary in the future in controlled catchment studies to qualify the observations, suggestions and inferences made in this analysis.

The analysis seemed to have raised more questions than it provided answers, perhaps, because of the nature of the subject of dealing with uncertainties masked with natural dynamism and human interventions. At present the inadequacy of climatic data did not allow for an evaluation of the spatial and temporal variability of evapotranspiration and hence it was not possible to compare the periodic trends of observed and predicted streamflows. The questions raised however, should serve as a baseline for future hydrologic model evaluation, and refinement for streamflow simulations in Kenya.

CHAPTER VIII

CONCLUSIONS AND RECOMMENDATIONS

8.1. CONCLUSIONS ON THE RESEARCH HYPOTHESES

The environmental changes on the earth from a higher use rate of its natural resources (deforestation, large-scale agriculture, construction of hydraulic structures) have altered its natural biogeophysical system. Scientific research is required in the development of utilities, facilities and technologies that are environmentally friendly and which advance sustainable utilisation of these resources. Considering the impact of human activities on the water resource and its sustained availability, a need was strongly felt for information on human-induced watershed changes, and climatic interactions so as to establish the extent and nature of their effects on small hydrologic systems. It is only by fully understanding these interactions that appropriate and sustainable watershed development plans can be designed and implemented. Consequently, the world community during the 1992 Rio Earth summit recommended specific and immediate action plans concerned with hydrological systems experiencing extreme human actions.

The progress in understanding the full array of human-induced interventions in hydrologic systems assumed upward momentum in the 1990s with the increased availability of data such as from the satellite imagery. Before this period, major progress was mainly in the study of the effects of large hydraulic structures. Research into the effects of inter-basin transfer and human-induced changes gained much interest only very recently with exceptions of Pereira's work in East African catchments and research works in Machakos in Kenya. Information in this area is therefore scanty in Kenya, partly because of the difficulty of isolating inadvertent from advertent human influencing factors and partly because of previous abundance of natural resources that apparently encouraged uncontrolled use rate.

The indicators considered in this study were mainly the hydrological regime of the watershed before and after perceived anthropogenic interventions and climate changes. Through five objectives in the introduction, literature review and results from data analysis, this study has provided an understanding of the effects of human interventions in a tropical watershed. It recommends integrated watershed development plans for a sustainable ecological system. The general conclusions that can be drawn regarding man-induced hydrological changes are:

1. Human-induced changes on streamflows may be estimated quantitatively and qualitatively from a knowledge of long-term variations of flows in combination with the

analysis of the natural fluctuations of hydrometeorological factors and of developments in the watershed. Such an approach would allow for an estimation of the integrated influence of major human factors within the watershed and may serve as a basis for objective evaluation of land and water resources. The approach however is limited in scope because it does not reveal the influence of each human factor in isolation.

2. Quantitative estimation of human influence upon streamflows is difficult and rather complex even with reliable long-term observation of the influencing factors because human modifications occur simultaneously and are superimposed on natural variations in rainfall and river flow. This study however has shown that the difficulty can be circumvented by developing trends of long-term correlations between the river flows and the indices that characterize quantitatively the degree of human induced changes in the watershed. Taking into account the nature of the watershed, the extent of land and water use, and economic development, various hydrological indices(RI, relative rainfall and runoff factors (ki), runoff coefficients, Z-score values, residuals, CUSUM(t), trends of mass curve slopes, andhydraulicityfactors) are chosen to establish the trends and patterns of advertent human-induced factors.

3. The study also demonstrated that quantitative estimate of hydrological changes due to human activities is based on several methods which range from simple graphical analysis, temporal and spatial homogeneity analyses to deterministic and stochastic models. Statistical correlation and regression provided useful means of analysing past behaviour and for the prediction of future consequences. The statistical methods were advantageous over graphical analysis because of their ability to consider multi-variate problems and the availability of measures for testing their levels of significance. Analysis of time series before and after the perceived change in the river regimes provided an understanding of the corresponding changes in the various statistical elements of the hydrologic system.

8.1.1. Hypothesis 1: *The consensus scientific opinion that land use patterns modify streamflow characteristics can be shown to be true for small rural watersheds in Kenya.*

The study was conducted to describe and understand the hydrological system, during the pre-land use change in the 1960s and compared with the situation observed in the 1970s and 1980s. A few watersheds' characteristics and elements of the hydrologic system were selected for detailed analysis to characterise the watershed as influenced by human-induced land use change.

Firstly, human activity was found to influence not only the mean flows, but the magnitude and frequency of droughts from year-to-year and season-to-season. This variation directly affected the availability of water supplies for domestic, agricultural and industrial uses. The results from the land use analysis showed the watershed to have experienced a 20% increase in agriculture, 30% decrease in forestry and 10.4% increase in urbanisation. The combined effect of this sudden change in land use (1970-90) decreased the annual flow from 183 mm/yr in the 1960s, to 114 mm/yr in the 1970s, and 63 mm/yr in the 1980s when there was minimum change in rainfall. Evidence of increased flows after a few years of vegetation removal (due to droughts, and enhanced by human activities) were seen in the 1970s during an exceptionally heavy rainfall year. The effects of increased urbanisation increased runoff from rainfall of even small magnitudes and reduced moisture storage in the watershed leading to reduced river flows in the dry season.

Secondly, the demographic trends altered the land occupancy levels from 25 persons per km^2 in the late 1960s to 250 persons km^2 in the 1980s; a tenfold increase in less than three decades. Rural-urban out-migration jumped from 7.41 in 1979 to 18.88% per year in 1993; an increase equivalent to land conversion from rural to urban of roughly 2.38% yr^{-1}. Assuming a linear rate of this magnitude is maintained, Njoro division will most likely become fully urbanised in less than 22 years. Further, it was found that a lack of a well-defined land use policy created a dynamic change in land ownership that is frequently subdivided to small commercial plots thus increasing urbanisation. This human population explosion, urbanisation, and agricultural development increased water use rate abstracted directly from the Enjoro river and increased pumpage from boreholes in the area. This response directly and indirectly altered streamflow generation mechanism from perennial to ephemeral that occurred even after short meteorologic droughts.

These dramatic demographic changes and trends combined with other factors (unquantifiable factors) to produce a hydrologic drought regime in the river and ultimately water unavailability; a scenario never experienced before in the area. During the wet years, particularly high flows were recorded in 1978 which was attributed to increased urbanisation and imperviousness of the watershed. During the dry years however, the effects of land use anddemographic changes assumed extreme levels in that the flow regime changed from ephemeral in the 1960s to hydrologic drought regime in the 1980s. The effect was exaggerated as population pressure forced people to settle, farm and use natural resources in steep and fragile semi-arid parts of the watershed since arable land was already used and in most cases intensively cultivated.

8.1.2. Hypothesis 2: *It is possible to analyse limited data records from changing watersheds to identify the contributions of the regional rainfall regime and of human-induced factors to observed changes in streamflows.*

Before subjecting the hydrological data to critical analysis, they were checked for temporal and spatial homogeneity. The data originated from diffuse government departments and institutions: archives, files, maps, correspondences and thus required total quality checks and control. In addition, the source of the data is a region with contrasting geography and rainfall variability. The philosophy behind this quality control was also to identify the data variability and establish their cause-effect relationship. It was however, difficult to account for direct surface water abstractions from the river and the indirect ground water abstractions. Both these human actions affected the rivers' minimum base flow. The experience gained in handling such diffuse data has demonstrated the possibility of analysing limited data records to extract hydrological information related to effects of watershed changes on streamflows.

Streamflow being an integrated output from the watershed, the runoff coefficient provided an understanding of the changing effects of mean annual rainfall on streamflow generation. The hydrological behaviour of the watershed was thus ascertained by examining its rainfall as modified by its losses and human induced factors. Using this approach, it was established that the regional and local rainfall was not the sole factor that contributed to the changed flow regime. Existence of decreasing trends of moisture supply to the watershed was found with the rainfall index (RI) of 1.0 in the 1960s, 0.99 in the 1970s and 0.97 in the 1980s. The small reduction was however not significant to effect the observed change in streamflow regime. Hence, the large reduction in flows does not appear to be the result of decreasing rainfall.

There appears however to exist a trend towards a generally increased temperature and increased rainfall variability in the East African region. Documentation from the literature (chapters' 2,3) and results from the analyses of the local climatic data (in chapters 5, 6) confirmed that the regional and the local climates did not change markedly from its long-term mean. However, an increased variability of these factors was observed. For example, meteorologic drought occurrences were more frequent in the 1970s and 1980s than in the 1960s. The study also did not find a human-induced rainfall and climatic change. However, during the periods of low RI, the drought effects were felt more during years of increased human interventions, presumably because of increased natural resource use rate.

A proportion of the changes in streamflows however, could be associated with the rainfall patterns through the study of the rainfall-runoff relationship. Water availability in the watershed declined from the 1960s to reach its lowest in the 1980s. The annual runoff coefficient declined from 22% in 1960s to 10% in the 1970s, and 8% in the 1980s. This was ascribed to reduced *in situ* moisture storage of the watershed before the onset of the rainy season. The reduction also reflected effects of changed physical watershed characteristics (increased deforestation, increased agriculture, and increased urbanisation), which reduced moisture recharge and increased usage of water resource. It was however impossible to specifically apportion the percentage of land use and water use to a certain percentage drop in runoff coefficient, except to postulate that, these effects occurred simultaneously and combined with other human factors to produce the observed low flow regime in the 1980s.

8.1.3. Hypthesis 3: *The normalized difference vegetation index (NDVI) derived from satellite imagery, could be used to monitor and predict streamflow responses to changes in vegetation cover in ungauged watersheds.*

The study considered the effects of vegetation cover change on streamflows. Effects of extensive land use changes were reviewed and established to have occurred at unprecedented proportions in tropical Africa, particularly in small rural watersheds. Coupled with the naturally rapid hydrological cycle occurrence in these regions, rainfall patterns became important as well as the meteorological responses to surface heat fluxes and heat balances. Thus the net effect of these processes during periods of climatic irregularity during which land use changed was established to be critical in the region. It was found that an extensive but transient interference with land vegetation cover corresponded to periods during which large effect on the river's hydrologic regime occurred.

The remote sensing technology through normalized difference vegetation index (NDVI) was used to extract spatial and temporal change in land covers and was demonstrated as a potential indicator for monitoring and hence predicting changes in streamflows (chapters IV and VI). Using this NDVI and the rainfall index (RI), it was possible to show that as the rainfall index increased NDVI also increased. This in turn influenced streamflow generation. The rainfall index thus indirectly provided a good indicator of the extent of land degradation. For example, the heavy rainfall in 1978/79 and 1989/90 encouraged increased vegetation cover reflected by increased NDVI value. A simple stepwise regression between flows and NDVI was developed to evaluate effects of changing land vegetation cover on streamflows.

A direct effect of vegetation change on evapotranspiration was difficult to evaluate because of introduced errors in the estimation of evapotranspiration and in the assessment and inference of land vegetation cover on streamflows. Because these activities lead to changes in radiative energy balance and evaporation, they indirectly affect streamflows in a feedback mechanism of: vegetation removal altering potential evapotranspiration elements and hence changes in local micro-climate (chapters IV, V and VI). It was not possible to demonstrate this feedback mechanism in this case, hence there is a need to specify and quantify more precisely vegetation cover change with re-growth of vegetation and/or subsequent land use in experimental catchments particularly in the arid and semi-arid areas in Kenya where human settlement and changes in land ownership occur rapidly.

8.1.4. Hypothesis 4: *That even with limited and discontinuous hydrologic data, existing conceptual hydrologic models could be calibrated and optimised to simulate the hydrological processes that will allow an evaluation of the effects of different land uses on streamflows.*

This study applied the Institute of Hydrology (UK), HYRROM model on 27 years of daily rainfall, evapotranspiration and streamflows to optimise watershed physical parameters in the higher elevation part of the Enjoro river watershed in Kenya. The watershed is characterised with varying physiography and climate. Three simulations were carried out for each year and comparisons of the observed and predicted flows made. A sensitivity analysis was conducted on selected model parameters so as to evaluate the effects of perceived land use changes on streamflows.

The performance of the model in simulating the flows from the combined effect of land use, for three periods examined (1960s, 1970, and 1980s) was relatively good using different values of the parameters for each year. Comparison of the observed and predicted flows indicated very small error differences ranging between -1.30 to 5.90 % during the wet years, and between -0.70 and 14.54 % during the dry years. The coefficient of efficiency ranged between 0.70 and 1.10. A simple regression fit between the observed and predicted annual flows yielded a regression coefficient $R^2 = 0.998$. The three model performance measures indicate an excellent performance for the data series used, but to obtain reliable results, all parameters of the model must be optimised individually, in addition to having a continuous good quality data.

Sensitivity analyses showed the three 'store' parameters (SS, RDEL, and GDEL), to be particularly sensitive to changes in vegetation cover resulting from meteorologic droughts and rainfall regime. This was evident especially during the driest years (1965, 1973, 1984 and 1986), and wet years (1964, 1970, 1978, and 1990) and during decades in which human activities increased (1970s, 1980s). Consequently, the changes were identified with corresponding changes in runoff time delay (RDEL), and groundwater time delay (GDEL) to contribute to total watershed streamflow output. The SS parameter decreased with decreased moisture availability and increased with increased rainfall. This was attributed to increased vegetation re-growth, which in turn increased surface interception. The RDEL parameter varied inconsistently as there was no defined pattern when the watershed rainfall regime changed, but assumed an extreme variability during the years of increased watershed disturbances. The GDEL parameter increased with dryness of the watershed and decreased as the watershed got saturated after a long duration of rainfall. The magnitudes of these three parameters were distinctively apparent in the 1980s, coinciding with periods of increased natural resource use rate and the river regime assuming hydrologic drought status.

It was difficult to estimate and to validate long-term model parameters because of the inadequacy of the data, the geographical contrasts of the region and the limitation of the model to optimise longer-periodic data series. These results however, have shown to some extent that some conceptual lumped models can simulate parameters of a watershed that can be used to infer changes in streamflows resulting from changes in land use patterns.

There is a need however, to invest more time and financial resources to improve the accuracy and reliability of the HYRROM model. It should be made dynamic enough to simulate temporal and spatial changes in land use. It can be improved perhaps, by incorporating some of the features of the energy-balance, micro-process models, to account for the role of vegetation growth. The study for example, found the effects of human activities on the land vegetation cover to be significant mainly at the watershed level, although, they were difficult to explicitly quantify from the model. An inference was thus made since the effects occurred simultaneously in about the same time periods. Hence, there is a need to develop models that account for specific problems at particular climates and land use changes. This could be by bringing engineering hydrologists and biometeorologists together in the development and verification of these conceptual models.

8.2. RECOMMENDATIONS FOR FURTHER RESEARCH

The recommendations for further studies are based on the investigative procedures used and on the interpretation of the findings. The broader requirement of scientific research directed at the evolution of strategies and approaches for sustainable conservation of the natural environment forms the backbone of these recommendations.

8.2.1. The Methodology and Validation procedures

Difficulties and limitations were met with in the search and selection of appropriate methods for the appraisal of human-induced changes in non-experimental watersheds. Subjective evaluations from graphics and plots alone were insufficient to reveal hidden historical trends, hence several hydrologic indices were utilised. Statistical analyses and homogeneity tests effectively assisted in the data quality control, and validation.

The use of the models to assess the effects of land use on streamflows had their inherent limitations in inferring parameters from estimated variables (since all the water balance components were by themselves a series of estimates). The IH model proved useful however, when applied to selected years during which land use changes occurred. The lack of continuous and good quality data limited the extent of its use in inferring effects of land use on streamflows. Sensitivity analyses aimed to show which parameters of the watershed physical characteristic are sensitive and dominant were relatively unreliable because the data sets were of unknown quality. Consequently, if conceptual modelling was to be improved for a better precision and a more reliable quantitative estimate, a rigorous long-term experimentation and monitoring on small watersheds in the ecologically fragile semi-arid tropics is highly recommended.

8.2.2. Analysis and Interpretation of the Results

The biogeophysical and socio-economic effects of anthropogenic projects are difficult to interpret especially from the perspective of the cumulative long-term consequences of watershed management and development. The problem of land uses itself introducing systematic errors in the rain gauges and river rating stations complicated the matter. The correlations of rainfall and flows with the proportion of the area covered by a different land use did not yield conclusive results because of inadequate quantified land use values. Therefore, future studies should concentrate on a continuous measurement and recording of different land use areas besides rainfall and streamflow data for any meaningful interpretation of their relationships.

8.2.3. The specific Recommendations

Adaptive research is the key in evaluating the effects of different watershed conservation and development strategies especially in diverse agro-ecological zones in Kenya. Studies that start at catchment level and evolve into larger watersheds are recommended in these arid and semi-arid regions in Africa. It will then be possible to quantify the effects of human-induced watershed changes on hydrological regimes. The specific recommendations of this study are:

1. An interdisciplinary approach (by agriculturalists, hydrologists, micro climatologists, geographers) is necessary to achieve a full understanding and the precision needed to detect effects of land use changes on local water resources. This is because land vegetation changes contribute to changes in surface temperatures by changing the surface albedo, rainfall interception, runoff and by modifying energy balance and evapotranspiration. Consequently an inter-and-multi-disciplinary research is necessary to quantify these feedback mechanisms.

2. There is need for small scale experimental studies to observe the effects of vegetation removal and land use under different climates in the tropics (arid, semi-arids, deserts) on streamflows and to solve existing disagreements on both the magnitude and extent of the effects of deforestation on rainfall and streamflows. Such a study should establish the extent and magnitudes of increases and/or decreases in flows following land re-vegetation in these regions where the variability of hydrologic elements relative to temperate climates is considered too dynamic.

3. Hydrologists should stimulate further investigation of conceptual models that address specifically to tropical Africa where extensive land use changes coupled with human population pressure are intrinsically masked to produce extreme local hydrological changes. To advance this knowledge frontier, respective governments in the region should increase research funding, and encourage exchange of information and knowledge between research institutions in the humid and temperate North and those in the South.

4. This study demonstrated the potential application of remote sensing technology in hydrology. Remote sensing satellite data can be incorporated into existing hydrologic models to improve the understanding of the hydrological processes at watershed level. The dependence of albedos on the soil moisture and land use will then be included when estimating evapotranspiration and hence increase the accuracy of rainfall-runoff models. Research that incorporates satellite derived NDVI in a HYRROM type model is thus recommended.

8.3. APPLICATIONS OF THE RESULTS FROM THE RESEARCH

The recognition of the difficulty of analysing limited and discontinuous data in a discontinuous domain calls for the planning and establishment of small experimental watersheds in different ecological zones of the country. These will continuously collect quality data for use in other parts of the country. A dire need exists for procedures that collect, check and process data routinely.

This study has demonstrated the importance of human activities in natural resource use and misuse. Hence a continuous review of the environmental conservation policy is necessary to include the conservation, integration and total watershed planning of all parts of country. It is also recognised that the effects of a large-scale vegetation change as a result of human activities in one country may produce some local climatic and hydrologic effects in neighbouring countries in the region. Hence there is a need to evolve policies aimed at funding research on the effects of human activities at regional levels in Africa.

The study also emphasised the urgency and importance for integrated approach to total river basin conservation, development and management as recommended in the 1992 Rio Earth Summit. However, since a single country is not blessed with adequate resources to harness expertise from all disciplines to study a large and complex region such as the Eastern and Central Africa, countries in the region should pull resources together to fund a regional research and development project that will address the problem.

Finally, for a continued survival of the Lake Nakuru ecosystem, Enjoro river watershed must be rehabilitated.The ongoing soil and water conservation meaures by the Government of Kenya should encompass the whole spectrum of human development. The watershed rehabilitation programmes must concentrate on balancing the basic needs of human life and the availability of land and water resources so as to achieve an ecologically sustainable hydrologic system. This can be achieved by formulating and implementing a thorough land use policy. Such a policy should clearly specify the minimum land economic unit that is environmentally sound and can sustain water and food exploitation without adversely affecting hydrologic regimes of rivers. Land fragmentation and subdivision into small commercial plots must be discouraged in the policy.

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APPENDICES

Appendix A1.1. Time series of the monthly rainfall (mm) at Teret rainfall station No.09035233

YEAR	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	Annual Totals
1963	47.50	38.60	90.30	219.00	223.90	97.40	46.80	157.00	47.00	45.70	153.90	212.00	1379.10
1964	36.30	19.00	105.20	202.10	75.30	131.60	141.70	122.80	124.10	106.50	97.90	80.20	1242.70
1965	48.80	0.00	26.70	99.40	105.60	14.20	74.10	38.10	23.40	62.00	65.60	25.80	583.70
1966	9.10	48.00	55.10	196.80	47.10	45.20	74.60	143.90	112.10	55.90	85.00	6.30	879.10
1967	5.30	0.00	38.20	159.50	154.00	49.70	226.00	56.80	35.90	125.80	253.90	0.00	1105.10
1968	0.00	178.50	74.83	372.60	116.61	82.79	101.66	91.30	21.30	32.80	140.50	61.66	1274.55
1969	54.20	94.70	155.90	62.10	110.50	132.60	97.20	52.20	112.70	36.30	93.50	38.40	1040.30
1970	202.90	16.60	266.70	277.20	185.80	101.50	91.80	158.90	117.60	30.40	105.00	70.40	1624.80
1971	70.50	0.00	32.70	257.90	113.00	184.50	77.50	230.50	37.00	30.00	58.00	102.80	1194.40
1972	50.50	137.70	14.80	49.80	81.00	107.60	44.50	102.30	34.60	94.60	161.20	24.70	903.30
1973	54.20	31.60	12.40	65.80	106.10	28.90	74.30	124.80	77.90	36.00	59.20	4.30	675.50
1974	15.20	21.70	103.60	159.60	35.60	45.60	128.80	181.80	155.60	81.20	70.00	18.50	1017.20
1975	15.90	8.90	36.20	161.50	123.80	128.50	134.90	285.80	92.70	103.00	17.20	49.20	1157.60
1976	1.20	30.60	20.80	95.90	87.60	53.70	164.10	52.20	58.70	33.20	45.80	64.60	708.40
1977	74.10	46.80	37.60	301.20	168.70	62.80	201.90	126.90	45.90	99.80	222.00	85.40	1473.10
1978	75.40	135.20	133.40	112.50	59.40	82.80	131.30	137.90	94.90	83.20	86.60	148.00	1280.60
1979	102.70	196.30	164.90	228.90	112.80	89.00	57.40	153.90	70.70	28.90	128.80	71.90	1406.20
1980	78.80	3.80	66.90	103.90	226.00	63.60	19.30	38.70	31.20	32.80	129.70	10.00	804.70
1981	9.50	31.80	120.30	122.10	90.70	62.20	120.40	157.40	104.10	51.90	39.20	33.80	943.40
1982	9.40	144.50	4.40	254.80	175.60	58.90	52.70	129.70	65.20	122.80	140.60	117.60	1276.20
1983	12.20	31.60	31.50	122.70	125.50	119.60	105.20	103.80	127.70	115.90	47.70	151.10	1094.50
1984	11.40	10.40	12.00	99.20	17.20	48.40	77.30	67.40	48.10	96.20	91.80	32.00	611.40
1985	11.70	48.20	116.70	276.10	140.30	113.20	94.70	89.30	60.70	13.60	157.90	9.70	1132.10
Means	43.34	55.41	74.83	173.94	116.61	82.80	101.66	121.89	73.87	66.02	106.57	61.67	1078.61

Appendix A1.2. Time series of the monthly rainfall (mm) at Nessuiet rainfall station No. 0903511[illegible]

YEAR	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	Annual Total
1963	110.40	57.40	73.30	260.70	197.50	103.60	63.20	182.10	0.00	14.50	139.80	222.20	1424.70
1964	137.00	0.00	98.90	201.50	71.40	86.90	166.40	181.40	93.40	85.00	47.40	27.80	1197.10
1965	37.20	0.00	17.50	132.50	62.00	15.00	157.70	61.00	7.60	96.50	77.30	51.80	716.10
1966	0.00	90.30	104.70	193.40	60.90	157.50	130.50	153.60	97.20	59.20	161.80	4.10	1213.20
1967	2.00	0.00	34.40	137.90	203.00	77.40	110.40	97.80	12.20	79.70	186.30	0.00	941.10
1968	0.00	146.30	161.90	342.40	54.80	51.90	128.60	109.70	6.30	87.70	225.50	0.00	1315.10
1969	108.00	84.60	114.90	113.20	155.50	37.00	86.90	114.30	164.00	41.80	130.00	33.00	1183.20
1970	296.10	35.00	225.50	198.00	232.10	34.00	150.70	178.20	81.70	43.80	95.10	61.60	1631.80
1971	43.00	0.00	32.00	166.50	101.00	195.00	134.00	247.10	51.00	17.00	107.00	145.50	1239.10
1972	37.00	175.00	12.00	14.80	94.10	92.30	63.60	145.70	46.00	93.50	150.90	37.00	961.90
1973	71.20	44.60	2.10	151.00	100.60	50.70	72.60	127.60	123.50	48.60	63.00	3.00	858.50
1974	0.00	24.00	106.20	167.10	26.70	55.30	130.00	315.50	101.80	85.70	61.00	0.00	1073.30
1975	0.00	16.90	30.80	148.00	144.60	133.00	142.30	248.80	62.50	174.40	27.80	26.30	1155.40
1976	5.90	32.10	31.10	120.50	68.80	52.60	177.40	162.90	154.30	143.00	105.10	67.20	1120.90
1977	56.90	53.90	48.50	334.30	205.50	54.00	82.30	94.80	60.40	78.60	101.76	60.20	1231.16
1978	98.80	166.30	216.30	172.30	5.40	114.20	228.60	114.70	78.40	105.70	43.00	96.60	1440.30
1979	111.20	134.90	147.70	160.50	201.70	71.20	90.80	148.40	62.00	27.60	81.00	37.00	1274.00
1980	55.20	13.50	42.00	93.20	171.40	82.30	24.60	43.40	4.40	25.00	198.40	6.90	760.30
1981	5.00	32.70	174.40	167.30	92.10	50.60	160.90	131.40	69.20	24.40	21.00	67.40	996.40
1982	16.80	17.80	8.90	186.70	200.90	92.50	34.00	257.70	34.90	166.30	121.50	103.80	1241.80
1983	0.00	58.90	6.30	146.50	109.70	87.70	96.50	272.30	95.80	85.20	83.70	48.70	1091.30
1984	56.30	9.70	17.30	100.20	29.30	40.00	95.00	94.20	55.50	59.10	136.10	64.10	756.80
1985	23.00	53.40	49.30	261.50	129.30	123.60	58.20	110.90	74.20	28.10	81.30	19.80	1012.60
Means	55.26	54.23	76.35	172.61	118.19	80.80	112.40	156.24	66.80	72.63	106.34	51.48	1123.31

Appendix A1.3 Time series of the monthly rainfall (mm) at Egerton Rainfall Station No. 09035092

YEAR	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	Totals
1944	20.60	0.80	88.10	287.80	105.70	108.70	148.80	97.00	91.20	56.60	106.20	36.80	1148.30
1945	11.90	25.40	97.00	0.00	121.40	282.70	299.00	251.50	171.70	39.60	91.90	0.30	1392.40
1946	8.60	0.00	25.90	260.90	65.50	65.50	83.10	245.60	194.30	61.70	76.70	0.50	1088.30
1947	94.00	101.90	176.50	363.70	233.40	161.80	102.60	186.90	109.20	84.60	84.10	33.00	1731.70
1948	4.60	7.10	93.00	88.10	110.20	139.20	156.20	198.10	75.70	70.40	55.40	43.70	1041.70
1949	0.00	4.60	8.90	129.00	136.10	94.50	105.70	191.50	207.80	24.10	13.20	8.90	924.30
1950	15.70	2.30	47.20	110.70	57.40	46.50	187.20	70.10	43.70	17.30	15.00	9.40	622.50
1951	0.80	19.30	107.70	330.70	59.70	56.40	74.70	117.60	62.70	103.90	106.40	180.60	1220.50
1952	0.00	11.20	9.90	144.80	132.30	16.50	45.00	90.90	39.90	59.90	45.00	11.20	606.60
1953	11.20	3.30	41.10	101.90	56.10	80.30	59.90	83.30	40.10	40.10	92.70	41.90	651.90
1954	3.50	2.00	3.30	212.30	228.60	157.00	142.50	55.90	88.40	42.40	55.40	22.90	1014.20
1955	60.20	91.20	15.00	134.40	71.90	49.50	81.30	240.00	126.70	79.50	40.60	91.90	1082.20
1956	121.70	53.60	72.10	160.30	74.90	128.80	149.60	332.50	70.90	92.70	72.40	36.30	1365.80
1957	55.10	7.10	64.30	121.90	251.70	108.50	50.30	113.80	20.10	30.20	109.70	53.10	985.80
1958	54.90	186.20	106.70	111.50	189.50	64.30	187.20	79.50	84.30	65.50	25.40	80.50	1235.50
1959	52.60	26.40	73.20	80.80	197.90	101.90	152.90	70.90	112.30	33.50	106.90	18.50	1027.80
1960	29.50	13.20	93.70	69.60	150.10	28.20	50.50	194.60	70.10	60.50	34.30	45.00	839.30
1961	2.80	9.10	18.00	137.90	83.10	45.00	39.40	222.80	58.70	230.90	439.90	141.70	1429.30
1962	70.90	3.30	82.80	100.30	132.80	82.30	99.10	112.80	102.60	61.70	73.90	61.20	983.70
1963	59.20	58.20	87.90	187.50	170.80	45.70	27.40	144.70	25.10	11.90	129.70	227.80	1175.90
1964	2.50	44.20	78.00	285.00	67.50	40.30	135.90	124.50	99.90	113.30	49.70	46.90	1087.70
1965	41.70	2.80	21.20	98.60	88.60	27.40	53.30	69.90	22.90	79.10	51.20	45.60	602.30
1966	9.30	67.30	102.70	271.90	54.30	88.10	93.90	91.90	106.90	36.10	107.00	10.10	1039.50
1967	2.60	6.10	28.40	97.10	203.70	113.20	112.30	80.70	27.30	50.10	140.70	40.60	902.80
1968	0.00	183.60	155.00	273.60	72.60	53.40	137.30	62.50	18.00	50.30	127.70	83.40	1217.40
1969	77.70	79.70	84.70	48.20	129.90	32.70	57.20	66.00	110.10	31.60	75.50	18.60	811.90
1970	146.00	13.40	138.50	153.80	161.80	56.80	83.50	106.50	73.60	57.20	102.10	32.70	1125.90

Appendix A1.3 contd													
1971	48.00	0.00	29.10	155.80	123.10	96.40	54.20	212.20	65.20	29.90	68.10	89.00	971.00
1972	20.40	109.40	16.10	18.80	136.20	67.10	58.90	142.20	36.40	77.70	101.10	19.90	804.20
1973	48.30	58.70	1.10	58.80	136.30	45.40	97.70	195.10	104.40	23.60	49.30	6.30	825.00
1974	11.50	16.60	122.90	207.30	57.30	71.80	143.40	282.90	94.50	38.00	43.70	11.10	1101.00
1975	12.90	10.40	9.50	156.80	210.00	158.80	150.00	238.90	81.50	256.70	41.70	40.70	1367.90
1976	3.10	32.50	11.00	110.10	77.20	81.90	136.00	139.00	77.50	18.90	71.10	66.40	824.70
1977	65.30	29.30	38.20	359.30	197.60	54.00	82.30	42.70	60.40	78.60	269.70	60.20	1337.60
1978	100.20	123.30	189.90	183.50	91.00	71.00	220.00	107.80	94.80	113.30	43.60	96.20	1434.60
1979	89.20	232.10	142.60	165.00	26.00	76.00	76.60	107.00	50.00	17.50	81.00	38.10	1101.10
1980	51.70	15.00	70.30	132.90	153.60	74.00	190.00	78.00	19.00	15.90	129.00	3.90	933.30
1981	2.40	30.70	151.80	145.50	78.50	33.90	154.40	122.20	112.50	17.20	28.90	51.30	929.30
1982	9.90	22.80	4.40	151.20	237.00	28.60	54.00	273.30	29.20	114.70	162.50	69.90	1157.50
1983	20.00	31.50	9.50	128.30	64.50	44.00	71.70	203.20	115.50	84.10	73.50	115.50	961.30
1984	6.60	13.10	13.80	112.20	21.60	35.90	61.40	93.80	77.20	67.50	110.80	52.50	666.40
1985	21.60	47.20	38.20	394.30	130.50	143.10	27.10	73.30	32.60	29.30	72.70	27.00	1036.90
1986	3.60	5.40	42.60	189.60	76.80	120.70	127.30	109.70	124.60	17.40	53.30	44.30	915.30
1987	15.40	10.00	39.30	93.60	120.60	134.10	59.40	83.10	31.80	17.60	151.00	7.00	762.90
1988	84.80	26.00	58.10	240.10	129.30	99.80	118.40	151.10	70.10	45.90	61.30	55.00	1139.90
1989	14.30	88.20	78.90	136.80	110.80	24.40	124.60	96.20	96.10	120.40	108.90	95.70	1095.30
1990	96.50	132.70	127.50	153.90	112.50	64.50	58.20	73.40	62.70	91.60	43.70	47.00	1064.20
1991	57.40	2.50	80.60	129.10	72.90	86.60	148.90	205.60	35.20	99.20	35.40	5.60	959.00
1992	12.40	12.50	26.60	162.40	69.20	64.70	80.80	77.30	52.00	68.90	69.30	99.30	795.40
Means	35.78	42.31	65.77	162.20	119.18	80.65	106.35	138.98	77.09	63.85	87.72	51.53	1031.41

Appendix A1.4. Time series of the monthly rainfall (mm) at Njoro Plant Breeding Research Station No. 09035021

YEAR	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	nnual Total
1930	54.60	56.10	147.80	157.50	199.40	112.80	136.90	109.20	102.40	44.50	124.20	35.80	1281.20
1931	27.90	40.60	72.90	171.50	83.30	37.30	80.30	147.60	27.70	42.90	50.80	41.70	824.50
1932	27.20	40.10	111.80	151.90	90.70	49.00	143.00	101.90	105.70	61.20	39.40	128.30	1050.20
1933	27.20	16.00	10.20	24.60	66.80	30.20	93.70	205.50	107.20	52.10	43.70	89.90	767.10
1934	0.00	24.50	29.50	75.70	47.20	162.00	115.80	85.10	21.60	20.60	48.00	21.80	651.80
1935	0.80	95.20	46.00	79.80	103.10	119.10	87.40	61.20	49.30	82.80	49.30	96.30	870.30
1936	86.40	146.60	83.30	181.10	57.90	73.90	81.30	142.50	65.00	58.20	43.20	56.60	1076.00
1937	45.20	20.30	66.30	142.20	186.70	132.80	198.10	129.80	21.30	50.30	120.40	15.20	1128.60
1938	18.80	1.00	70.90	77.70	113.30	65.50	153.40	186.70	40.60	43.90	44.20	32.00	848.00
1939	18.80	4.80	38.60	136.70	42.20	61.50	102.90	67.80	8.60	22.10	116.80	3.30	624.10
1940	33.00	45.50	154.20	129.50	169.70	35.80	77.50	120.40	8.90	32.80	99.30	8.40	915.00
1941	18.30	48.70	127.50	255.30	213.40	81.50	95.00	85.10	37.30	47.00	107.40	53.60	1170.10
1942	4.60	35.30	255.00	130.60	130.30	139.40	105.20	124.00	65.00	16.80	22.90	28.40	1057.50
1943	1.50	28.20	6.90	67.60	79.50	88.10	153.40	151.90	84.10	120.70	59.40	40.60	881.90
1944	9.40	0.80	44.50	141.20	150.40	49.80	134.60	98.60	47.00	42.90	71.90	39.10	830.20
1945	25.90	21.80	11.40	8.40	88.10	199.90	173.00	163.60	101.90	29.70	56.10	6.10	885.90
1946	8.10	0.00	30.00	161.50	52.60	89.20	77.20	203.50	124.20	40.90	43.70	26.20	857.10
1947	85.60	30.20	98.30	249.40	191.50	115.80	122.70	145.50	130.00	54.40	22.60	44.20	1290.20
1948	1.30	6.60	54.90	104.40	63.80	160.80	103.90	141.70	89.90	32.30	23.40	57.40	840.40
1949	0.80	21.60	14.50	126.20	81.50	81.30	99.30	117.60	111.50	20.80	32.00	52.80	759.90
1950	24.60	0.00	66.00	116.80	48.50	63.00	130.30	103.10	62.00	51.80	30.20	22.10	718.40
1951	6.40	12.40	106.40	329.20	85.60	76.70	95.00	109.20	53.30	107.20	147.80	148.80	1278.00
1952	0.80	20.30	6.90	143.80	145.30	17.30	58.90	84.30	52.80	38.90	54.60	10.40	634.30
1953	8.90	7.40	58.90	89.40	37.10	89.90	42.90	88.40	39.90	46.20	81.30	16.80	607.10
1954	0.00	17.80	6.10	235.00	244.90	156.70	153.70	77.20	49.30	77.50	45.50	16.80	1080.50
1955	35.00	123.40	11.70	115.60	95.50	46.20	64.30	168.70	163.60	54.00	73.40	70.40	1021.80
1956	116.10	49.30	81.30	103.60	68.10	114.60	104.90	274.80	87.10	76.20	75.40	37.10	1188.50
1957	37.80	6.40	61.20	61.20	235.20	78.00	37.60	63.20	11.90	33.00	122.90	59.20	807.60
1958	57.70	129.00	148.10	120.10	217.40	42.90	167.90	126.50	121.40	70.90	34.50	77.20	1313.60
1959	49.30	27.40	75.90	75.90	212.90	69.30	95.30	39.10	92.50	34.30	90.70	21.30	883.90

Contd Appendix A1.4													
1960	9.70	17.50	116.10	67.10	111.30	17.00	31.20	227.10	94.20	59.40	44.20	25.40	820.20
1961	1.30	5.60	19.60	87.60	76.20	49.00	52.80	161.20	37.80	183.90	413.80	148.10	1236.90
1962	83.00	8.40	83.30	121.70	133.40	100.10	135.60	87.10	89.70	188.40	116.80	58.40	1205.90
1963	64.50	67.10	80.50	170.40	157.70	68.30	26.40	116.30	12.70	13.50	93.00	191.00	1061.40
1964	1.50	43.40	47.80	248.90	121.70	41.70	77.10	111.30	82.80	74.20	71.40	39.60	961.40
1965	36.80	1.30	16.80	115.60	74.40	30.00	39.90	51.30	12.20	79.50	65.90	29.20	552.90
1966	3.60	46.20	59.20	200.90	56.40	92.50	104.60	97.80	105.20	50.50	124.00	9.40	950.30
1967	3.60	0.30	27.90	89.70	138.90	117.30	99.27	52.80	27.40	52.80	94.00	1.30	705.27
1968	0.00	156.00	174.60	286.40	99.40	34.30	126.20	66.40	25.80	29.50	100.60	63.20	1162.40
1969	72.20	69.90	98.50	44.90	150.00	24.90	58.00	60.30	107.60	45.30	52.60	16.00	800.20
1970	136.30	14.00	130.50	181.90	154.90	64.60	76.20	127.40	65.70	59.60	108.60	24.80	1144.50
1971	43.90	0.00	21.70	134.30	134.30	146.70	55.40	173.80	75.30	16.40	51.10	61.30	914.20
1972	17.80	138.90	14.40	21.60	152.70	62.30	90.70	128.70	37.10	59.40	107.70	6.90	838.20
1973	22.90	46.30	1.70	41.50	107.40	20.20	84.00	202.90	124.30	31.60	53.60	4.90	741.30
1974	10.80	14.10	114.80	199.60	70.60	61.10	115.10	284.30	95.70	49.70	43.90	16.70	1076.40
1975	12.80	26.70	23.10	156.80	210.00	158.80	150.00	193.00	81.50	104.80	27.60	48.10	1193.20
1976	3.90	37.50	11.00	110.10	77.20	31.50	106.40	153.60	77.50	24.40	33.90	43.80	710.80
1977	61.30	30.40	16.10	294.10	147.50	45.30	105.60	40.90	49.10	121.40	208.60	48.90	1169.20
1978	83.50	99.90	194.50	165.70	54.60	57.80	57.80	137.80	154.80	77.30	103.80	23.30	1210.80
1979	55.20	165.80	89.20	158.90	85.90	63.30	60.90	80.10	38.10	15.70	66.40	32.30	911.80
1980	54.20	4.90	78.70	125.20	221.40	75.20	18.80	52.90	12.00	19.00	113.60	4.30	780.20
1981	0.80	22.10	112.90	211.70	108.60	36.60	88.10	153.60	118.40	23.20	29.30	34.90	940.20
1982	7.00	21.00	4.90	144.10	154.50	25.60	42.80	231.10	28.60	94.90	139.90	62.20	956.60
1983	24.20	27.90	14.40	119.10	97.90	37.90	73.00	170.60	123.50	62.30	81.70	118.10	950.60
1984	1.30	15.10	6.80	106.50	29.10	34.30	63.20	65.40	60.00	80.40	91.40	34.90	588.40
1985	31.00	44.90	121.90	342.80	104.60	130.50	47.60	64.60	22.20	30.70	70.50	20.50	1031.80
Means	29.91	39.33	67.46	141.26	118.44	76.20	94.14	125.29	68.61	56.33	80.05	45.09	942.12

Appendix A1.5. Time series of the monthly rainfall (mm) at Ogilgei rainfall station No. 0903500

YEAR	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	Annual Totals
1963	93.70	65.20	56.40	193.80	130.70	63.10	23.60	117.60	11.80	18.00	80.60	183.20	1037.70
1964	8.70	63.40	59.00	272.00	113.40	30.70	125.70	107.50	121.50	102.80	71.10	52.00	1127.80
1965	36.10	33.00	108.80	78.60	22.30	54.30	58.90	17.70	7.17	7.17	78.90	37.10	540.04
1966	8.90	59.80	76.60	184.10	66.30	134.80	111.00	123.30	98.90	45.60	140.80	10.90	1061.00
1967	2.00	0.50	49.60	101.20	203.10	100.30	107.30	50.80	30.40	42.70	136.70	2.80	827.40
1968	0.00	148.50	213.90	317.60	57.70	34.90	52.30	74.40	44.30	39.70	89.20	71.90	1144.40
1969	66.50	70.20	95.30	40.50	136.40	32.40	81.60	55.40	83.90	40.20	67.30	20.90	790.60
1970	208.10	11.20	173.30	325.10	147.00	129.30	58.30	169.20	49.70	0.00	0.00	0.00	1271.20
1971	0.00	0.00	0.00	54.90	220.60	289.30	48.20	162.60	74.90	20.60	75.30	85.90	1032.30
1972	17.30	155.00	9.10	38.20	111.70	105.80	11.50	152.30	35.10	67.60	131.70	39.90	875.20
1973	43.50	45.90	0.00	32.10	88.20	26.40	92.40	136.80	128.80	18.40	64.50	9.50	686.50
1974	12.80	13.00	97.90	137.10	67.00	54.70	130.00	177.10	77.30	31.40	36.80	13.30	848.40
1975	2.10	26.10	25.80	132.10	147.10	161.70	168.20	173.90	119.10	74.40	24.70	23.20	1078.40
1976	0.00	27.20	17.00	81.10	99.00	29.70	115.70	130.20	37.70	27.00	34.00	0.00	598.60
1977	3.40	15.20	18.20	440.40	326.10	68.70	93.20	142.80	123.30	103.80	65.02	96.00	1496.12
1978	102.80	199.10	344.40	462.00	20.40	90.50	135.70	234.00	63.44	96.00	9.80	80.00	1838.14
1979	87.20	220.60	57.70	122.80	46.40	77.10	111.70	107.60	0.00	54.60	0.00	30.10	915.80
1980	60.20	2.40	81.60	193.40	117.84	87.28	3.50	49.50	34.60	21.40	64.00	1.10	716.82
Means	41.85	64.24	82.48	178.17	117.85	87.28	84.93	121.26	63.44	45.08	65.02	42.10	993.69

Appendix A1.6. Time series of the monthly rainfall (mm) at Technology Farm rainfall station No. 090

YEAR	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	Annual Totals
1963	81.00	92.90	49.00	211.30	180.00	102.50	45.70	132.40	41.70	31.20	67.90	183.80	1219.40
1964	7.20	53.10	72.00	233.20	91.90	83.50	129.30	111.10	181.70	127.20	105.40	41.60	1237.20
1965	25.10	0.00	32.00	92.20	90.90	42.40	29.10	45.90	11.10	57.10	54.60	53.10	533.50
1966	8.90	29.30	62.80	242.90	27.70	132.50	78.50	150.00	124.40	44.00	85.70	1.00	987.70
1967	4.60	0.00	31.30	115.50	205.10	83.80	89.60	91.30	32.70	34.40	98.80	2.30	789.40
1968	0.00	112.10	193.20	223.60	143.80	63.70	89.10	78.20	41.40	56.70	98.40	71.00	1171.20
1969	58.30	88.50	81.70	41.40	117.60	26.40	56.10	46.30	103.80	29.20	88.90	29.00	767.20
1970	162.20	9.50	184.20	190.00	165.30	77.50	93.80	121.40	108.70	70.10	30.20	24.50	1237.40
1971	57.40	0.00	29.50	143.60	238.80	131.40	59.30	124.20	72.40	20.70	44.50	58.00	979.80
1972	22.00	114.10	7.00	32.20	101.50	132.60	68.00	179.40	38.50	103.30	93.50	29.60	921.70
1973	27.40	56.80	2.00	39.00	166.30	29.60	61.90	187.00	161.10	36.40	68.00	8.40	843.90
1974	8.60	15.20	109.00	158.60	59.60	71.30	97.30	140.70	84.10	34.90	52.60	4.00	835.90
1975	4.50	23.00	17.60	112.80	198.50	130.70	85.10	191.90	85.10	140.50	42.60	23.20	1055.50
1976	0.00	30.80	6.70	86.40	92.70	51.10	112.20	258.00	48.30	5.20	32.70	28.00	752.10
1977	50.30	15.40	13.10	242.30	108.40	64.16	101.10	74.80	46.90	70.10	118.50	43.10	948.16
1978	86.30	111.50	190.80	104.00	59.90	56.90	110.20	164.10	161.70	72.80	55.20	66.40	1239.80
1979	34.70	132.00	8.10	233.20	119.50	79.80	68.70	58.00	23.40	30.80	5.10	14.80	808.10
1980	24.80	2.20	48.80	118.60	361.70	114.90	3.50	49.50	34.60	21.40	64.00	1.10	845.10
1981	0.00	27.50	112.50	174.80	189.60	64.60	133.20	168.10	137.10	56.00	36.60	12.60	1112.60
1982	28.20	0.00	0.00	159.90	165.00	39.60	19.00	167.70	28.90	110.30	119.90	48.10	886.60
1983	1.50	52.80	11.00	63.40	68.70	61.20	59.80	133.40	178.40	197.50	67.20	62.40	957.30
1984	15.10	16.10	6.80	106.50	30.40	45.70	49.00	44.80	94.00	72.10	62.90	0.00	543.40
1985	17.40	16.70	111.50	206.30	139.90	129.90	70.10	65.20	13.80	25.50	52.40	19.20	867.90
Means	31.54	43.46	60.03	144.86	135.77	78.95	74.33	121.02	80.60	62.93	67.20	35.88	936.56

Appendix A1.7. Time series of the monthly rainfall at Nakuru Meteorological Station No. 0936261

YEAR	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	Annual Totals
1964	5.2	30	130.7	207.1	88.9	55.2	180.2	68.2	195.6	114.5	61.6	30.1	1167.3
1965	18.6	2.8	34	107.4	79.6	32	66.7	84.8	22.5	52.4	61.5	52	614.3
1966	14.4	31.5	61.1	263.6	33.8	127.8	89.7	177.8	106.3	60.6	61.2	12.2	1040
1967	2.1	0	32.7	141.6	172.6	73.3	146	91.8	59.1	54	83.1	39	895.3
1968	0	103.9	173.8	205.1	148.4	42.2	82.9	68.4	56.7	66	103.6	55.7	1106.7
1969	93.7	116.7	115.6	27.6	116.1	24.2	39.6	49.9	98.5	36.1	54.6	24.8	797.4
1970	115.7	4.8	206.4	167.3	140	43.5	104.7	72.4	116.4	83.4	35.2	23.7	1113.5
1971	58.2	0	23.7	86.7	147.2	118.5	60	136.9	159.2	17.6	43.3	56	907.3
1972	19.9	100.5	23.4	35	102.5	113.9	81.5	128.2	40.8	102.6	80.4	29.6	858.3
1973	20.1	33.7	3	43.5	115.9	24.7	62.8	188.7	165.9	34	60.2	64	816.5
1974	14.5	17	101.6	138.4	61.7	112.7	137.2	197.2	83.6	56.1	40.5	9.5	970
1975	3.9	7.1	14.7	149.5	144.9	121.2	72	193.3	155.2	163	102.48	44.1	1171.38
1976	29.4	37.9	15.9	104.9	118.5	70	107.2	168.5	81.4	50.6	55	39.7	879
1977	61.1	17.5	22.8	220.8	175.3	78.6	193.1	91.8	58.9	65.6	157.7	74.6	1217.8
1978	56.7	123.2	188.7	136.25	117.8	54.5	97.7	105.6	128.3	100.7	60.2	65.5	1235.15
1979	56.5	165.6	75	120	98.3	91.4	81.9	127.2	53.8	29.1	45.2	14	958
1980	28.6	0.9	36.4	102.3	92.4	82.8	29.3	58.6	18.2	36.9	87.4	10.3	584.1
1981	0.3	26	109.9	138.5	165.3	77.6	138.1	169.2	63.2	72	38.3	117	1115.4
1982	2.6	15	2.9	156.1	188.9	20.7	49.9	191.3	49.3	141.4	186.5	66.7	1071.3
1983	14.8	84.6	15.5	83.3	68.3	65.9	67.1	142.8	142.9	89.5	90.1	68.6	933.4
1984	3.6	6.1	11.8	65.1	32	67	87.6	36	69.4	83	62.1	63.6	587.3
1985	27.8	39.4	96.8	297.5	99.6	123.7	88.5	69.5	27.9	23.6	73.5	8.5	976.3
Means	29.44	43.83	68.02	136.25	114.00	73.70	93.80	119.00	88.78	69.67	74.71	44.05	955.26

Appendix A.2. Time Series of the Annual Rainfall in Enjoro River Watershed (mm)

YEAR	*NPBRS Station*	*EGERTON Station*	*NAKURU Station*	*OGILGEI Station*	*TECFARM Station*	*NESSUIET Station*	*TERETStation*
1960	820.2	839.3					
1961	1236.9	1429.3					
1962	1205.9	983.7					
1963	1061.4	1175.9		1037.7	1219.4	1424.7	1379.1
1964	961.4	1087.7	1167.3	1127.8	1237.2	1197.1	1242.7
1965	552.9	602.3	614.3	540.04	533.5	716.1	583.7
1966	950.3	1039.5	1040	1061	987.7	1213.2	879.1
1967	705.27	902.8	895.3	827.4	789.4	941.1	1105.1
1968	1162.4	1217.4	1106.7	1144.4	1171.2	1315.1	1274.55
1969	800.2	811.9	797.4	790.6	767.2	1183.2	1040.3
1970	1144.5	1125.9	1113.5	1271.2	1237.4	1631.8	1624.8
1971	914.2	971	907.3	1032.3	979.8	1239.1	1194.4
1972	838.2	804.2	858.3	875.2	921.7	961.9	903.3
1973	741.3	825	816.5	686.5	843.9	858.5	675.5
1974	1076.4	1101	970	848.4	835.9	1073.3	1017.2
1975	1193.2	1367.9	1171.38	1078.4	1055.5	1155.4	1157.6
1976	710.8	824.7	879	598.6	752.1	1120.9	708.4
1977	1169.2	1337.6	1217.8	1496.12	948.16	1231.16	1473.1
1978	1210.8	1434.6	1235.15	1838.14	1239.8	1440.3	1280.6
1979	911.8	1101.1	958	915.8	808.1	1274	1406.2
1980	780.2	933.3	584.1	716.82	845.1	760.3	804.7
1981	940.2	929.3	1115.4		1112.6	996.4	943.4
1982	956.6	1157.5	1071.3		886.6	1241.8	1276.2
1983	950.6	961.3	933.4		957.3	1091.3	1094.5
1984	588.4	666.4	587.3		543.4	756.8	611.4
1985	1031.8	1036.9	976.3		867.9	1012.6	1132.1
1986		915.3					
1987		762.9					
1988		1139.3					
1989		1095.3					
1990		1064.2					
Means	946.73	1020.79	955.26	993.69	936.56	1123.31	1078.61

Appendix A3.1 Time series of the monthly Kenya Pan 'A' evaporation (mm/yr) at Egerton weather station.

YEAR	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	Annual Totals	Annual ETo
1965	137.70	165.30	168.70	107.70	112.10	112.40	112.30	117.90	148.90	131.20	89.30	136.60	1540.10	1155.08
1966	153.80	135.90	136.20	130.40	109.80	100.70	92.40	93.20	107.90	118.10	104.80	143.60	1426.80	1070.10
1967	166.10	174.60	158.80	119.60	112.20	102.20	89.80	112.20	113.10	105.10	89.70	136.60	1480.00	1110.00
1968	176.00	108.60	111.00	89.10	98.10	93.90	87.70	84.50	121.50	105.50	92.90	119.90	1288.70	966.53
1969	144.90	105.20	116.60	110.00	101.90	89.20	87.50	87.30	110.50	125.60	92.20	115.10	1286.00	964.50
1970	103.40	144.90	127.50	86.50	105.60	84.40	87.20	90.00	99.50	113.20	91.60	115.20	1249.00	936.75
1971	133.90	152.50	165.60	97.60	70.10	74.90	81.70	94.70	83.80	99.80	100.60	106.40	1261.60	946.20
1972	105.13	142.70	140.90	139.30	113.50	120.40	150.40	154.20	170.60	159.70	152.90	160.60	1710.33	1282.75
1973	214.70	225.70	291.10	270.80	319.80	134.40	122.20	162.50	99.40	90.60	87.80	183.30	2202.30	1651.73
1974	102.30	89.60	139.50	168.00	99.20	96.00	120.00	176.70	120.00	83.70	150.80	83.70	1429.50	1072.13
1975	101.60	135.50	140.50	116.50	102.50	97.10	103.10	119.30	117.30	113.30	114.20	102.20	1363.10	1022.33
1976	100.60	140.40	143.50	120.50	107.40	115.60	106.20	124.10	120.80	120.10	132.10	152.40	1483.70	1112.78
1977	156.00	184.80	170.50	156.60	151.20	219.00	136.40	198.40	159.00	195.30	150.00	220.10	2097.30	1572.98
1978	198.40	131.60	120.90	116.00	156.60	144.00	142.60	204.60	168.00	151.90	162.00	183.00	1879.60	1409.70
1979	176.70	134.40	279.00	156.00	189.10	180.00	198.40	179.80	132.00	102.30	116.00	150.80	1994.50	1495.88
1980	186.00	124.70	164.30	147.00	158.10	165.00	161.20	124.00	216.00	161.20	113.10	151.90	1872.50	1404.38
1981	198.40	190.40	114.70	132.00	161.20	162.00	99.20	62.00	198.00	298.40	123.00	133.30	1872.60	1404.45
1982	151.90	103.60	176.70	108.00	108.50	105.00	102.30	102.30	108.00	93.00	87.00	111.60	1357.90	1018.43
1983	124.00	114.80	158.10	126.00	108.50	99.00	99.20	102.30	117.00	108.50	108.00	102.30	1367.70	1025.78
1984	130.20	159.50	164.30	132.00	133.30	117.00	102.30	111.60	132.00	99.20	90.00	111.60	1483.00	1112.25
1985	142.60	134.40	148.80	120.00	108.50	120.00	99.20	111.60	93.00	111.00	99.00	120.90	1409.00	1056.75
1986	145.70	148.40	142.60	105.00	111.60	87.00	105.40	111.60	120.00	114.00	96.00	108.90	1396.20	1047.15
1987	133.30	148.30	164.30	117.00	108.50	90.00	124.00	124.00	126.00	135.00	99.00	130.20	1499.60	1124.70
1988	130.20	142.10	139.50	108.00	105.40	111.00	89.90	99.20	93.00	114.00	96.00	105.40	1333.70	1000.28
1989	136.40	131.60	155.00	99.00	120.90	108.00	99.20	120.90	165.00	111.00	99.00	105.40	1451.40	1088.55
1990	130.20	142.80	120.90	117.00	120.90	111.00	120.90	120.90	129.00	111.00	102.00	114.70	1441.30	1080.98
1991	148.80	151.20	151.20	108.00	105.40	105.00	86.80	111.60	123.00	111.00	99.00	117.80	1418.80	1064.10
1992	136.40	142.80	142.80	117.00	114.70	108.00	102.30	86.80	108.00	105.00	96.00	102.30	1362.10	1021.58
1993	89.90	109.20	109.20	120.00	120.90	99.00	99.20	114.70	123.00	117.00	99.00	108.90	1310.00	982.50
Mean Epan	143.28	141.91	153.89	125.54	125.36	115.56	110.66	120.79	128.39	124.30	108.03	128.78		
Monthly ETo	107.46	106.44	115.41	94.15	94.02	86.67	82.99	90.59	96.29	93.23	81.03	96.59		

ETo= Kpan * ETpan: Kpan = 0.75

Appendix B.1. Time series of the current meter discharge measurements in the river gauging stations (RGS)

2FC05 GAUGING STATION

Date of Record	Width(m)	Area(m^2)	Velocity(m/s)	Gauge Height(m)	DISCHARGE(m3/s)	Hydraulic Coefficient(k*)
03.08.1949	4.802	1.768	0.686	0.701	1.210	0.296
22.09.1959	4.573	3.058	0.070	0.274	0.214	0.050
11.09.1959	4.878	3.828	0.235	0.448	0.904	0.179
26.09.1959	4.573	2.909	0.052	0.253	0.150	0.036
29.01.1964	3.963	1.604	0.195	0.287	0.312	0.097
05.05.1965	2.134	0.637	0.137	0.183	0.087	0.055
18.09.1967	2.134	0.307	0.098	0.213	0.029	0.023
15.03.1968	4.421	1.992	0.616	0.485	1.226	0.317
30.08.1968	3.506	1.549	0.445	0.405	0.692	0.226
27.02.1969	2.591	0.774	0.393	0.293	0.305	0.154
16.05.1969	3.354	1.504	0.384	0.421	0.600	0.201
23.01.1970	3.201	1.381	0.299	0.338	0.414	0.151
20.04.1970	4.268	2.156	0.591	0.503	1.278	0.326
29.01.1971	1.390	0.303	0.216	0.128	0.065	0.070
16.01.1971	1.192	0.118	0.229	0.082	0.027	0.045
30.06.1971	3.659	1.385	0.326	0.341	0.452	0.152
28.06.1972	2.729	0.973	0.122	0.247	0.118	0.055
22.09.1972	1.223	0.285	0.387	0.183	0.110	0.123
24.02.1973	1.143	0.211	0.171	0.107	0.036	0.049
30.03.1973	0.939	0.070	0.101	0.046	0.007	0.016
20.04.1978	5.701	5.637	0.433	0.591	2.443	0.379
26.01.1979	5.802	4.720	0.143	0.390	0.670	0.114
30.05.1979	5.802	4.575	0.131	0.381	0.599	0.103
06.07.1979	5.902	5.722	0.265	0.530	1.510	0.232
30.08.1979	5.802	4.611	0.119	0.360	0.542	0.093
09.11.1979	5.500	3.680	0.030	0.220	0.114	0.023
21.11.1980	4.802	2.602	0.049	0.201	0.130	0.031
03.06.1980	5.802	3.666	0.095	0.305	0.344	0.065
09.06.1980	5.500	3.903	0.082	0.311	0.324	0.062
15.07.1980	5.402	3.161	0.055	0.250	0.173	0.036
30.12.1980	2.302	0.733	0.040	0.079	0.028	0.017
29.01.1981	2.101	0.495	0.018	0.040	0.009	0.007
23.04.1981	4.802	3.183	0.101	0.299	0.322	0.071
02.04.1981	2.399	0.679	0.110	0.189	0.074	0.043
15.04.1981	5.101	3.882	0.287	0.460	1.105	0.212
14.05.1981	5.101	3.571	0.216	0.412	0.790	0.158
20.07.1981	4.878	3.050	0.079	0.250	0.243	0.054

Appendix B.1 Contd								
	18.08.1981	5.402	5.413	0.561	0.668	3.032	0.484	
	17.09.1981	4.899	3.029	0.049	0.250	0.152	0.034	
	22.10.1981	5.000	3.009	0.055	0.241	0.168	0.037	
	05.11.1981	4.802	2.515	0.040	0.201	0.099	0.024	
	30.12.1981	2.101	0.483	0.055	0.110	0.027	0.020	
	10.02.1982	1.000	0.260	0.061	0.098	0.016	0.022	
	16.03.1982	0.899	0.191	0.116	0.040	0.022	0.039	
	19.04.1982	0.899	0.180	0.445	0.119	0.080	0.130	
	25.05.1982	2.500	0.559	0.207	0.220	0.116	0.069	
	13.08.1982	5.000	3.243	0.143	0.250	0.459	0.100	
	14.09.1982	1.399	0.392	0.546	0.259	0.214	0.190	
	10.02.1983	2.500	0.565	0.088	0.140	0.050	0.031	
	28.09.1983	2.701	0.945	0.428	0.351	0.404	0.182	
	17.02.1984	2.101	0.384	0.064	0.079	0.025	0.020	
	23.03.1984	2.101	0.578	0.057	0.079	0.033	0.023	
	9.04.1984	2.101	0.318	0.035	0.049	0.011	0.010	
	11.07.1984	1.299	0.332	0.084	0.040	0.028	0.033	
	21.09.1984	1.399	0.230	0.087	0.049	0.020	0.025	
	07.11.1984	2.101	0.608	0.044	0.061	0.027	0.019	
	21.12.1984	0.649	0.062	0.291	0.070	0.018	0.054	
	12.02.1985	0.799	0.170	0.099	0.018	0.017	0.035	
	08.04.1985	1.399	0.561	0.690	0.320	0.387	0.292	
	12.05.1985	2.000	0.433	0.540	0.290	0.234	0.164	
	01.11.1985	2.802	1.530	0.229	0.290	0.350	0.135	
	05.12.1985	0.899	0.008	0.247	0.119	0.020	0.092	
	10.02.1986	0.799	0.057	0.070	0.040	0.004	0.011	
	04.03.1986	0.899	0.067	0.075	0.021	0.005	0.013	
	26.05.1986	0.701	0.062	0.433	0.110	0.027	0.072	
	02.09.1986	2.250	0.730	0.440	0.299	0.321	0.178	
	06.10.1986	2.201	0.559	0.199	0.229	0.111	0.070	
	19.01.1988	1.899	0.390	0.192	0.259	0.075	0.057	
	25.10.1988	4.601	1.919	0.277	0.360	0.531	0.140	
	20.02.1989	4.500	1.135	0.089	0.229	0.101	0.033	
	03.10.1989	1.800	1.695	0.261	0.350	0.442	0.201	
	13.06.1990	2.900	0.875	0.309	0.300	0.270	0.122	
	13.08.1990	1.200	0.234	0.350	0.190	0.082	0.098	
	17.05.1991	1.900	0.291	0.144	0.080	0.042	0.039	
	23.09.1992	3.200	1.351	0.182	0.260	0.246	0.093	
					* k =Q/A*R^0.667 (see equation 4..)			

Appendix B.2. Time series of the current meter discharge measurements at Little Shuru tributary RGS No. 2FC

Date of Record	Width(m)	Area(m^2)	Velocity (m/s)	Gauge Height (m)	Discharge (m3/s)	Hydraulic coefficient(k*)
28.04.1966	2.134	0.432	0.140	0.274	0.061	0.051
19.05.1966	1.982	0.217	0.070	0.171	0.015	0.020
27.07.1966	2.363	0.227	0.058	0.177	0.013	0.017
12.10.1966	2.439	0.251	0.082	0.192	0.021	0.025
09.06.1967	1.982	0.240	0.189	0.195	0.046	0.057
15.03.1968	2.134	0.478	0.107	0.259	0.050	0.037
10.05.1968	3.405	2.006	0.241	0.832	0.485	0.164
30.08.1968	2.034	0.518	0.287	0.354	0.148	0.117
27.02.1969	2.591	0.412	0.155	0.262	0.064	0.056
16.05.1969	2.268	0.545	0.299	0.335	0.163	0.121
23.01.1970	2.210	0.358	0.195	0.259	0.070	0.069
22.07.1970	2.134	0.421	0.515	0.232	0.187	0.147
29.01.1971	2.439	0.271	0.061	0.180	0.016	0.017
16.04.1971	1.448	0.119	0.064	0.037	0.008	0.007
30.06.1971	2.134	0.410	0.229	0.250	0.093	0.078
22.03.1972	1.433	0.129	0.030	0.107	0.012	0.020
28.06.1972	2.235	0.191	0.110	0.137	0.021	0.027
26.08.1972	2.110	0.378	0.290	0.195	0.111	0.088
08.12.1972	2.134	0.241	0.098	0.116	0.023	0.021
24.02.1973	1.677	0.164	0.055	0.095	0.009	0.010
30.03.1973	1.119	0.074	0.067	0.073	0.005	0.011
10.08.1973	1.982	0.342	0.262	0.210	0.089	0.081
26.02.1974	0.899	0.048	0.040	0.049	0.002	0.005
18.05.1974	1.381	0.093	0.162	0.070	0.015	0.026
08.07.1974	2.104	0.480	0.302	0.250	0.145	0.104
25.03.1975	1.601	0.011	0.046	0.070	0.005	0.073
26.05.1975	1.899	0.152	0.049	0.079	0.011	0.013
11.06.1975	1.802	0.168	0.119	0.079	0.020	0.021
15.08.1975	2.683	0.583	0.424	0.290	0.246	0.162
06.11.1975	2.601	0.651	0.232	0.280	0.150	0.087
21.01.1976	2.101	0.153	0.079	0.091	0.012	0.015
28.06.1976	1.899	0.198	0.098	0.091	0.019	0.019
22.11.1976	1.899	0.198	0.098	0.091	0.019	0.019
06.07.1977	2.302	0.529	0.247	0.259	0.130	0.087
20.04.1978	2.802	1.275	0.552	0.500	0.704	0.284
26.05.1978	2.601	0.620	0.284	0.290	0.175	0.108
15.02.1979	2.899	1.177	0.716	0.500	0.842	0.370
30.05.1979	2.500	0.668	0.280	0.299	0.187	0.108
06.07.1979	2.601	0.857	0.369	0.390	0.317	0.166
30.08.1979	2.762	0.614	0.253	0.311	0.155	0.101
09.11.1979	2.802	0.487	0.113	0.189	0.055	0.034
03.01.1980	2.360	0.335	0.058	0.149	0.019	0.015
15.02.1980	2.201	0.232	0.061	0.131	0.014	0.014
03.06.1980	2.000	0.244	0.079	0.140	0.019	0.019
09.06.1980	2.302	0.311	0.110	0.171	0.034	0.031

	Appendix B.2 Contd						
	15.07.1980	2.201	0.307	0.110	0.171	0.034	0.031
	21.11.1980	2.201	0.300	0.101	0.171	0.030	0.028
	30.12.1980	1.899	0.161	0.043	0.101	0.007	0.009
	29.01.1981	1.302	0.081	0.067	0.070	0.005	0.011
	02.04.1981	2.101	0.276	0.101	0.162	0.028	0.027
	15.04.1981	2.601	0.829	0.412	0.360	0.342	0.177
	23.04.1981	2.399	0.462	0.183	0.229	0.084	0.061
	20.07.1981	2.101	0.252	0.104	0.149	0.026	0.027
	18.08.1981	2.701	0.930	0.564	0.412	0.525	0.262
	17.09.1981	2.101	0.315	0.098	0.189	0.031	0.029
	22.10.1981	2.201	0.293	0.159	0.180	0.046	0.045
	05.11.1981	2.101	0.255	0.110	0.140	0.028	0.027
	30.12.1981	1.701	0.175	0.058	0.119	0.010	0.013
	16.03.1982	1.500	0.204	0.040	0.079	0.008	0.007
	19.04.1982	2.000	0.125	0.070	0.131	0.015	0.028
	25.05.1982	2.101	0.264	0.114	0.150	0.030	0.029
	30.06.1982	1.899	0.213	0.085	0.131	0.018	0.020
	13.08.1982	2.302	0.433	0.232	0.241	0.101	0.080
	14.09.1982	2.201	0.345	0.165	0.189	0.057	0.049
	21.10.1982	2.302	0.466	0.308	0.250	0.143	0.107
	10.02.1983	2.195	0.279	0.073	0.159	0.020	0.019
	28.09.1983	2.501	0.634	0.311	0.290	0.197	0.118
	17.02.1984	2.101	0.251	0.064	0.131	0.016	0.015
	23.03.1984	2.201	0.250	0.040	0.119	0.010	0.009
	09.04.1984	2.000	0.182	0.027	0.101	0.005	0.006
	22.05.1984	1.899	0.196	0.030	0.110	0.006	0.007
	21.09.1984	1.701	0.162	0.030	0.101	0.005	0.006
	21.12.1984	1.802	0.173	0.058	0.101	0.010	0.012
	18.04.1985	1.802	0.494	0.203	0.241	0.101	0.068
	22.05.1985	2.000	0.346	0.192	0.201	0.066	0.058
	11.11.1985	1.899	0.297	0.227	0.220	0.067	0.072
	05.12.1985	1.701	0.190	0.132	0.149	0.025	0.033
	10.02.1986	0.701	0.065	0.092	0.080	0.006	0.015
	04.03.1986	0.701	0.037	0.108	0.079	0.004	0.017
	08.04.1986	0.701	0.043	0.130	0.090	0.006	0.022
	02.09.1986	1.899	0.251	0.116	0.171	0.029	0.032
	27.03.1987	0.841	0.065	0.062	0.091	0.004	0.011
	19.01.1988	1.899	0.394	0.174	0.229	0.070	0.058
	20.02.1989	2.000	0.323	0.071	0.159	0.023	0.019
	13.06.1990	2.400	0.534	0.178	0.240	0.095	0.061
	13.08.1990	3.100	0.496	0.177	0.230	0.088	0.061
	03.10.1990	2.300	0.496	0.177	0.230	0.088	0.059
	17.05.1991	2.200	0.390	0.051	0.140	0.020	0.013
	14.08.1991	2.500	0.733	0.270	0.300	0.198	0.105
	23.09.1992	2.400	0.416	0.050	0.170	0.021	0.014

Appendix B.3. Time series of the current meter discharge measurements at the river gauging station 2FC09

Date of Record	Width(m)	Area(m^2)	Velocity(m/s)	Gauge Height(m)	Discharge(m3/s)	Hydraulic coefficient (k')
05.09.1974	11.585	6.683	0.390	0.762	2.611	0.291
13.11.1974	8.232	2.273	0.165	0.524	0.377	0.097
21.11.1974	7.866	1.787	0.640	0.412	0.147	0.041
06.07.1977	8.201	2.452	0.095	0.521	0.235	0.056
30.05.1979	8.701	2.701	0.207	0.570	0.557	0.127
13.06.1979	8.302	4.200	0.073	0.521	0.297	0.041
06.07.1979	8.701	3.864	0.387	0.671	1.492	0.261
30.08.1979	8.604	2.591	0.183	0.561	0.473	0.111
15.07.1980	8.101	2.233	0.085	0.530	0.190	0.050
21.11.1980	8.101	1.758	0.052	0.390	0.093	0.026
21.01.1981	0.601	0.032	0.107	0.091	0.003	0.018
02.04.1981	8.003	1.343	0.076	0.360	0.103	0.036
23.04.1981	8.503	2.272	0.162	0.521	0.367	0.094
21.05.1981	4.701	3.347	0.247	0.601	0.684	0.122
20.07.1981	4.701	2.344	0.049	0.381	0.111	0.022
18.08.1981	11.503	7.614	0.402	0.799	3.067	0.309
17.09.1981	8.201	2.056	0.067	0.421	0.138	0.034
22.10.1981	8.104	1.941	0.061	0.399	0.122	0.031
05.11.1981	7.402	1.845	0.049	0.360	0.092	0.023
10.02.1982	0.799	0.061	0.116	0.110	0.007	0.022
25.05.1982	8.302	1.762	0.085	0.399	0.151	0.043
13.08.1982	8.802	2.373	0.174	0.570	0.412	0.107
14.09.1982	8.402	2.016	0.091	0.500	0.181	0.051
21.10.1982	10.701	3.785	0.348	0.659	1.309	0.235
09.02.1983	7.104	1.193	0.046	0.220	0.056	0.016
28.09.1983	7.701	2.358	0.180	0.520	0.424	0.104
05.01.1984	2.802	0.633	0.515	0.479	0.326	0.255
11.02.1984	2.101	0.178	0.140	0.180	0.025	0.039
23.03.1984	1.601	0.130	0.062	0.171	0.008	0.016
11.07.1984	0.799	0.049	0.062	0.119	0.010	0.041
21.12.1984	2.500	0.272	0.092	0.229	0.025	0.030
12.02.1985	0.750	0.046	0.152	0.101	0.007	0.027
22.05.1985	2.302	0.273	0.288	0.491	0.258	0.463
11.11.1985	1.799	1.118	0.150	0.421	0.170	0.066
05.12.1985	0.799	0.220	0.132	0.229	0.029	0.037
26.05.1986	0.899	0.190	0.131	0.280	0.025	0.042
02.09.1986	3.302	0.538	0.468	0.509	0.252	0.244
06.10.1986	3.500	0.362	0.320	0.412	0.116	0.150
19.01.1988	2.802	0.617	0.418	0.500	0.258	0.212
25.10.1988	3.302	1.028	0.709	0.561	0.730	0.390
20.02.1989	5.800	1.301	0.076	0.370	0.099	0.035
03.10.1989	2.300	0.705	0.784	0.550	0.553	0.408
13.08.1990	3.600	0.241	0.320	0.390	0.077	0.145
28.05.1991	2.000	0.215	0.326	0.310	0.070	0.122
14.08.1991	7.400	3.486	0.245	0.500	0.855	0.138
23.09.1992	3.100	0.783	0.352	0.480	0.276	0.177

Contd. Appendix B.4												
1968	0.062	0.159	1.248	8.531	3.288	0.548	0.396	1.646	0.226	0.095	0.307	1.987	1.541
1969	0.153	0.241	0.200	0.144	0.449	0.091	0.063	0.074	0.179	0.058	0.066	0.052	0.148
1970	0.157	0.082	0.586	2.266	3.368	1.758	0.596	1.524	1.392	0.818	0.419	0.123	1.091
1971	0.075	0.044	0.034	0.060	0.746	0.909	1.057	7.732	4.760	0.348	0.128	0.180	1.339
1972	0.136	0.460	0.208	0.044	0.045	0.066	0.356	2.604	1.600	0.111	0.725	0.116	0.539
1973	0.076	0.030	0.029	0.030	0.031	0.041	0.050	0.265	0.410	0.143	0.074	0.069	0.104
1974	0.046	0.022	0.065	0.108	0.111	0.145	0.278	0.411	0.936	0.443	0.346	0.205	0.260
1975	0.064	0.014	0.008	0.137	0.113	0.250	0.312	0.412	0.938	0.444	0.279	0.113	0.257
1976	0.042	0.026	0.016	0.028	0.031	0.188	0.345	0.874	0.970	0.137	0.052	0.053	0.230
1977	0.065	0.038	0.023	1.075	4.886	0.519	1.892	2.561	0.373	0.205	2.686	1.506	1.319
1978	0.980	0.067	0.024	0.551	1.077	0.680	1.422	1.540	1.821	0.765	0.624	0.627	0.848
1979	0.040	0.010	0.026	0.308	0.602	0.380	0.794	1.396	0.286	0.120	0.098	0.090	0.346
1980	0.025	0.052	0.045	0.065	1.162	0.274	0.234	0.080	0.055	0.030	0.025	0.026	0.173
1981	0.020	0.005	0.009	0.013	0.232	0.055	0.047	2.144	1.182	0.219	0.104	0.056	0.341
1982	0.032	0.008	0.025	0.131	0.237	0.239	0.042	0.718	0.343	0.275	1.717	3.832	0.633
1983	0.985	0.073	0.030	0.060	0.220	0.229	0.083	0.854	0.500	1.080	0.600	0.043	0.396
1984	0.020	0.033	0.018	0.058	0.092	0.158	0.224	0.272	0.137	0.020	0.034	0.036	0.092
1985	0.027	0.018	0.188	0.349	0.487	0.421	0.272	0.330	0.196	0.062	0.049	0.030	0.202
1986	0.010	0.003	0.008	0.023	0.070	0.078	0.050	0.175	0.511	0.203	0.011	0.085	0.102
1987	0.043	0.001	0.005	0.014	0.060	0.339	0.077	0.047	0.023	0.004	0.036	0.517	0.097
1988	0.250	0.011	0.004	0.016	0.042	0.054	0.065	0.033	0.012	0.001	0.029	0.093	0.051
1989	0.044	0.030	0.019	0.225	0.501	0.218	0.695	0.637	1.359	0.583	0.442	0.209	0.414
1990	0.112	0.177	0.377	5.260	1.413	0.356	0.174	0.113	0.113	0.096	0.067	0.030	0.691
Means	0.207	0.085	0.103	0.682	0.870	0.375	0.537	1.271	0.957	0.311	0.483	0.542	
Ki	0.356	0.146	0.177	1.171	1.496	0.644	0.922	2.184	1.644	0.535	0.830	0.931	
Zi	-1.462	-1.940	-1.869	0.389	1.125	-0.808	-0.176	2.687	1.462	-1.057	-0.385	-0.156	
Dev.	-0.375	-0.497	-0.479	0.100	0.288	-0.207	-0.045	0.689	0.375	-0.271	-0.099	-0.040	

Where Qm =ith month mean flow, Ki = Qi/Qm, Z-score (Zi) = (Qi-QM)/Qvar, Dev = Qi-Qm and Qi is the ith monthly flow

Appendix B.5. Time series of mean monthly discharges(m3/s) from RGS 2FC09 in Enjoro River

YEAR	*JAN*	*FEB*	*MAR*	*APR*	*MAY*	*JUN*	*JUL*	*AUG*	*SEP*	*OCT*	*NOV*	*DEC*	***Annual Mean***
1960		-----	-----	-----	0.056	0.047	0.053	0.451	1.055	0.066	0.032	0.022	0.223
1961	0.327	0.122	-----	0.328	-----	-----	0.223	-----	0.04	0.146	0.12	-----	0.187
1962	-----	0.153	-----	0.069	0.481	0.189	0.324	0.695	0.621	0.27	0.115	0.145	0.306
1963	0.578	0.111	0.07	0.909	3.166	2.335	0.17	0.494	0.643	0.097	0.084	-----	0.787
1964	-----	0.122	0.1	0.328	-----	-----	0.223	-----	-----	-----	0.12	0.078	0.162
1965	-----	0.122	0.09	0.328	-----	-----	0.223	-----	-----	-----	0.12	0.078	0.160
1966	-----	-----	0.09	-----	-----	-----	0.223	0.352	-----	0.094	0.164	0.078	0.167
1967	0.327	-----	0.004	0.008	0.073	0.111	-----	0.84	0.099	-----	0.12	0.078	0.184
1968		-----	-----	-----	-----	0.382	0.229	1.525	0.207	0.072	0.12	0.078	0.373
1969	0.578	0.136	0.115	-----	-----	-----	0.043	0.084	0.156	0.037	0.043	0.04	0.137
1970	0.075	0.043	0.177	-----	-----	-----	0.521	2.289	1.923	1.198	0.283	0.104	0.735
1971	0.086	0.052	0.009	0.019	0.58	0.583	0.701	-----	-----	0.407	0.112	0.136	0.269
1972	0.089	0.899	0.047	0.011	0.039	0.036	0.098	1.424	0.511	0.077	0.467	-----	0.336
1973	-----	0.018	-----	-----	0.008	0.019	0.02	0.199	0.499	0.068	0.066	-----	0.112
1974	-----	-----	-----	-----	0.019	0.063	-----	1.202	1.52	0.403	0.16	-----	0.561
1975	-----	-----	-----	-----	0.03	0.243	0.824	-----	-----	0.947	0.276	0.116	0.406
1976		0.013	0.006	0.015	0.017	0.016	0.147	0.604	1.578	0.179	0.049	-----	0.262
1977	0.027	0.017	0.006	-----	-----	0.365	-----	-----	0.362	0.157	-----	1.341	0.325
1978	0.769	0.478	-----	-----	1.503	0.234	1.739	-----	-----	1.334	0.343	0.943	0.918
1979	0.812	-----	0.745	-----	1.515	0.338	0.734	1.296	0.272	0.119	0.085	0.075	0.599
1980	0.056	0.047	0.095	0.158	-----	0.486	-----	-----	0.22	0.117	0.055	0.035	0.141
1981	0.009	-----	-----	0.299	0.408	0.079	0.134	-----	-----	-----	-----	-----	0.186
1982	-----	-----	-----	-----	-----	0.199	0.044	0.699	0.294	0.405	-----	-----	0.328
1983	0.123	-----	0.026	-----	0.05	-----	-----	-----	-----	0.017	-----	-----	0.054
1984	-----	0.025	-----	-----	-----	-----	0.015	-----	0.011	0.031	0.053	0.03	0.028
1985	-----	-----	-----	0.046	0.188	0.641	0.208	-----	-----	0.041	0.063	0.032	0.174
1986	-----	-----	-----	-----	0.065	0.092	-----	0.144	0.447	0.097	0.024	0.041	0.130
1987	-----	-----	-----	-----	0.059	0.46	0.074	0.026	0.022	0.01	0.036	-----	0.098
1988	0.006	0.001	-----	0.017	-----	-----	-----	-----	0.002	-----	-----	0.002	0.006
1989	-----	-----	-----	-----	1.928	0.442	0.327	0.554	0.543	0.574	-----	-----	0.728
1990	-----	0.179	0.26	-----	-----	0.267	0.137	0.11	0.094	0.069	0.06	-----	0.147
Means	0.276	0.149	0.123	0.195	0.566	0.347	0.310	0.722	0.505	0.270	0.127	0.182	
Qmi/Q	0.941	0.510	0.419	0.666	1.931	1.183	1.057	2.463	1.725	0.923	0.433	0.620	

Appendix C.1. Monthly reference evapotranspiration estimated with Penman-Monteith Method (Egerton Weather Station)

MONTH	WIND(kph)	TEMP(oC)	RH(%)	ea(Tmin)	ea(Tmax)	ea	ed	VPC	n(hr)	N(hr)	Ra	Rns	Rnl	Rn	g	ea-ed	DIF	P-M ETo(mm/day)	Days	Monthly ETo	ETm=ETo*kc*f
1977	6.70	20.35	74	20.00	28.10	2.41	1.73	0.11	5.70	11.95	15.00	4.29	8.15613E-05	4.29	0.054	0.68	0.29	2.78	31	86.09	63.93
2	7.40	21.30	75	22.00	29.80	2.59	1.90	0.12	7.20	12.00	15.50	4.99	9.3056E-05	4.99	0.054	0.69	0.31	2.96	28	82.87	61.53
3	7.60	21.20	72	21.30	29.80	2.56	1.79	0.12	4.60	12.05	15.70	4.05	6.6976E-05	4.05	0.054	0.77	0.31	3.26	31	101.20	75.15
4	5.30	17.70	70	14.00	28.10	2.11	1.31	0.10	4.00	12.15	15.30	3.71	7.04578E-05	3.71	0.054	0.80	0.25	2.94	30	88.28	65.55
5	5.70	17.85	79	16.50	24.90	2.07	1.57	0.10	6.90	12.20	14.40	4.49	9.91407E-05	4.49	0.054	0.50	0.26	2.03	31	62.97	46.76
6	5.60	18.40	79	18.20	24.02	2.11	1.64	0.10	6.50	12.25	13.90	4.19	9.18708E-05	4.19	0.054	0.48	0.26	1.89	30	56.71	42.11
7	5.20	17.45	78	16.90	23.40	2.02	1.53	0.10	5.80	12.20	14.10	4.02	8.70213E-05	4.02	0.054	0.48	0.25	1.86	31	57.80	42.92
8	6.10	18.25	72	17.40	24.90	2.12	1.47	0.10	5.80	12.20	14.80	4.22	8.86864E-05	4.22	0.054	0.64	0.27	2.61	31	80.83	60.02
9	6.20	16.75	64	13.30	26.40	1.99	1.13	0.10	5.70	12.10	15.30	4.35	9.89414E-05	4.35	0.054	0.85	0.26	3.52	30	105.70	78.49
10	6.40	19.85	75	18.20	28.30	2.33	1.66	0.11	6.60	12.05	15.40	4.72	9.35054E-05	4.72	0.054	0.66	0.28	2.71	31	84.04	62.41
11	4.90	17.90	78	18.20	21.00	1.96	1.52	0.09	3.50	12.00	15.10	3.50	5.99589E-05	3.50	0.054	0.44	0.24	1.64	30	49.22	36.55
12	6.00	18.85	69	19.40	24.00	2.17	1.48	0.10	5.60	11.95	14.80	4.19	8.74947E-05	4.19	0.054	0.69	0.27	2.75	31	85.14	63.22
1978	6.90	18.95	69	18.20	26.40	2.23	1.49	0.11	6.90	11.95	15.00	4.73	0.000103695	4.73	0.054	0.74	0.29	3.18	31	98.43	73.09
14	6.70	20.35	75	19.40	29.80	2.46	1.76	0.12	6.90	12.00	15.50	4.88	9.40766E-05	4.88	0.054	0.70	0.29	2.87	28	80.33	59.65
15	5.40	19.90	70	19.40	28.10	2.38	1.61	0.11	1.50	12.05	15.70	2.87	3.40653E-05	2.87	0.054	0.77	0.27	2.72	31	84.29	62.59
16	5.20	18.80	75	18.20	26.40	2.23	1.62	0.11	2.60	12.15	15.30	3.20	4.68607E-05	3.20	0.054	0.61	0.26	2.21	30	66.39	49.30
17	5.60	15.85	81	12.30	26.40	1.94	1.36	0.09	6.00	12.20	14.40	4.18	9.48129E-05	4.18	0.054	0.58	0.25	2.32	31	71.78	53.30
18	5.70	15.60	84	12.20	26.00	1.91	1.40	0.09	5.50	12.25	13.90	3.86	8.70168E-05	3.86	0.054	0.51	0.25	2.10	30	63.13	46.88
19	5.00	14.05	78	10.80	23.40	1.71	1.15	0.08	5.00	12.20	14.10	3.75	8.79063E-05	3.75	0.054	0.56	0.23	2.17	31	67.34	50.00
20	5.20	14.90	78	11.10	24.90	1.80	1.20	0.09	5.60	12.20	14.80	4.15	9.47335E-05	4.15	0.054	0.60	0.24	2.37	31	73.37	54.49
21	6.10	15.25	69	11.10	26.40	1.88	1.08	0.09	5.60	12.10	15.30	4.31	9.93621E-05	4.31	0.054	0.80	0.26	3.33	30	99.94	74.22
22	5.60	14.90	69	10.70	26.40	1.86	1.05	0.09	4.50	12.05	15.40	3.94	8.4699E-05	3.94	0.054	0.80	0.25	3.21	31	99.36	73.78
23	5.80	14.85	69	11.10	24.90	1.80	1.06	0.09	3.90	12.00	15.10	3.65	7.59995E-05	3.64	0.054	0.74	0.25	3.04	30	91.12	67.67
24	6.00	17.00	69	16.50	24.90	2.07	1.37	0.10	5.10	11.95	14.80	4.01	8.42945E-05	4.01	0.054	0.70	0.26	2.82	31	87.53	65.00
1979	5.60	15.25	69	12.10	26.40	1.93	1.15	0.09	4.10	11.95	15.00	3.70	7.68489E-05	3.70	0.054	0.78	0.25	3.07	31	95.10	70.62
26	6.10	15.65	66	10.70	26.40	1.86	1.01	0.09	5.70	12.00	15.50	4.42	0.000104095	4.42	0.054	0.85	0.26	3.56	28	99.62	73.97
27	7.20	16.00	61	11.20	28.20	1.97	0.98	0.10	6.80	12.05	15.70	4.89	0.000121102	4.89	0.054	0.99	0.28	4.44	31	137.62	102.19
28	5.80	15.80	65	11.50	28.10	1.98	1.06	0.10	4.90	12.15	15.30	4.04	8.96011E-05	4.04	0.054	0.92	0.26	3.64	30	109.21	81.10
29	5.40	15.30	75	11.10	26.40	1.88	1.17	0.09	4.90	12.20	14.40	3.80	8.59485E-05	3.80	0.054	0.70	0.25	2.75	31	85.18	63.25
30	5.70	15.10	74	11.20	26.00	1.86	1.16	0.09	4.70	12.25	13.90	3.59	8.33258E-05	3.59	0.054	0.70	0.25	2.83	30	84.79	62.97
31	5.40	14.90	73	11.10	24.90	1.80	1.12	0.09	4.50	12.20	14.10	3.58	8.18832E-05	3.58	0.054	0.68	0.24	2.69	31	83.25	61.82
32	5.40	14.95	65	10.70	26.40	1.86	0.99	0.09	4.10	12.20	14.80	3.62	7.98448E-05	3.62	0.054	0.87	0.24	3.35	31	104.00	77.23
33	6.30	15.90	70	11.10	28.10	1.96	1.11	0.10	5.60	12.10	15.30	4.31	9.81476E-05	4.31	0.054	0.85	0.27	3.54	30	106.23	78.88
34	6.50	16.25	57	11.30	29.80	2.06	0.93	0.10	5.80	12.05	15.40	4.42	0.000107881	4.42	0.054	1.12	0.27	4.65	31	144.15	107.04
35	6.10	15.90	61	12.30	24.90	1.86	1.00	0.09	3.90	12.00	15.10	3.65	7.74716E-05	3.64	0.054	0.86	0.26	3.55	30	106.43	79.03
36	7.40	16.70	56	16.10	28.30	2.22	1.15	0.11	6.70	11.95	14.80	4.59	0.000113492	4.59	0.054	1.07	0.30	4.66	31	144.47	107.28
Jun-09	8.30	18.55	53	14.00	31.70	2.29	1.03	0.11	6.20	11.95	15.00	4.47	0.000110933	4.47	0.054	1.26	0.32	5.69	31	176.50	131.07
38	7.80	18.40	63	14.00	31.70	2.29	1.22	0.11	7.30	12.00	15.50	5.03	0.000118489	5.03	0.054	1.06	0.31	4.72	28	132.17	98.14
39	7.90	18.40	81	14.00	31.70	2.29	1.57	0.11	5.80	12.05	15.70	4.51	8.66439E-05	4.51	0.054	0.71	0.31	3.24	31	100.35	74.52
40	6.70	19.15	66	16.10	29.80	2.30	1.38	0.11	4.30	12.15	15.30	3.82	7.26216E-05	3.82	0.054	0.92	0.29	3.73	30	111.83	83.04
41	5.20	18.90	75	17.00	28.10	2.28	1.59	0.11	3.90	12.20	14.40	3.45	6.26691E-05	3.45	0.054	0.67	0.26	2.39	31	74.14	55.05
42	5.00	18.40	78	16.50	26.20	2.14	1.58	0.10	4.90	12.25	13.90	3.66	7.45937E-05	3.66	0.054	0.56	0.25	2.02	30	60.53	44.94
43	5.50	17.15	68	15.00	24.90	2.00	1.27	0.10	6.70	12.20	14.10	4.33	0.000106926	4.33	0.054	0.72	0.25	2.81	31	87.13	64.70
44	5.90	19.30	72	18.40	26.70	2.26	1.57	0.11	5.50	12.20	14.80	4.12	8.23105E-05	4.12	0.054	0.69	0.27	2.67	31	82.89	61.56
45	7.00	20.20	61	18.00	31.50	2.48	1.40	0.12	6.00	12.10	15.30	4.46	9.42235E-05	4.46	0.054	1.08	0.30	4.38	30	131.45	97.61
46	7.00	20.90	57	19.40	31.70	2.56	1.37	0.12	5.10	12.05	15.40	4.16	8.36693E-05	4.16	0.054	1.18	0.30	4.72	31	146.40	108.71
47	5.80	16.25	69	13.10	24.90	1.90	1.18	0.09	4.70	12.00	15.10	3.94	8.39185E-05	3.94	0.054	0.72	0.25	2.90	30	86.98	64.59
48	7.60	17.15	65	12.20	29.80	2.10	1.13	0.10	5.20	11.95	14.80	4.05	9.30522E-05	4.05	0.054	0.97	0.30	4.37	31	135.34	100.50

Contd. Appendix C.

1981	8.60	17.20	53	11.50	33.60	2.26	0.91	0.11	8.30	11.95	15.00	5.24	0.000148061	5.24	0.054	1.35	0.32	6.27	31	194.31	144.29
50	8.20	17.55	53	11.50	33.60	2.26	0.91	0.11	7.50	12.00	15.50	5.10	0.000135278	5.10	0.054	1.35	0.31	6.11	28	171.18	127.12
51	6.80	17.55	65	13.10	31.70	2.24	1.21	0.11	4.20	12.05	15.70	3.90	7.61863E-05	3.90	0.054	1.03	0.29	4.27	31	132.33	98.26
52	5.40	17.45	70	14.00	28.10	2.11	1.31	0.10	5.30	12.15	15.30	4.19	8.75784E-05	4.19	0.054	0.80	0.26	3.00	30	89.85	66.72
53	5.20	16.00	79	12.30	26.40	1.94	1.33	0.09	6.10	12.20	14.40	4.21	9.72042E-05	4.21	0.054	0.61	0.24	2.34	31	72.63	53.93
54	5.80	15.60	73	11.10	28.00	1.96	1.16	0.10	7.30	12.25	13.90	4.46	0.000118988	4.46	0.054	0.79	0.26	3.20	30	95.92	71.23
55	4.80	14.55	83	10.80	23.80	1.73	1.23	0.09	4.50	12.20	14.10	3.58	7.87891E-05	3.58	0.054	0.50	0.23	1.89	31	58.69	43.58
56	5.40	14.65	73	11.30	24.90	1.81	1.13	0.09	5.10	12.20	14.80	3.98	8.98425E-05	3.98	0.054	0.68	0.24	2.68	31	83.12	61.72
57	6.00	16.00	69	12.00	26.40	1.92	1.14	0.09	5.60	12.10	15.30	4.31	9.73213E-05	4.31	0.054	0.78	0.26	3.22	30	96.53	71.68
58	6.30	17.10	65	13.30	26.40	1.99	1.15	0.10	6.60	12.05	15.40	4.72	0.000111287	4.72	0.054	0.84	0.27	3.50	31	108.41	80.51
59	6.30	17.25	65	14.00	26.40	2.02	1.19	0.10	5.30	12.00	15.10	4.16	9.21148E-05	4.16	0.054	0.83	0.27	3.44	30	103.16	76.60
60	6.80	16.85	64	12.80	28.10	2.05	1.13	0.10	6.70	11.95	14.80	4.59	0.000114421	4.59	0.054	0.92	0.28	3.95	31	122.46	90.94
1982	7.50	17.00	52	11.50	31.70	2.16	0.88	0.11	7.80	11.95	15.00	5.06	0.00014191	5.06	0.054	1.28	0.30	5.65	31	175.05	129.99
62	7.60	17.75	53	11.90	33.60	2.28	0.93	0.11	7.00	12.00	15.50	4.91	0.000126569	4.91	0.054	1.34	0.30	5.84	28	163.58	121.47
63	8.40	18.10	42	11.50	35.70	2.36	0.73	0.11	7.70	12.05	15.70	5.23	0.000147026	5.23	0.054	1.63	0.32	7.34	31	227.45	168.90
64	5.80	17.30	65	14.00	28.10	2.11	1.21	0.10	5.10	12.15	15.30	4.12	8.76952E-05	4.12	0.054	0.89	0.26	3.46	30	103.86	77.12
65	4.90	16.40	74	13.10	26.20	1.97	1.29	0.10	5.70	12.20	14.40	4.08	9.3034E-05	4.08	0.054	0.67	0.24	2.46	31	76.40	56.74
66	5.60	15.90	73	11.20	26.50	1.89	1.15	0.09	7.00	12.25	13.90	4.36	0.000115306	4.36	0.054	0.74	0.25	2.94	30	88.30	65.57
67	5.40	15.10	72	10.80	26.40	1.86	1.10	0.09	6.80	12.20	14.10	4.36	0.000114726	4.36	0.054	0.76	0.24	2.97	31	92.22	68.48
68	5.00	15.45	83	12.30	24.90	1.86	1.37	0.09	5.30	12.20	14.80	4.05	8.55747E-05	4.05	0.054	0.49	0.24	1.90	31	58.94	43.76
69	5.90	15.85	69	11.50	27.20	1.94	1.12	0.09	6.40	12.10	15.30	4.61	0.0001094	4.61	0.054	0.82	0.26	3.34	30	100.19	74.40
70	5.30	15.95	69	12.30	26.40	1.94	1.16	0.09	4.80	12.05	15.40	4.05	8.58142E-05	4.05	0.054	0.78	0.25	2.97	31	92.05	68.35
71	5.20	16.15	74	13.10	24.90	1.90	1.27	0.09	4.30	12.00	15.10	3.79	7.60896E-05	3.79	0.054	0.63	0.24	2.41	30	72.37	53.74
72	6.30	15.90	64	12.20	25.00	1.86	1.05	0.09	6.50	11.95	14.80	4.52	0.000114551	4.52	0.054	0.81	0.26	3.46	31	107.25	79.64
1983	7.50	16.75	56	12.30	29.80	2.11	0.98	0.10	7.70	11.95	15.00	5.02	0.000135588	5.02	0.054	1.13	0.29	5.04	31	156.20	115.99
74	8.10	17.55	57	12.70	31.70	2.22	1.03	0.11	7.20	12.00	15.50	4.99	0.00012504	4.99	0.054	1.19	0.31	5.40	28	151.16	112.25
75	8.30	19.00	50	12.20	35.70	2.40	0.91	0.11	7.00	12.05	15.70	4.97	0.000127126	4.97	0.054	1.49	0.32	6.62	31	205.33	152.48
76	6.10	18.10	66	13.80	31.70	2.28	1.27	0.11	5.70	12.15	15.30	4.34	9.4091E-05	4.34	0.054	1.01	0.28	3.91	30	117.27	87.08
77	5.60	17.15	70	13.00	28.30	2.07	1.25	0.10	7.40	12.20	14.40	4.66	0.000117251	4.66	0.054	0.82	0.26	3.17	31	98.38	73.05
78	5.90	16.75	69	12.30	28.30	2.03	1.18	0.10	5.90	12.25	13.90	3.99	9.8983E-05	3.99	0.054	0.85	0.26	3.37	30	101.04	75.03
79	5.40	15.55	73	11.90	26.40	1.92	1.20	0.09	5.50	12.20	14.10	3.92	9.3375E-05	3.92	0.054	0.72	0.25	2.79	31	86.42	64.17
80	5.20	15.70	73	12.30	26.40	1.94	1.23	0.09	5.00	12.20	14.80	3.94	8.57555E-05	3.94	0.054	0.71	0.24	2.69	31	83.49	62.00
81	6.20	15.95	69	11.70	27.10	1.94	1.13	0.10	6.80	12.10	15.30	4.75	0.000114563	4.75	0.054	0.81	0.26	3.40	30	102.14	75.84
82	5.70	16.30	70	12.70	26.50	1.96	1.20	0.10	5.50	12.05	15.40	4.31	9.41661E-05	4.31	0.054	0.76	0.25	3.02	31	93.61	69.51
83	5.90	15.85	65	11.50	26.20	1.89	1.04	0.09	6.80	12.00	15.10	4.71	0.00011896	4.71	0.054	0.85	0.25	3.47	30	104.17	77.36
84	6.50	15.65	60	12.10	24.90	1.85	0.98	0.09	5.00	11.95	14.80	3.98	9.49687E-05	3.98	0.054	0.87	0.26	3.76	31	116.59	86.58
1984	7.50	16.10	47	10.70	29.80	2.03	0.74	0.10	7.90	11.95	15.00	5.10	0.000150826	5.10	0.054	1.28	0.29	5.77	31	178.99	132.92
86	8.60	16.85	40	10.70	33.60	2.22	0.65	0.11	9.50	12.00	15.50	5.86	0.000182459	5.86	0.054	1.57	0.32	7.32	28	204.88	152.14
87	8.10	17.50	45	11.40	33.60	2.25	0.77	0.11	7.40	12.05	15.70	5.12	0.000140298	5.12	0.054	1.48	0.31	6.68	31	206.99	153.70
88	6.40	17.45	62	13.10	29.80	2.15	1.13	0.10	6.60	12.15	15.30	4.67	0.000111343	4.67	0.054	1.02	0.28	4.15	30	124.48	92.43
89	6.30	16.85	56	11.50	29.90	2.07	0.93	0.10	8.00	12.20	14.40	4.87	0.000139828	4.87	0.054	1.14	0.27	4.65	31	144.17	107.05
90	6.20	15.70	63	10.70	28.30	1.95	0.98	0.10	7.40	12.25	13.90	4.49	0.000128217	4.49	0.054	0.97	0.26	4.02	30	120.63	89.58
91	5.40	15.45	72	11.20	26.70	1.90	1.14	0.09	5.70	12.20	14.10	3.99	9.81458E-05	3.99	0.054	0.76	0.25	2.95	31	91.52	67.96
92	5.40	15.85	73	11.50	28.10	1.98	1.19	0.10	6.40	12.20	14.80	4.44	0.000105857	4.44	0.054	0.79	0.25	3.04	31	94.15	69.91
93	6.40	16.00	55	10.50	28.90	1.97	0.85	0.10	7.30	12.10	15.30	4.94	0.000134187	4.94	0.054	1.12	0.27	4.70	30	141.03	104.73
94	6.00	15.55	59	11.50	26.40	1.90	0.95	0.09	4.70	12.05	15.40	4.01	9.08967E-05	4.01	0.054	0.95	0.26	3.88	31	120.29	89.32
95	5.80	15.60	64	12.10	24.80	1.85	1.04	0.09	5.10	12.00	15.10	4.09	9.4031E-05	4.09	0.054	0.80	0.25	3.28	30	98.29	72.99
96	6.60	15.95	59	12.80	26.40	1.98	1.02	0.10	6.80	11.95	14.80	4.63	0.000120285	4.63	0.054	0.94	0.27	4.04	31	125.18	92.96
1985	7.20	16.90	52	11.50	31.70	2.16	0.88	0.11	7.90	11.95	15.00	5.10	0.000143465	5.10	0.054	1.28	0.29	5.53	31	171.44	127.31

Contd. Appendix C.																					
98	7.30	17.40	45	12.30	30.00	2.12	0.79	0.10	6.40	12.00	15.50	4.69	0.000123803	4.69	0.054	1.33	0.29	5.79	28	162.01	120.31
99	7.40	17.20	52	12.10	31.70	2.19	0.91	0.11	6.20	12.05	15.70	4.66	0.000114869	4.66	0.054	1.28	0.30	5.55	31	172.14	127.83
100	5.30	16.50	74	13.10	26.40	1.98	1.30	0.10	5.20	12.15	15.30	4.15	8.66266E-05	4.15	0.054	0.68	0.25	2.60	30	78.06	57.96
101	5.10	16.10	74	12.60	26.20	1.94	1.26	0.09	6.30	12.20	14.40	4.28	0.000102098	4.28	0.054	0.68	0.24	2.57	31	79.68	59.17
102	5.20	15.35	73	11.10	26.40	1.88	1.14	0.09	7.50	12.25	13.90	4.52	0.00012256	4.52	0.054	0.73	0.24	2.83	30	84.89	63.04
103	5.40	14.50	72	11.10	24.90	1.80	1.11	0.09	6.30	12.20	14.10	4.19	0.000107623	4.19	0.054	0.69	0.24	2.77	31	85.76	63.68
104	5.40	15.30	72	10.60	23.40	1.70	1.05	0.08	6.80	12.20	14.80	4.58	0.000116859	4.58	0.054	0.65	0.24	2.65	31	82.18	61.02
105	5.80	15.80	64	10.80	28.10	1.95	1.00	0.10	7.30	12.10	15.30	4.94	0.00012717	4.94	0.054	0.95	0.26	3.80	30	113.99	84.65
106	5.90	16.20	56	11.50	29.40	2.05	0.93	0.10	6.10	12.05	15.40	4.53	0.00011274	4.53	0.054	1.12	0.26	4.41	31	136.78	101.57
107	6.00	16.05	59	12.10	26.20	1.92	0.98	0.09	4.90	12.00	15.10	4.01	9.31758E-05	4.01	0.054	0.94	0.26	3.82	30	114.64	85.13
108	6.80	16.85	61	12.80	28.10	2.05	1.07	0.10	7.00	11.95	14.80	4.70	0.000120881	4.70	0.054	0.97	0.28	4.17	31	129.29	96.01
1986	7.40	17.10	56	11.50	31.70	2.16	0.95	0.10	7.70	11.95	15.00	5.02	0.000137028	5.02	0.054	1.21	0.29	5.32	31	164.98	122.51
110	8.10	17.55	45	11.50	33.60	2.26	0.77	0.11	7.90	12.00	15.50	5.25	0.000148578	5.25	0.054	1.48	0.31	6.68	28	186.99	138.85
111	7.70	17.35	49	12.10	31.70	2.19	0.86	0.11	6.50	12.05	15.70	4.77	0.000121704	4.77	0.054	1.33	0.30	5.90	31	182.80	135.74
112	5.40	17.15	80	12.70	28.10	2.04	1.40	0.10	4.70	12.15	15.30	3.97	7.7244E-05	3.97	0.054	0.64	0.25	2.46	30	73.70	54.73
113	5.50	16.30	69	12.40	26.40	1.94	1.16	0.09	7.20	12.20	14.40	4.59	0.000117866	4.59	0.054	0.78	0.25	3.05	31	94.40	70.10
114	4.50	15.35	78	12.30	23.40	1.79	1.26	0.09	4.70	12.25	13.90	3.59	8.05456E-05	3.59	0.054	0.53	0.22	1.91	30	57.27	42.53
115	5.00	14.60	78	11.10	24.90	1.80	1.20	0.09	6.40	12.20	14.10	4.23	0.000105629	4.23	0.054	0.60	0.23	2.32	31	71.87	53.37
116	5.50	14.75	72	10.60	26.40	1.85	1.09	0.09	7.00	12.20	14.80	4.65	0.000118133	4.65	0.054	0.76	0.25	3.04	31	94.22	69.96
117	5.40	15.30	73	10.80	26.40	1.86	1.12	0.09	6.40	12.10	15.30	4.61	0.000109264	4.61	0.054	0.74	0.24	2.93	30	87.79	65.19
118	5.90	16.30	61	11.50	29.50	2.05	1.01	0.10	6.40	12.05	15.40	4.65	0.000113880	4.65	0.054	1.04	0.28	4.12	31	127.01	94.70
119	5.80	16.85	64	12.10	28.10	1.91	1.06	0.09	4.30	12.00	15.10	3.79	8.18448E-05	3.79	0.054	0.85	0.25	3.41	30	102.40	76.04
120	6.20	16.45	65	13.00	26.40	1.97	1.13	0.10	5.70	11.95	14.80	4.23	9.99356E-05	4.23	0.054	0.84	0.26	3.47	31	107.48	79.82
1987	6.60	17.35	65	12.30	31.70	2.20	1.15	0.11	6.60	11.95	15.00	4.62	0.000111975	4.62	0.054	1.05	0.28	4.31	31	133.52	99.15
122	7.20	18.10	52	12.30	33.60	2.30	0.94	0.11	7.80	12.00	15.50	5.22	0.000138479	5.22	0.054	1.36	0.30	5.74	28	160.60	119.25
123	7.10	18.75	54	12.20	35.70	2.40	0.98	0.12	7.30	12.05	15.70	5.08	0.000128359	5.08	0.054	1.41	0.30	5.82	31	180.46	134.01
124	6.00	17.00	66	12.30	29.80	2.11	1.15	0.10	6.40	12.15	15.30	4.60	0.000107765	4.60	0.054	0.96	0.27	3.80	30	113.90	84.58
125	5.00	17.15	74	13.10	28.10	2.06	1.32	0.10	5.60	12.20	14.40	4.04	9.079E-05	4.04	0.054	0.74	0.25	2.68	31	82.98	61.62
126	4.30	16.15	78	12.60	26.40	1.95	1.33	0.10	5.30	12.25	13.90	3.79	8.635E-05	3.79	0.054	0.62	0.23	2.11	30	63.21	46.94
127	5.30	15.65	73	11.50	26.40	1.90	1.17	0.09	8.60	12.20	14.10	4.97	0.000136908	4.97	0.054	0.73	0.24	2.84	31	87.92	65.29
128	5.30	16.30	68	12.00	28.10	2.01	1.14	0.10	6.60	12.20	14.80	4.51	0.000110383	4.51	0.054	0.86	0.25	3.25	31	100.90	74.93
129	6.10	17.20	56	11.70	31.70	2.17	0.96	0.11	7.00	12.10	15.30	4.83	0.000124557	4.83	0.054	1.21	0.27	4.77	30	143.16	106.31
130	6.30	17.75	53	12.00	33.60	2.28	0.94	0.11	7.40	12.05	15.40	5.02	0.000131908	5.02	0.054	1.34	0.28	5.28	31	163.56	121.45
131	5.20	16.85	65	13.00	28.10	2.06	1.16	0.10	3.80	12.00	15.10	3.61	7.21178E-05	3.61	0.054	0.90	0.25	3.29	30	98.68	73.28
132	6.60	16.70	57	11.50	29.80	2.07	0.95	0.10	8.20	11.95	14.80	5.14	0.000144575	5.14	0.054	1.12	0.28	4.70	31	145.74	108.23
1988	6.70	17.15	61	12.20	29.80	2.10	1.06	0.10	6.80	11.95	15.00	4.69	0.000118669	4.69	0.054	1.04	0.28	4.39	31	136.18	101.12
134	7.10	17.85	53	12.20	33.60	2.29	0.95	0.11	7.80	12.00	15.50	5.22	0.000137877	5.22	0.054	1.34	0.30	5.63	28	157.59	117.02
135	6.80	18.20	57	12.30	31.70	2.20	1.01	0.11	5.70	12.05	15.70	4.47	0.000103558	4.47	0.054	1.19	0.28	4.93	31	152.91	113.55
136	4.80	17.55	70	14.00	28.10	2.11	1.31	0.10	4.30	12.15	15.30	3.82	7.44087E-05	3.82	0.054	0.80	0.24	2.78	30	83.38	61.92
137	4.90	16.35	75	12.50	26.40	1.95	1.27	0.10	6.50	12.20	14.40	4.35	0.000104292	4.35	0.054	0.67	0.24	2.48	31	76.98	57.17
138	5.40	15.25	73	11.50	26.40	1.90	1.17	0.09	7.90	12.25	13.90	4.66	0.000126838	4.66	0.054	0.73	0.25	2.86	30	85.65	63.60
139	4.30	15.20	83	11.60	24.90	1.83	1.31	0.09	4.60	12.20	14.10	3.62	7.79677E-05	3.62	0.054	0.51	0.22	1.80	31	55.65	41.32
140	4.00	15.00	78	11.60	24.90	1.83	1.23	0.09	5.40	12.20	14.80	4.08	9.08587E-05	4.08	0.054	0.59	0.22	1.98	31	61.41	45.60
141	4.80	15.80	74	12.00	26.40	1.92	1.22	0.09	4.60	12.10	15.30	3.94	8.0983E-05	3.94	0.054	0.70	0.24	2.54	30	76.17	56.56
142	5.70	16.30	61	11.90	28.10	2.00	1.02	0.10	6.50	12.05	15.40	4.68	0.000114939	4.68	0.054	0.98	0.26	3.85	31	119.20	88.51
143	5.40	15.45	64	12.00	24.80	1.84	1.04	0.09	5.00	12.00	15.10	4.05	9.27564E-05	4.05	0.054	0.80	0.24	3.15	30	94.59	70.24
144	6.20	15.50	64	11.80	25.00	1.84	1.03	0.09	5.90	11.95	14.80	4.30	0.000106635	4.30	0.054	0.81	0.26	3.45	31	106.88	79.37
1989	6.50	16.15	59	12.00	28.10	2.01	0.99	0.10	6.60	11.95	15.00	4.62	0.000118353	4.62	0.054	1.01	0.27	4.26	31	132.02	98.04
146	6.90	16.00	55	11.50	28.10	1.98	0.90	0.10	7.00	12.00	15.50	4.91	0.0001281	4.91	0.054	1.08	0.28	4.72	28	132.02	98.03

Contd. Appendix C.

147	6.50	17.15	56	12.00	31.80	2.19	0.98	0.11	7.30	12.05	15.70	5.08	0.000128641	5.08	0.054	1.21	0.28	4.95	31	153.30	113.84
148	4.90	16.25	69	13.10	26.40	1.98	1.21	0.10	5.20	12.15	15.30	4.15	8.92538E-05	4.15	0.054	0.77	0.24	2.79	30	83.60	62.08
149	4.90	16.00	74	12.30	28.00	2.02	1.26	0.10	7.10	12.20	14.40	4.56	0.000112553	4.56	0.054	0.75	0.24	2.73	31	84.61	62.83
150	5.30	15.25	73	11.20	26.40	1.88	1.15	0.09	0.00	12.25	13.90	2.03	1.87792E-05	2.03	0.054	0.73	0.24	2.76	30	82.69	61.40
151	4.50	15.00	78	11.60	24.90	1.83	1.23	0.09	5.10	12.20	14.10	3.79	8.68239E-05	3.79	0.054	0.59	0.23	2.12	31	65.59	48.70
152	4.80	16.20	73	12.30	28.10	2.02	1.25	0.10	5.80	12.20	14.80	4.22	9.57612E-05	4.22	0.054	0.77	0.24	2.75	31	85.25	63.30
153	5.40	16.95	69	13.10	27.40	2.03	1.22	0.10	6.20	12.10	15.30	4.53	0.000102708	4.53	0.054	0.80	0.25	3.07	30	91.96	68.29
154	5.00	16.40	69	13.00	26.40	1.97	1.20	0.10	5.70	12.05	15.40	4.38	9.69164E-05	4.38	0.054	0.77	0.24	2.84	31	87.89	65.26
155	4.70	15.15	89	12.80	24.90	1.89	1.50	0.09	4.90	12.00	15.10	4.01	7.77514E-05	4.01	0.054	0.38	0.23	1.44	30	43.30	32.15
156	5.30	16.25	74	13.00	26.40	1.97	1.29	0.10	5.20	11.95	14.80	4.05	8.7976E-05	4.05	0.054	0.68	0.25	2.61	31	80.79	59.99
1990	5.50	15.85	55	11.50	28.10	1.98	0.90	0.10	7.40	11.95	15.00	4.91	0.000134725	4.91	0.054	1.08	0.25	4.17	31	129.13	95.89
158	5.00	17.60	61	13.10	29.80	2.15	1.11	0.10	5.00	12.00	15.50	4.16	9.03747E-05	4.16	0.054	1.03	0.25	3.64	28	101.96	75.71
159	4.60	17.20	70	14.00	28.10	2.11	1.31	0.10	4.50	12.05	15.70	4.01	7.75345E-05	4.01	0.054	0.80	0.24	2.72	31	84.22	62.54
160	4.60	17.05	75	13.40	28.10	2.08	1.36	0.10	4.80	12.15	15.30	4.01	7.95537E-05	4.01	0.054	0.71	0.24	2.46	30	73.91	54.88
161	4.70	16.80	75	12.50	28.10	2.03	1.30	0.10	7.20	12.20	14.40	4.59	0.000112814	4.59	0.054	0.73	0.24	2.60	31	80.53	59.80
162	4.70	15.70	73	11.50	26.50	1.90	1.17	0.09	8.30	12.25	13.90	4.79	0.000132256	4.79	0.054	0.73	0.23	2.65	30	79.51	59.04
163	4.40	15.85	78	12.00	26.40	1.92	1.29	0.09	6.10	12.20	14.10	4.13	9.84938E-05	4.13	0.054	0.63	0.23	2.20	31	68.33	50.74
164	4.60	16.70	73	13.10	26.40	1.98	1.28	0.10	5.60	12.20	14.80	4.15	9.21613E-05	4.15	0.054	0.70	0.23	2.45	31	76.04	56.46
165	5.70	16.55	61	11.20	29.40	2.03	0.99	0.10	7.90	12.10	15.30	5.16	0.000136432	5.16	0.054	1.04	0.26	4.07	30	122.10	90.67
166	5.50	16.45	56	10.80	28.50	1.97	0.88	0.10	5.10	12.05	15.40	4.16	9.92901E-05	4.16	0.054	1.09	0.25	4.16	31	129.10	95.87
167	5.30	16.15	65	12.00	26.40	1.92	1.07	0.09	5.50	12.00	15.10	4.23	9.87883E-05	4.23	0.054	0.85	0.24	3.24	30	97.17	72.15
168	6.00	16.45	55	12.10	26.80	1.95	0.92	0.09	6.20	11.95	14.80	4.41	0.000115404	4.41	0.054	1.03	0.26	4.17	31	129.14	95.90
1991	6.60	17.35	47	12.00	31.70	2.19	0.82	0.11	7.90	11.95	15.00	5.10	0.000146565	5.10	0.054	1.37	0.28	5.59	31	173.27	128.66
170	6.80	18.10	45	12.30	33.60	2.30	0.81	0.11	8.30	12.00	15.50	5.40	0.000152807	5.40	0.054	1.48	0.29	6.06	28	169.80	126.09
171	6.20	18.25	49	13.10	33.60	2.34	0.92	0.11	6.50	12.05	15.70	4.77	0.000118894	4.77	0.054	1.41	0.28	5.43	31	168.24	124.93
172	5.40	17.15	65	13.10	28.40	2.08	1.17	0.10	5.90	12.15	15.30	4.41	0.000100253	4.41	0.054	0.91	0.25	3.42	30	102.61	76.19
173	4.60	17.05	70	13.40	26.80	2.01	1.25	0.10	5.70	12.20	14.40	4.08	9.43711E-05	4.08	0.054	0.76	0.24	2.64	31	81.83	60.77
174	5.10	16.95	79	13.10	28.10	2.06	1.41	0.10	5.70	12.25	13.90	3.93	8.90483E-05	3.93	0.054	0.65	0.25	2.40	30	71.87	53.37
175	4.00	15.05	78	12.30	23.40	1.79	1.26	0.09	3.80	12.20	14.10	3.35	6.87925E-05	3.35	0.054	0.53	0.22	1.77	31	54.88	40.75
176	5.00	15.65	73	11.70	26.40	1.91	1.18	0.09	5.40	12.20	14.80	4.08	9.24537E-05	4.08	0.054	0.72	0.24	2.69	31	83.43	61.95
177	5.70	16.30	65	11.50	28.10	1.98	1.06	0.10	6.70	12.10	15.30	4.72	0.000115803	4.72	0.054	0.92	0.26	3.63	30	108.91	80.87
178	5.40	16.50	56	10.80	28.40	1.96	0.88	0.10	4.60	12.05	15.40	3.97	9.16077E-05	3.97	0.054	1.08	0.25	4.10	31	127.15	94.42
179	5.10	16.00	69	12.00	26.40	1.92	1.14	0.09	4.70	12.00	15.10	3.94	8.52574E-05	3.94	0.054	0.78	0.24	2.92	30	87.68	65.11
180	6.00	16.45	56	12.30	26.70	1.95	0.94	0.10	5.30	11.95	14.80	4.09	0.000100671	4.09	0.054	1.01	0.26	4.07	31	126.10	93.64
1992	6.40	17.15	49	12.00	31.70	2.19	0.85	0.11	6.90	11.95	15.00	4.73	0.000129049	4.73	0.054	1.33	0.28	5.35	31	165.72	123.06
182	6.80	18.30	46	12.40	33.60	2.30	0.83	0.11	7.30	12.00	15.50	5.03	0.000135811	5.03	0.054	1.47	0.29	5.97	28	167.22	124.17
183	6.80	18.70	42	13.10	35.70	2.44	0.81	0.12	6.80	12.05	15.70	4.89	0.000128816	4.89	0.054	1.63	0.30	6.49	31	201.22	149.42
184	5.20	18.00	66	13.40	31.80	2.26	1.24	0.11	4.80	12.15	15.30	4.01	8.27737E-05	4.01	0.054	1.02	0.26	3.59	30	107.67	79.95
185	5.00	16.70	65	12.40	28.10	2.03	1.12	0.10	7.20	12.20	14.40	4.59	0.000119741	4.59	0.054	0.91	0.24	3.29	31	102.11	75.82
186	4.50	16.75	74	12.60	28.10	2.04	1.29	0.10	5.10	12.25	13.90	3.73	8.49938E-05	3.73	0.054	0.75	0.24	2.54	30	76.29	56.65
187	4.60	15.35	72	11.70	24.90	1.83	1.15	0.09	5.30	12.20	14.10	3.86	9.22629E-05	3.86	0.054	0.68	0.23	2.46	31	76.25	56.62
188	4.00	15.15	78	12.20	24.90	1.86	1.28	0.09	4.80	12.20	14.80	3.87	8.15875E-05	3.87	0.054	0.58	0.22	1.92	31	59.62	44.27
189	4.90	15.70	64	11.50	26.40	1.90	1.03	0.09	5.70	12.10	15.30	4.35	0.000102668	4.35	0.054	0.87	0.24	3.19	30	95.70	71.06
190	5.10	16.60	65	12.30	26.40	1.94	1.09	0.09	4.60	12.05	15.40	3.97	8.49615E-05	3.97	0.054	0.84	0.24	3.14	31	97.25	72.22
191	5.40	15.65	59	11.40	26.40	1.89	0.94	0.09	5.30	12.00	15.10	4.16	0.000100464	4.16	0.054	0.95	0.25	3.67	30	110.06	81.73
192	5.30	16.30	64	12.90	26.40	1.97	1.11	0.10	5.40	11.95	14.80	4.12	9.84372E-05	4.12	0.054	0.86	0.25	3.24	31	100.45	74.59

Appendix C.2. Monthly evapotranspiration (mm/month) estimated with Hargreaves and Pan A methods

Month	Temp(oC)	Radiation()	Days	TD(oC)	Hargreaves ETo mm/day	Monthly ETm(Hargreave)	Corr. Hargreaves (ETm)	Monthly Pan A (ETm)	Corr.Pan A (ETm)
1977	20.35	15.00	31.00	5.50	2.95	91.53	63.56	156.00	99.15
2	21.30	15.50	28.00	4.60	2.86	80.07	55.60	184.60	117.33
3	21.20	15.70	31.00	5.60	3.19	98.82	68.62	170.50	108.37
4	17.70	15.30	30.00	11.00	3.96	118.89	82.57	156.60	99.53
5	17.85	14.40	31.00	6.90	2.97	91.97	63.87	151.20	96.10
6	18.40	13.90	30.00	4.60	2.37	71.23	49.46	219.00	139.19
7	17.45	14.10	31.00	5.30	2.52	78.04	54.19	136.40	86.69
8	18.25	14.80	31.00	5.70	2.80	86.87	60.33	198.40	126.10
9	16.75	15.30	30.00	11.10	3.87	116.24	80.72	159.00	101.06
10	19.85	15.40	31.00	7.10	3.40	105.37	73.17	195.30	124.13
11	17.90	15.10	30.00	3.60	2.25	67.51	46.88	150.00	95.34
12	18.85	14.80	31.00	3.50	2.23	69.21	48.06	220.10	139.89
1978	18.95	15.00	31.00	5.90	2.95	91.32	63.42	198.40	126.10
14	20.35	15.50	28.00	6.50	3.32	92.87	64.49	131.60	83.64
15	19.90	15.70	31.00	5.80	3.14	97.22	67.51	120.90	76.84
16	18.80	15.30	30.00	5.40	2.86	85.88	59.64	116.00	73.73
17	15.85	14.40	31.00	11.90	3.68	114.00	79.17	156.60	99.53
18	15.60	13.90	30.00	12.20	3.57	107.02	74.32	144.00	91.53
19	14.05	14.10	31.00	11.50	3.35	103.86	72.13	142.60	90.64
20	14.90	14.80	31.00	12.40	3.75	116.23	80.71	204.60	130.04
21	15.25	15.30	30.00	13.30	4.06	121.71	84.52	168.00	106.78
22	14.90	15.40	31.00	13.20	4.03	124.78	86.65	151.90	96.55
23	14.85	15.10	30.00	12.70	3.87	115.96	80.53	162.00	102.97
24	17.00	14.80	31.00	6.40	2.87	88.86	61.71	183.00	116.31
1979	15.25	15.00	31.00	13.50	4.01	124.23	86.27	176.70	112.31
26	15.65	15.50	28.00	13.50	4.19	117.35	81.49	134.40	85.42
27	16.00	15.70	31.00	14.80	4.49	139.23	96.69	279.00	177.33
28	15.80	15.30	30.00	13.40	4.14	124.20	86.25	156.00	99.15
29	15.30	14.40	31.00	13.40	3.84	118.99	82.64	189.10	120.19
30	15.10	13.90	30.00	13.00	3.63	108.82	75.57	180.00	114.41
31	14.90	14.10	31.00	12.80	3.63	112.50	78.13	198.40	126.10
32	14.95	14.80	31.00	13.50	3.92	121.46	84.35	179.80	114.28
33	15.90	15.30	30.00	14.80	4.36	130.92	90.91	132.00	83.90
34	16.25	15.40	31.00	14.90	4.45	138.04	95.86	102.30	65.02
35	15.90	15.10	30.00	11.80	3.85	115.37	80.12	116.00	73.73
36	16.70	14.80	31.00	9.40	3.44	106.77	74.14	150.80	95.85
Jun-09	18.55	15.00	31.00	13.10	4.34	134.59	93.47	183.00	118.22
38	18.40	15.50	28.00	12.80	4.42	123.66	85.87	124.70	79.26
39	18.40	15.70	31.00	12.80	4.47	138.67	96.30	164.30	104.43

Contd. Appendix C.2									
40	19.15	15.30	30.00	9.90	3.91	117.40	81.53	147.00	93.43
41	18.90	14.40	31.00	7.40	3.16	98.05	68.09	158.10	100.49
42	18.40	13.90	30.00	7.60	3.05	91.55	63.58	165.00	104.87
43	17.15	14.10	31.00	8.30	3.12	96.83	67.24	161.20	102.46
44	19.30	14.80	31.00	6.00	2.96	91.73	63.70	124.00	78.81
45	20.20	15.30	30.00	8.60	3.75	112.53	78.15	216.00	137.29
46	20.90	15.40	31.00	7.60	3.61	112.05	77.81	161.20	102.46
47	16.25	15.10	30.00	10.30	3.63	108.91	75.63	113.10	71.89
48	17.15	14.80	31.00	13.90	4.24	131.52	91.34	151.90	96.55
1981	17.20	15.00	31.00	16.80	4.73	146.76	101.91	198.40	126.10
50	17.55	15.50	28.00	16.90	4.96	138.75	96.36	190.40	121.02
51	17.55	15.70	31.00	13.90	4.55	141.12	98.00	114.70	72.90
52	17.45	15.30	30.00	10.90	3.92	117.52	81.61	132.00	83.90
53	16.00	14.40	31.00	12.00	3.71	114.99	79.85	161.20	102.46
54	15.60	13.90	30.00	14.20	3.85	115.46	80.18	162.00	102.97
55	14.55	14.10	31.00	12.30	3.52	109.10	75.76	99.20	63.05
56	14.65	14.80	31.00	11.30	3.55	110.10	76.46	62.00	39.41
57	16.00	15.30	30.00	13.50	4.18	125.41	87.09	198.00	125.85
58	17.10	15.40	31.00	11.90	4.08	126.45	87.81	298.40	189.66
59	17.25	15.10	30.00	10.30	3.74	112.11	77.85	123.00	78.18
60	16.85	14.80	31.00	11.40	3.81	118.09	82.00	133.30	84.72
1982	17.00	15.00	31.00	16.40	4.65	144.17	100.12	151.90	96.55
62	17.75	15.50	28.00	16.50	4.92	137.88	95.75	103.60	65.85
63	18.10	15.70	31.00	17.60	5.20	161.26	111.99	176.70	112.31
64	17.30	15.30	30.00	11.60	4.02	120.72	83.83	108.00	68.64
65	16.40	14.40	31.00	10.20	3.46	107.27	74.49	108.50	68.96
66	15.90	13.90	30.00	13.00	3.72	111.47	77.41	105.00	66.74
67	15.10	14.10	31.00	13.20	3.71	114.94	79.82	102.30	65.02
68	15.45	14.80	31.00	11.30	3.64	112.82	78.35	102.30	65.02
69	15.85	15.30	30.00	13.50	4.16	124.85	86.70	108.00	68.64
70	15.95	15.40	31.00	11.90	3.94	122.28	84.92	93.00	59.11
71	16.15	15.10	30.00	10.30	3.62	108.59	75.41	87.00	55.30
72	15.90	14.80	31.00	11.40	3.70	114.85	79.76	111.60	70.93
1983	16.75	15.00	31.00	13.90	4.25	131.77	91.51	124.00	78.81
74	17.55	15.50	28.00	13.90	4.49	125.84	87.39	114.80	72.97
75	19.00	15.70	31.00	16.40	5.15	159.57	110.81	158.10	100.49
76	18.10	15.30	30.00	13.40	4.42	132.70	92.16	126.00	80.08
77	17.15	14.40	31.00	12.50	3.91	121.35	84.27	108.50	68.96
78	16.75	13.90	30.00	12.90	3.79	113.84	79.06	99.00	62.92
79	15.55	14.10	31.00	12.10	3.60	111.56	77.47	99.20	63.05
80	15.70	14.80	31.00	11.60	3.71	115.16	79.98	102.30	65.02

C.2 contd									
81	15.95	15.30	30.00	13.10	4.11	123.35	85.66	117.00	74.36
82	16.30	15.40	31.00	11.80	3.97	123.03	85.44	108.50	68.96
83	15.85	15.10	30.00	12.70	3.98	119.51	82.99	108.00	68.64
84	15.65	14.80	31.00	11.50	3.69	114.50	79.51	102.30	65.02
1984	16.10	15.00	31.00	15.20	4.36	135.21	93.89	130.20	82.75
86	16.85	15.50	28.00	17.30	4.91	137.61	95.56	159.50	101.38
87	17.50	15.70	31.00	17.00	5.03	155.84	108.22	164.30	104.43
88	17.45	15.30	30.00	13.50	4.36	130.79	90.82	132.00	83.90
89	16.85	14.40	31.00	15.10	4.27	132.23	91.83	133.30	84.72
90	15.70	13.90	30.00	15.00	3.97	119.03	82.66	117.00	74.36
91	15.45	14.10	31.00	13.70	3.82	118.35	82.19	102.30	65.02
92	15.85	14.80	31.00	13.50	4.03	124.80	86.66	111.60	70.93
93	16.00	15.30	30.00	15.20	4.44	133.07	92.41	132.00	83.90
94	15.55	15.40	31.00	12.70	4.03	124.83	86.68	99.20	63.05
95	15.60	15.10	30.00	11.20	3.71	111.40	77.36	90.00	57.20
96	15.95	14.80	31.00	12.70	3.92	121.40	84.31	111.60	70.93
1985	16.90	15.00	31.00	16.00	4.58	141.99	98.61	142.60	90.64
98	17.40	15.50	28.00	14.00	4.49	125.75	87.33	134.40	85.42
99	17.20	15.70	31.00	14.80	4.65	144.17	100.12	148.80	94.58
100	16.50	15.30	30.00	10.80	3.79	113.83	79.05	120.00	76.27
101	16.10	14.40	31.00	10.80	3.53	109.41	75.98	108.50	68.96
102	15.35	13.90	30.00	12.30	3.56	106.66	74.07	120.00	76.27
103	14.50	14.10	31.00	12.00	3.47	107.60	74.72	99.20	63.05
104	15.30	14.80	31.00	10.00	3.41	105.65	73.37	111.60	70.93
105	15.80	15.30	30.00	14.20	4.26	127.86	88.79	93.00	59.11
106	16.20	15.40	31.00	14.00	4.31	133.61	92.79	111.00	70.55
107	16.05	15.10	30.00	12.30	3.94	118.31	82.16	99.00	62.92
108	16.85	14.80	31.00	12.70	4.02	124.64	86.55	120.90	76.84
1986	17.10	15.00	31.00	15.60	4.55	141.01	97.93	145.70	92.61
110	17.55	15.50	28.00	16.90	4.96	138.75	96.36	148.40	94.32
111	17.35	15.70	31.00	15.10	4.72	146.25	101.56	142.60	90.64
112	17.15	15.30	30.00	11.30	3.95	118.64	82.39	105.00	66.74
113	16.30	14.40	31.00	11.60	3.68	114.06	79.21	111.60	70.93
114	15.35	13.90	30.00	10.90	3.35	100.41	69.73	87.00	55.30
115	14.60	14.10	31.00	12.20	3.51	108.82	75.57	105.40	66.99
116	14.75	14.80	31.00	13.70	3.92	121.61	84.45	111.60	70.93
117	15.30	15.30	30.00	14.00	4.17	125.06	86.85	120.00	76.27
118	16.30	15.40	31.00	14.00	4.32	134.01	93.06	114.00	72.46
119	15.95	15.10	30.00	12.10	3.90	117.00	81.25	96.00	61.02
120	16.45	14.80	31.00	11.30	3.75	116.21	80.70	108.90	69.22
1987	17.35	15.00	31.00	14.50	4.42	136.93	95.09	133.30	84.72

Contd. Appendix C.2									
122	18.10	15.50	28.00	16.00	4.90	137.11	95.21	148.30	94.26
123	18.75	15.70	31.00	15.90	5.03	156.05	108.37	164.30	104.43
124	17.00	15.30	30.00	14.80	4.51	135.19	93.88	117.00	74.36
125	17.15	14.40	31.00	11.90	3.82	118.40	82.23	108.50	68.96
126	16.15	13.90	30.00	10.90	3.43	102.83	71.41	90.00	57.20
127	15.65	14.10	31.00	13.30	3.78	117.31	81.46	124.00	78.81
128	16.30	14.80	31.00	13.20	4.03	125.05	86.84	124.00	78.81
129	17.20	15.30	30.00	15.60	4.65	139.59	96.94	126.00	80.08
130	17.75	15.40	31.00	15.90	4.80	148.88	103.39	135.00	85.81
131	16.85	15.10	30.00	11.90	3.97	119.12	82.72	99.00	62.92
132	16.70	14.80	31.00	15.00	4.35	134.87	93.66	130.20	82.75
1988	17.15	15.00	31.00	13.70	4.27	132.34	91.90	130.20	82.75
134	17.85	15.50	28.00	15.90	4.85	135.73	94.26	142.10	90.32
135	18.20	15.70	31.00	14.00	4.65	144.23	100.16	139.50	88.67
136	17.55	15.30	30.00	11.50	4.04	121.05	84.06	108.00	68.64
137	16.35	14.40	31.00	11.50	3.67	113.73	78.98	105.40	66.99
138	15.25	13.90	30.00	12.50	3.57	107.20	74.44	111.00	70.55
139	15.20	14.10	31.00	11.80	3.52	109.01	75.70	89.90	57.14
140	15.00	14.80	31.00	11.40	3.61	111.78	77.63	99.20	63.05
141	15.80	15.30	30.00	12.20	3.95	118.51	82.30	93.00	59.11
142	16.30	15.40	31.00	13.20	4.20	130.12	90.36	114.00	72.46
143	15.45	15.10	30.00	11.30	3.71	111.39	77.36	96.00	61.02
144	15.50	14.80	31.00	12.00	3.76	116.43	80.86	105.40	66.99
1989	16.15	15.00	31.00	12.70	3.99	123.77	85.95	136.40	86.69
146	16.00	15.50	28.00	14.40	4.37	122.46	85.04	131.60	83.64
147	17.15	15.70	31.00	15.10	4.69	145.42	100.99	155.00	98.52
148	16.25	15.30	30.00	11.50	3.89	116.60	80.97	99.00	62.92
149	16.00	14.40	31.00	11.60	3.65	113.06	78.51	120.90	76.84
150	15.25	13.90	30.00	13.30	3.69	110.57	76.79	108.00	68.64
151	15.00	14.10	31.00	11.60	3.47	107.43	74.60	99.20	63.05
152	16.20	14.80	31.00	12.60	3.93	121.82	84.60	120.90	76.84
153	16.95	15.30	30.00	11.10	3.90	116.91	81.19	165.00	104.87
154	16.40	15.40	31.00	11.00	3.84	119.13	82.73	111.00	70.55
155	15.15	15.10	30.00	10.70	3.58	107.42	74.59	99.00	62.92
156	16.25	14.80	31.00	11.40	3.74	116.04	80.58	105.40	66.99
1990	15.85	15.00	31.00	14.10	4.17	129.26	89.77	130.20	82.75
158	17.60	15.50	28.00	12.60	4.28	119.98	83.32	142.80	90.76
159	17.20	15.70	31.00	11.00	4.01	124.29	86.31	120.90	76.84
160	17.05	15.30	30.00	11.30	3.94	118.30	82.15	117.00	74.36
161	16.80	14.40	31.00	12.40	3.86	119.66	83.09	120.90	76.84
162	15.70	13.90	30.00	13.40	3.75	112.50	78.13	111.00	70.55

Appendix C.3. Annual streamflow, mean rainfall, and Pan A estimated evapotranspiration in millimetres (mm

YEAR	SWSII(mm/yr)	SWSI (mm/yr)	Rainfall (mm/yr)	Pan ETo (mm/yr)	Corrected ETm(mm/yr)	
1960	65.856	51.633	829.750	713.668	453.550	
1961	27.495	492.567	1333.100	1146.599	728.686	
1962	90.528	194.161	1094.800	941.637	598.429	
1963	232.677	330.295	1118.650	962.151	611.465	
1964	36.069	530.562	1024.550	881.215	560.029	
1965	36.069	19.199	577.600	1155.075	978.877	
1966	60.116	54.217	994.900	1070.100	906.864	
1967	17.443	170.935	804.035	1110.000	940.678	
1968	142.799	466.556	1189.900	966.525	819.089	
1969	24.169	44.655	806.050	964.500	817.373	
1970	217.237	330.220	1135.200	936.750	793.856	
1971	79.382	405.503	942.600	946.200	801.864	
1972	99.392	163.256	821.200	1282.748	1087.074	
1973	33.150	31.486	783.150	1651.725	1399.767	
1974	165.909	78.608	1088.700	1072.125	908.581	
1975	120.034	77.793	1280.550	1022.325	866.377	
1976	77.579	69.682	767.750	1112.775	943.030	
1977	96.086	399.347	1253.400	1572.975	1333.030	
1978	271.370	256.779	1322.700	1409.700	1194.661	
1979	177.124	104.700	1006.450	1495.875	1267.691	
1980	41.687	52.299	856.750	1404.375	1190.148	
1981	54.932	103.085	934.750	1404.450	1190.212	
1982	97.032	191.714	1057.050	1018.425	863.072	
1983	36.365	120.013	955.950	1025.775	869.301	
1984	10.742	27.802	627.400	1112.250	942.585	
1985	51.485	61.281	1034.350	1056.750	895.551	
1986	38.435	30.956	915.300	1047.150	887.415	
1987	29.016	29.417	762.900	1124.700	953.136	
1988	7.884	15.390	1139.900	1000.275	847.691	
1989	215.233	125.185	1095.300	1088.550	922.500	
1990	43.461	209.096	1064.200	1080.975	916.081	

Contd. Appendix C.3

163	15.85	14.10	31.00	12.10	3.63	112.56	78.17	120.90	76.84
164	16.70	14.80	31.00	10.80	3.69	114.44	79.47	120.90	76.84
165	16.55	15.30	30.00	15.50	4.55	136.56	94.83	129.00	81.99
166	16.45	15.40	31.00	13.70	4.30	133.15	92.46	111.00	70.55
167	16.15	15.10	30.00	12.50	3.99	119.62	83.07	102.00	64.83
168	16.45	14.80	31.00	12.70	3.97	123.20	85.56	114.70	72.90
1991	17.35	15.00	31.00	15.30	4.54	140.65	97.68	148.80	94.58
170	18.10	15.50	28.00	16.40	4.96	138.81	96.40	151.20	96.10
171	18.25	15.70	31.00	15.30	4.87	150.99	104.85	151.20	96.10
172	17.15	15.30	30.00	12.30	4.13	123.78	85.96	108.00	68.64
173	17.05	14.40	31.00	10.90	3.65	113.00	78.47	105.40	66.99
174	16.95	13.90	30.00	11.90	3.67	109.97	76.37	105.00	66.74
175	15.05	14.10	31.00	10.10	3.24	100.39	69.72	86.80	55.17
176	15.65	14.80	31.00	12.30	3.82	118.41	82.23	111.60	70.93
177	16.30	15.30	30.00	13.60	4.23	126.99	88.19	123.00	78.18
178	16.50	15.40	31.00	13.40	4.25	131.87	91.58	111.00	70.55
179	16.00	15.10	30.00	12.20	3.92	117.66	81.71	99.00	62.92
180	16.45	14.80	31.00	12.70	3.97	123.20	85.56	117.80	74.87
1992	17.15	15.00	31.00	15.30	4.51	139.85	97.12	136.40	86.69
182	18.30	15.50	28.00	15.80	4.89	137.01	95.14	142.80	90.76
183	18.70	15.70	31.00	16.20	5.07	157.30	109.24	142.80	90.76
184	18.00	15.30	30.00	13.00	4.34	130.34	90.52	117.00	74.36
185	16.70	14.40	31.00	12.40	3.85	119.31	82.85	114.70	72.90
186	16.75	13.90	30.00	12.10	3.68	110.26	76.57	108.00	68.64
187	15.35	14.10	31.00	12.10	3.58	110.89	77.00	102.30	65.02
188	15.15	14.80	31.00	11.10	3.57	110.81	76.95	86.80	55.17
189	15.70	15.30	30.00	13.20	4.10	122.90	85.35	108.00	68.64
190	16.60	15.40	31.00	12.40	4.10	127.23	88.35	105.00	66.74
191	15.65	15.10	30.00	12.50	3.93	117.86	81.85	96.00	61.02
192	16.30	14.80	31.00	11.40	3.75	116.21	80.70	102.30	65.02

Appendix D.1 Time series of the mean monthly NDVI at selected sites in Enjoro river watershed

MONTH	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	Annual mean
1982ndvi													
2FC11RGS	0.380	0.395	0.455	0.410	0.377	0.435	0.540	0.433	0.345	0.375	0.445	0.360	0.420
2FC05RGS	0.380	0.395	0.455	0.408	0.380	0.435	0.540	0.437	0.345	0.375	0.445	0.360	0.420
EGERTON Met. Stn	0.375	0.330	0.270	0.333	0.473	0.500	0.465	0.410	0.340	0.425	0.375	0.325	0.360
NPBR Met.stn	0.235	0.260	0.275	0.143	0.163	0.385	0.550	0.443	0.315	0.290	0.375	0.340	0.290
2FC09RGS	0.235	0.260	0.275	0.143	0.163	0.385	0.550	0.443	0.315	0.290	0.375	0.340	0.290
2FC10RGS	0.205	0.245	0.175	0.125	0.067	0.405	0.495	0.353	0.275	0.335	0.300	0.275	0.220
1982 mean	0.343	0.345	0.364	0.323	0.348	0.439	0.524	0.431	0.336	0.366	0.410	0.346	
1983ndvi													
2FC11RGS	0.455	0.390	0.390	0.393	0.347	0.355	0.395	0.380	0.415	0.420	0.470	0.530	0.480
2FC05 RGS	0.455	0.390	0.390	0.363	0.320	0.355	0.395	0.380	0.415	0.420	0.470	0.520	0.480
EGERTON MET	0.375	0.405	0.415	0.253	0.317	0.385	0.350	0.450	0.365	0.330	0.310	0.430	0.430
NPBR	0.340	0.265	0.300	0.175	0.237	0.350	0.420	0.433	0.375	0.325	0.465	0.420	0.390
2FC09 RGS	0.340	0.265	0.300	0.175	0.240	0.350	0.400	0.433	0.450	0.360	0.465	0.420	0.390
2FC10 RGS	0.345	0.240	0.235	0.135	0.240	0.360	0.405	0.357	0.275	0.245	0.365	0.375	0.410
1983 mean	0.406	0.363	0.374	0.296	0.305	0.361	0.390	0.411	0.393	0.374	0.429	0.475	
1984ndvi													
2FC11RGS	0.485	0.290	0.385	0.370	0.453	0.265	0.245	0.233	0.315	0.360	0.455	0.475	0.450
2FC05 RGS	0.485	0.290	0.385	0.360	0.430	0.265	0.245	0.233	0.315	0.360	0.455	0.475	0.450
EGERTON MET	0.385	0.400	0.355	0.275	0.427	0.425	0.425	0.393	0.360	0.415	0.290	0.265	0.430
NPBR	0.325	0.250	0.270	0.158	0.220	0.180	0.195	0.090	0.150	0.325	0.185	0.200	0.310
2FC09 RGS	0.325	0.250	0.270	0.168	0.257	0.180	0.195	0.090	0.165	0.325	0.185	0.200	0.310
2FC10 RGS	0.310	0.280	0.275	0.198	0.227	0.155	0.170	0.133	0.255	0.275	0.225	0.165	0.340
1984 mean	0.420	0.308	0.349	0.291	0.383	0.284	0.278	0.238	0.285	0.365	0.346	0.354	
1985ndvi													
2FC11RGS	0.420	0.320	0.250	0.325	0.423	0.445	0.575	0.457	0.410	0.445	0.545	0.380	0.420
2FC05 RGS	0.420	0.270	0.250	0.323	0.423	0.445	0.575	0.457	0.410	0.445	0.545	0.380	0.420
EGERTON MET	0.480	0.205	0.280	0.340	0.513	0.545	0.530	0.340	0.460	0.500	0.480	0.385	0.480
NPBR	0.350	0.120	0.145	0.200	0.433	0.495	0.575	0.453	0.415	0.420	0.330	0.215	0.350
2FC09 RGS	0.350	0.120	0.145	0.200	0.433	0.495	0.575	0.453	0.415	0.420	0.330	0.215	0.350
2FC10 RGS	0.260	0.020	0.170	0.243	0.400	0.455	0.505	0.367	0.210	0.375	0.405	0.210	0.260
1985 mean	0.418	0.229	0.231	0.297	0.448	0.483	0.564	0.427	0.424	0.453	0.475	0.340	

Contd. Appendix D.1													
1986ndvi													
2FC11RGS	0.225	0.285	0.315	0.230	0.303	0.530	0.485	0.530	0.590	0.515	0.390	0.395	0.220
2FC05 RGS	0.225	0.285	0.315	0.230	0.303	0.530	0.485	0.530	0.590	0.515	0.390	0.395	0.220
EGERTON MET	0.375	0.310	0.355	0.378	0.517	0.545	0.465	0.440	0.420	0.285	0.455	0.330	0.390
NPBR	0.190	0.150	0.145	0.098	0.217	0.415	0.535	0.493	0.540	0.460	0.470	0.325	0.180
2FC09 RGS	0.190	0.150	0.145	0.098	0.217	0.415	0.535	0.493	0.540	0.460	0.470	0.325	0.180
2FC10 RGS	0.135	0.120	0.185	0.108	0.230	0.425	0.420	0.387	0.395	0.335	0.395	0.255	0.170
1986 mean	0.254	0.258	0.283	0.234	0.335	0.505	0.493	0.498	0.535	0.444	0.426	0.361	
1987ndvi													
2FC11RGS	0.320	0.225	0.380	0.368	0.297	0.500	0.520	0.527	0.445	0.415	0.305	0.240	0.370
2FC05 RGS	0.320	0.225	0.380	0.368	0.297	0.500	0.520	0.527	0.445	0.415	0.305	0.240	0.370
EGERTON MET	0.365	0.295	0.335	0.430	0.433	0.470	0.500	0.460	0.335	0.435	0.445	0.375	0.300
NPBR	0.255	0.150	0.190	0.153	0.273	0.405	0.490	0.517	0.380	0.295	0.240	0.155	0.250
2FC09 RGS	0.255	0.150	0.190	0.153	0.273	0.405	0.490	0.517	0.380	0.295	0.240	0.155	0.250
2FC10 RGS	0.165	0.170	0.215	0.178	0.317	0.420	0.495	0.450	0.295	0.220	0.125	0.145	0.170
1987 mean	0.315	0.224	0.321	0.329	0.325	0.469	0.508	0.508	0.401	0.390	0.324	0.253	
1988ndvi													
2FC11RGS	0.445	0.300	0.275	0.213	0.240	0.385	0.525	0.420	0.375	0.385	0.390	0.330	0.410
2FC05 RGS	0.445	0.300	0.275	0.213	0.240	0.385	0.525	0.420	0.375	0.385	0.380	0.315	0.410
EGERTON MET	0.400	0.440	0.430	0.495	0.580	0.580	0.495	0.310	0.305	0.345	0.330	0.210	0.390
NPBR	0.235	0.255	0.180	0.163	0.183	0.305	0.565	0.450	0.455	0.430	0.350	0.225	0.210
2FC09 RGS	0.235	0.255	0.180	0.163	0.183	0.305	0.565	0.450	0.455	0.430	0.350	0.225	0.210
2FC10 RGS	0.230	0.185	0.140	0.135	0.193	0.190	0.450	0.400	0.310	0.370	0.410	0.290	0.220
1988 mean	0.381	0.324	0.290	0.271	0.311	0.414	0.528	0.400	0.378	0.386	0.363	0.270	
1990ndvi													
2FC11RGS	0.610	0.570	0.320	0.443	0.530	0.465	0.565	0.440	0.485	0.495	0.570	0.545	0.610
2FC05 RGS	0.610	0.570	0.320	0.443	0.530	0.465	0.565	0.440	0.485	0.495	0.570	0.545	0.610
EGERTON MET	0.540	0.485	0.390	0.430	0.533	0.575	0.570	0.503	0.515	0.390	0.350	0.490	0.570
NPBR	0.440	0.440	0.315	0.343	0.483	0.490	0.475	0.467	0.410	0.405	0.390	0.270	0.460
2FC09 RGS	0.440	0.440	0.315	0.343	0.483	0.490	0.475	0.467	0.410	0.405	0.390	0.265	0.460
2FC10 RGS	0.435	0.310	0.225	0.350	0.453	0.465	0.520	0.403	0.295	0.305	0.330	0.095	0.460
1990 mean	0.550	0.516	0.336	0.414	0.519	0.499	0.544	0.463	0.474	0.446	0.470	0.463	

Appendix E.1 The objective function(F1), percentage error (%) and coefficient of efficiency (E1) of analysis set (1) in HYRROM

YEAR	Days Simulated	Days/Year	Obj.Function (F1)	%Error(MSE)	Qobs(mm)	Qpred(mm)	(Qobs/Qpred)	(Qo-Qm)^2	(Qp-Qo)^2	Coef. of (E1)
1964	365	1.00	1.889	0.65	445.00	447.80	1.01	92555.89	7.84	1.000
1965	365	1.00	0.039	-16.02	16.10	13.10	0.81	15542.61	9.00	0.999
1966	92	0.25	0.199	-45.3	13.40	7.30	0.54	487.60	37.21	0.924
1967	365	1.00	1.144	-74.69	145.00	36.70	0.25	17.89	11728.89	-654.505
1968	365	1.00	3.414	-62.08	389.50	147.70	0.38	61866.61	58467.24	0.055
1969	365	1.00	0.142	-31.44	37.20	25.50	0.69	10726.74	136.89	0.987
1970	366	1.00	1.010	-2.04	276.40	282.00	1.02	18291.03	31.36	0.998
1971	365	1.00	2.691	-68.71	298.20	93.30	0.31	24784.20	41984.01	-0.694
1972	151	0.41	0.131	-43.93	14.70	21.20	1.44	1895.41	42.25	0.978
1973	365	1.00	0.149	-53.14	26.50	12.40	0.47	13057.63	198.81	0.985
1974										
1975	150	0.41	0.144	-53.78	10.90	5.10	0.47	2204.37	33.64	0.985
1976	365	1.00	0.469	-76.64	58.20	13.60	0.23	6817.80	1989.16	0.708
1977										
1978	365	1.00	2.278	54.64	278.40	430.64	1.55	18942.02	23177.02	-0.224
1979	184	0.50	1.606	9.01	65.10	71.00	1.09	34.38	34.81	-0.012
1980	305	0.84	0.180	-69.81	27.90	8.40	0.30	8051.42	380.25	0.953
1981	153	0.42	1.641	18.2	79.10	93.50	1.18	403.70	207.36	0.486
1982	275	0.75	1.687	125.66	158.70	358.10	2.26	2771.01	39760.36	-13.349
1983	184	0.50	0.642	16.42	82.10	95.60	1.16	124.02	182.25	-0.470
1984	366	1.00	0.425	-59.7	17.10	7.10	0.42	15389.81	100.00	0.994
1985	306	0.84	0.783	293.37	45.10	177.40	3.93	5316.66	17503.29	-2.292
1986	365	1.00	0.180	-0.92	26.90	26.60	0.99	12966.38	0.09	1.000
1987	365	1.00	0.115	-29.94	14.70	10.30	0.70	15893.64	19.36	0.999
1988	90	0.25	0.007	-13	0.80	0.70	0.88	1149.92	0.01	1.000
1989	275	0.75	0.682	14.82	139.70	160.30	1.15	1131.68	424.36	0.625
1990	365	1.00	2.221	17.04	235.20	275.30	1.17	8917.02	1608.01	0.820

where Qobs and Qpred are the observed and predicted flows respectively (mm/yr), Qm is the long-term mean flow (mm/yr), the days/year ratio was multiplied with Qm to harmonise data set and the coefficient of efficiency (E1) was calculated using equation (7.9).

Appendix E.2 The objective function (F2), percentage error (%) and coefficient of efficiency (E2) of analysis set (2) in HYRROM

YEAR	Days simulated	Days/Year	Obj.Function(F2)	%Error(MSE)	Qobs	Qpred	Qpred/Qobs	(Qobs-Qm)^2	(Qpred-Qobs)^2	Coef. of (E2)
1964	365	1.00	1.846	1.510	445.00	451.70	0.99	92555.89	44.89	1.000
1965	365	1.00	0.038	-0.589	16.10	15.20	1.06	15542.61	0.81	1.000
1966	92	0.25	0.168	-28.880	13.40	9.50	1.41	487.60	15.21	0.969
1967	365	1.00	1.079	-70.000	145.00	42.30	3.43	17.89	10547.29	-588.468
1968	365	1.00	3.149	-59.080	389.50	159.40	2.44	61866.61	52946.01	0.144
1969	365	1.00	0.126	3.030	37.20	38.30	0.97	10726.74	1.21	1.000
1970	366	1.00	0.988	-0.490	276.40	275.00	1.01	18291.03	1.96	1.000
1971	365	1.00	2.621	-78.790	298.20	63.20	4.72	24784.20	55225.00	-1.228
1972	151	0.41	0.128	40.560	14.70	20.70	0.71	1895.41	36.00	0.981
1973	365	1.00	0.135	-25.820	26.50	19.70	1.35	13057.63	46.24	0.996
1974								0.00		
1975	150	0.41	0.132	-45.660	10.90	5.90	1.85	2204.37	25.00	0.989
1976	365	1.00	0.430	-61.510	58.20	22.40	2.60	6817.80	1281.64	0.812
1977										
1978	365	1.00	0.884	25.920	278.40	350.60	0.79	18942.02	5212.84	0.725
1979	184	0.50	1.179	70.130	65.10	110.80	0.59	34.38	2088.49	-59.746
1980	305	0.84	0.157	-46.000	27.90	15.00	1.86	8051.42	166.41	0.979
1981	153	0.42	1.083	26.330	79.10	100.00	0.79	403.70	436.81	-0.082
1982	275	0.75	1.404	47.400	158.70	233.90	0.68	2771.01	5655.04	-1.041
1983	184	0.50	0.625	5.570	82.10	86.70	0.95	124.02	21.16	0.829
1984	366	1.00	0.423	-23.750	17.10	13.50	1.27	15389.81	12.96	0.999
1985	306	0.84	0.309	8.360	45.10	48.90	0.92	5316.66	14.44	0.997
1986	365	1.00	0.173	16.730	26.90	31.40	0.86	12966.38	20.25	0.998
1987	365	1.00	0.108	9.360	14.70	16.10	0.91	15893.64	1.96	1.000
1988	90	0.25	0.007	-7.680	0.80	0.70	1.14	1149.92	0.01	1.000
1989	275	0.75	0.535	-18.820	139.70	113.40	1.23	1131.68	691.69	0.389
1990	365	1.00	1.549	25.270	235.20	294.70	0.80	8917.02	3540.25	0.603

* as described in appendix E.1

Appendix E3. The objective function (F3), percentage error (%) and coefficient of efficiency (E3) of analysis set (3) in HYRROM

YEAR	Days simulated	*Days/year*	*Obj.Function(F3)*	*% Error(MSE)*	*Qobs**	*Qpred**	*Qpred/Qobs*	*(Qobs-Qm*)^2*	*(Qpred-Qobs)^2*	*Coef.of (E3)*
1964	365	1.00	1.846	1.51	445.00	451.70	0.99	92598.49	44.89	0.9995
1965	365	1.00	0.038	-1.30	16.10	15.10	1.07	15525.16	1.00	0.9999
1966	92	0.25	0.138	0.83	13.40	13.50	0.99	486.82	0.01	1.0000
1967	365	1.00	0.943	7.84	145.00	155.30	0.93	18.49	106.09	-4.7377
1968	365	1.00	2.87	-4.59	389.50	375.50	1.04	61901.44	196.00	0.9968
1969	365	1.00	0.124	3.77	37.20	38.60	0.96	10712.25	1.96	0.9998
1970	366	1.00	0.971	-0.56	276.40	274.80	1.01	18310.02	2.56	0.9999
1971										
1972	151	0.41	0.098	-0.93	14.70	14.60	1.01	1892.89	0.01	1.0000
1973	365	1.00	0.127	-0.30	26.50	26.45	1.00	13041.64	0.00	1.0000
1974										
1975	150	0.41	0.103	5.95	10.90	11.00	0.99	2201.67	0.01	1.0000
1976	366	1.00	0.369	14.54	58.20	66.70	0.87	6806.25	72.25	0.9894
1977										
1978	365	1.00	0.771	2.77	278.40	287.80	0.97	18961.29	88.36	0.9953
1979	184	0.50	1.49	9.15	65.10	71.10	0.92	33.97	36.00	-0.0598
1980	305	0.84	0.133	-0.70	27.90	27.70	1.01	8040.93	0.04	1.0000
1981	153	0.42	1.641	18.20	79.10	93.50	0.85	404.88	207.36	0.4878
1982										
1983	184	0.50	0.62	4.32	82.10	85.70	0.96	124.81	12.96	0.8962
1984	366	1.00	0.422	0.85	17.70	17.90	0.99	15223.98	0.04	1.0000
1985	306	0.84	0.295	5.98	45.10	47.80	0.94	5308.10	7.29	0.9986
1986	365	1.00	0.18	-0.92	26.90	26.60	1.01	12950.44	0.09	1.0000
1987	365	1.00	0.108	-1.02	14.70	14.85	0.99	15876.00	0.02	1.0000
1988	90	0.25	0.006	0.64	0.80	0.80	1.00	1148.75	0.00	1.0000
1989	275	0.75	0.515	-0.32	139.70	139.20	1.00	1135.23	0.25	0.9998
1990	365	1.00	0.985	5.90	235.20	249.10	0.94	8930.25	193.21	0.9784

* as described in appendix E.1