SPATIO-TEMPORAL VARIABILITY OF RAINFALL & STREAMFLOW MODELING FOR A PENINSULAR BASIN IN INDIA

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BONAFIDE CERTIFICATE

This is to certify that the Project report entitled **"CLIMATIC VARIABILITY & RAINFALL-RUNOFF MODELING FOR A PENINSULAR BASIN USING CONCEPTUAL HYDROLOGICAL MODEL"** is a record of the bonafide dissertation work carried out by me, MOHD IZHARUDDIN ANSARI, towards the partial fulfilment of requirements for the award of degree of Master of Technology in HYDRAULICS & FLOOD CONTROL ENGINEERING.

Also, I do hereby state that I have not submitted the matter embodied in any other University/Institute for the award of any degree as per my knowledge and belief.

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ABSTRACT

Climate change is recognized to be one of the most serious challenges facing mankind today. Driven by anthropogenic activities, it is known to be a direct threat to our food and water supplies and an indirect threat to world security. Increase in the concentration of carbon dioxide and other greenhouse gases in the atmosphere will certainly affect hydrological regimes. The consequent global warming is expected to have major implications on water resources management. The increasing rate of global surface temperature is going to have significant impact on local hydrological regimes and thus on water resources, this leads to the assessment of water resources potential resulting from the climate change impacts. Main parameters that are closely related to the climate change are temperature, precipitation and runoff. Therefore, there is a growing need for an integrated analysis that can quantify the study of climate variability & its arising consequences on water resources.

In present study, Statistical trend analysis has been carried out on monthly, seasonal and annual scale using Parametric (Regression analysis) and non-parametric (Mann-kendall & Sen's Slope Estimator) methods.

Drought is a weather related natural hazards In fact, drought is complex and least understood of all natural hazards, affecting more people than any other hazard. Among all these kinds of droughts, only Meteorological drought is important from our study point of view. The Standardized Precipitation Index (SPI), which is one of the most commonly used and recommended drought indicator, is used to describe meteorological droughts. It was designed to enumerate precipitation deficit for multiple timescales. There are many indices to measure meteorological dryness, such as simple rainfall devation from historic norms, palmer drought severity index (PDSI) and standardized precipitation index (SPI). Among these indices, SPI has been widely used in recent years because of its computational simplicity and reliable interpretation. In our study, Standardized Precipitation Index (SPI) has been used for drought characterization in Chaliyar basin.

The study presented in this paper constitutes an initial approach to the problematic task of evaluating the spatial variation of ground water levels. In our study, spatiotemporal analysis of groundwater level fluctuations of 29 piezometric wells using geostatistical analysis was performed for chaliyar river basin. Geostatistical analysis was performed using ordinary kriging method in Arc GIS & finally the spatial prediction maps of water level were generated.

Our study describes the application of NAM (NedborAfstromnings Model), to investigate its performance, efficiency and suitability in Chaliyar river basin of kerala. It is deterministic, lumped and conceptual rainfall-runoff model that operates by continuously accounting for the moisture content in three different and mutually interrelated storages that represent overland flow, interflow and base flow. In the present study, the rainfall-runoff model has been developed, calibrated and validated using flow data at a river gauge station Kuniyil, in the Chaliyar River Basin, Kerala, India.

In brief, the present work intends to study climatic variability in the Chaliyar River Basin through (1) Precipitation Trend Analysis (2) Drought Characterization. (3) Study of Spatial Variation of Ground Water levels & (4) Rainfall-Runoff Modelling.

Keywords: NAM, Climatic Variability, Hydrological modelling, Kriging, Standardized precipitation index (SPI), Mann-kendall Test, Sen's Slope Estimator Test

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GLOSSARY

CWC	CENTRAL WATER COMMISSION
CWRDM	CENTRE FOR WATER RESOURCE DEVELOPMENT & MANAGEMENT
DEM	DIGITAL ELEVATION MODEL
ET	EVAPOTANSPIRATION
G.I.S.	GEOGRAPHICAL INFORMATION SYSTEM
IMD	*INDIAN METEOROLOGICAL DEPARTMENT*
MSL	MEAN SEA LEVEL
NAM	NEDBOR-AFSTROMINGS MODEL
NE	NORTH EAST
PET	POTENTIAL EVAPOTRANSPIRATION
SPI	STANDARD PRECIPITATION INDEX
SW	SOUTHWEST
SRTM	*SHUTTLE RADAR TOPOGRAPHY MISSION*
T_{avg}	DAILY AVERAGE TEMPERATURE (°C)
T_{max}	DAILY MAXIMUM TEMPERATURE (°C)
T_{min}	DAILY MINIMUM TEMPERATURE (°C)

CHAPTER 1

INTRODUCTION

1.1 GENERAL

Climate is a measure of the average pattern of variation of meteorological parameters (precipitation, humidity, temperature and others) in a given region over long period of time. Rising amount of greenhouse gases in the atmosphere can cause drastic changes in climate.

Climate change may be described as a change in the climate which can be recognized by changes in the statistical distribution of weather variables for a longer duration of time. Climate change is expected to have an impact on hydrological systems because of changes in precipitation, temperature and evapotranspiration, which are the primary input variables for the terrestrial part of the hydrological cycle. The fourth assessment report of the Intergovernmental Panel on Climate Change (IPCC) says that the annual average river runoff and water availability are projected to increase by 10-40% at high latitudes and in some wet tropical areas and decrease by 10-30% over some dry regions at mid-latitudes and in the dry tropics, some of which are presently water-stressed areas. Climate variability and change are expected to alter regional hydrological conditions & hydrological cycle and result in a variety of impacts on water resource systems throughout the world.

Quantitative estimation of the hydrological effects of climate change will be helpful in understanding potential water resource problems and making better planning decisions. With economic development and increase in population, the conflict between water use and water supply will become increasingly grave in the future. Understanding the possible impacts of climate change on water resources is of utmost importance for ensuring their appropriate management and utilization but since Most of river catchments in India are ungauged and the runoff information is not available for those catchments. Under such circumstances rainfall-runoff model can be developed to simulate the natural hydrological processes to estimate runoff from the catchment.

Hydrological modeling is simplified description of hydrological cycle to imitate the natural system. A rainfall-runoff model is a mathematical representation describing the rainfall-runoff relations of a catchment area, drainage basin or watershed. More precisely, it produces the surface runoff hydrograph as a response to a rainfall as an input.

In other words, Hydrological modeling is a mathematical representation of natural processes that influence primarily the energy and water balances of a watershed. The main purpose of using hydrological modeling is to provide information for planning and management of water resources in a sustained manner i.e. the hydrologic response of catchment to rainfall, estimates of catchment yield, and runoff data are of vital importance for hydrological analysis for the purpose of water resources planning, flood forecasting, pollution control and many other applications.

Temperature and precipitation are the key parameters of climate and variations in the pattern of these variables can affect human health, economic growth and development. An increase or decrease in precipitation pattern can result in the increase in the frequency of floods, instances of droughts and impact on water quality. Increase in Earth's temperature results in an increase in evaporation and cloud formation to occur, which increases precipitation, indicating that temperature and precipitation are interconnected. Therefore, it is necessary to carry out statistical analysis to find the trend for the most important important climatic parameters i.e. precipitation. The statistical analyses used in the present study are the Mann Kendall Test (Mann 1945, Kendall 1975) and Sen's Slope estimator (Sen, 1968). The Mann Kendall test has been used to detect trends in the time series of the precipitation and temperature. Sen's slope estimator has been used to find out the magnitude of the detected trend.

Possible climate change, in the sense of variations in temperature and, above all, in the quantity and intensity of precipitation (Eckhardt and Ulbrich 2003), may have a marked influence on the volume of water that comprises aquifer recharge (Scibek and Allen 2006), either increasing or decreasing it. Determining just how these variations affect natural water recharge to aquifers is thus a question of crucial importance, as water recharge constitutes a basic element in the water balance, and knowledge and evaluation of this parameter is absolutely essential to the efficient management of water resources (Sophocleous 1991 in Marechal et al. 2006).

1.2 HYDROLOGICAL MODELLING

Hydrological modeling is a mathematical representation of natural processes that influence primarily the energy and water balances of a watershed. The main purpose of using hydrological modeling is to provide information for planning and management of water resources in a sustained manner. Hydrological models are of two major types: (a) Stochastic models and (b) Process-based models. Stochastic models are based on mathematical and statistical concepts to relate a particular input (e.g. rainfall) to the model output (e.g. runoff) and also tries to compute the errors in model outcomes. Some of the stochastic hydrological models used are: transfer functions, artificial neural networks, and others. Process-Based Models represents the physical processes (surface runoff, subsurface flow, evapotranspiration) observed in the real world. Some of the process based hydrologic models are: SWAT (Soil and Water Assessment Tool), Variable infiltration Capacity (VIC), MIKE-SHE model (Dadhwal et al., 2010).

The rainfall-runoff process or hydrological modelling is a complex activity as it is influenced by a number of implicit and explicit factors such as precipitation distribution, evaporation, transpiration, abstraction, watershed topography and soil type.

Hydrologic models especially simple rainfall-runoff models are widely used in understanding and quantifying the impacts of land use changes, and to provide information that can be used in land-use decision making. Many hydrologic models are available; varying in nature, complexity and purpose (shoemkar et. al 1997). Various models have been developed to solve the rainfall-runoff relationship in engineering research and practices. The widely known rainfall-runoff models identified are the Rational method (Mcpherson, 1964), Soil Conservation Services (SCS) Curve Number method (Maidment, 1993), and Green-Ampt method (Green, 1911).

In present study, an appropriate hydrological model, the NAM (Nedbor-Afstromings model) has been identified & rainfall-runoff modeling has been carried out for CHALIYAR river basin in Kerala state.

1.3 MIKE 11 NAM

The hydrologic model used in the present study is NAM; which is a lumped rainfall-runoff model originally developed by the department of Hydrodynamics & water resources at the Technical University of Denmark. The model forms part of the mike 11 river modelling system for simulation of rainfall-runoff process in sub catchments.

NAM is the abbreviation of the Danish “Nedbor Afstromnings Model” means precipitation runoff model.NAM model is a part of MIKE 11 river module which simulates the rainfall-runoff process occurring at the catchment scale.

NAM is a deterministic, lumped and conceptual rainfall-runoff model. It operates by continuously accounting for the moisture content in four different and mutually interrelated storages, which represent snow, overland flow, interflow & base flow. This NAM model is a well-proven engineering tool that has been applied to a large number of catchments around the world, representing many different hydrological regimes and climatic conditions. MIKE11 NAM is a professional engineering software package, manufactured by Danish Hydraulic Institute, Denmark for water resource planning and management applications.

The basic input requirements for the NAM model consists of:

- Modal parameters
- Initial conditions
- Meteorological data
- Stream flow data for model calibration & validation

The basic meteorological data requirements are:

- Rainfall
- Potential evapotranspiration

1.4 PRECIPITATION TREND

Statistic and probability plays an important role in scientific and engineering community (**Ayyub and McCuen, 2011**) because statistical tools help to detect spatial and temporal trends for hydrological and environmental studies. Major schemes or projects are formulated based on the historical behaviour of environment under uncertain climatic conditions. Therefore, a study of trend assists to investigate the overall pattern of change over time in hydro-meteorological variables especially for water resources project on temporal and spatial scales.

Rainfall has been widely considered as one of the starting point towards the apprehension of climate change courses. Various studies have indicated due to climate change, rainfall pattern is most likely to change which would have adverse impacts on lives and livelihoods of millions of people. Analysis of the general rainfall trend is vital in understanding the underlying features, for the purpose of forecasting and in identifying the changes and impacts that are very crucial for an agro-based economy like the one of India.

Trends in data can be identified by using either parametric or non-parametric methods, and both the methods are extensively used. Testing the significance of observed trends in hydro-meteorological time series has received a great attention recently, especially in connection with climate change. The changing pattern of rainfall deserves urgent and systematic attention for planning, development, utilisation and management of water resources.

In the present study, to analyze the trends of the rainfall series of each individual station, the popular statistical methods; simple regression method (parametric), Mann-Kendall test and Sen's estimator of slope method (non-parametric) have been applied.

Among which Mann Kendall test has been used to detect significance of the trends in the time series of the precipitation & Sen's slope estimator has been used to find out the magnitude of the detected trend.

1.5 SPATIAL VARIABILITY OF GROUND WATER LEVELS

Rapid industrial development, urbanization and increase in agricultural production have led to decline in freshwater shortages in many parts of the world. To ensure proper supply of water for various purposes like agricultural, domestic and industrial, a greater emphasis is being laid for proper planning and optimal utilization of water resources. The water requirement for agriculture, municipal and industries is larger than the annual recharge. This may lead to depletion of ground water. On other hand, continuous withdrawals from groundwater reservoir in excess of replenishable recharge may result in lowering of water table & also the phenomenon like climate change, in the sense of variations in temperature and, above all, in the quantity and intensity of precipitation (Eckhardt and Ulbrich 2003), may have a marked influence on the volume of water that comprises ground water recharge recharge (Scibek and Allen 2006), either increasing or decreasing giving rise to spatial variation in ground water levels.

Ground water level fluctuation space to space (Spatial) and time to time (Temporal) is a major problem in India and the assessment of spatio-Temporal characteristics water level fluctuation trend is very important in the point of view of future development. Moreover, to have an intimate understanding of the changes in water level fluctuations, it is also important to relate them to the surrounding geomorphic, structural, climatic and geologic factors. This research serves two fold. The first one is to operationalize the use of RS and Geographic Information Systems (GIS) techniques to assess the change of water surface area in Chaliyar River Basin. The second is to present and interpret the available statistical data on water level fluctuations.

The techniques RS and Geographic Information Systems (GIS) used in our study of ground water level fluctuations is known as Kriging technique which is explained in the later part of thesis.

1.6 DROUGHT CHARACTERIZATION

Drought is a weather related hazards. However, there is no universally accepted definition of drought. Perhaps the most general definition is the one which considers drought as significant decrease of water availability during a long period of time and over a large area. In fact, drought is complex and least understood of all natural hazards, affecting more people than any other hazard (wilhite, 2000). Drought is a natural hazard affects about one third of the world's total population. On average more than 0.5 billion people in china and India are usually exposed to droughts, seriously affecting the economic development and environment of the region. Because of the slow development of a drought, people are often not aware of the emerging of droughts in time. More insights in the development of drought can help the people to be aware of a drought in an earlier stage. Drought impacts the poorer economies to a large extent and may cause fatalities as compared to developed countries. On an average, more than 30% of the population is exposed to drought annually in western African countries, thereby seriously threatening the livelihoods (ISDR, 2009; WWDR, 2009).

The drought has many facets and it always starts with the lack of precipitation, but may (or may not, depending on how long and severe it is) affects soil moisture, streams, ground water, ecosystem and human beings. This leads to the identification of different types of droughts viz., meteorological, agricultural, socio-economic and physiological droughts, which reflects the perspectives of different sectors on water shortages (Smakhtin and Hughes, 2004).

Classification of drought

Droughts are often classified in to four different categories:

1. Meteorological drought (deficit in precipitation)
2. Hydrological drought (deficit in surface water, ground water, and reservoir storage)
3. Agricultural drought (deficit in soil moisture)
4. Socio economic drought (imbalance in water supply and demand)

Among all these kinds of droughts, in the present study, only Meteorological drought is important from our study point of view. Meteorological drought is the earliest and the most explicit event in the process of occurrence and progression of drought conditions. Rainfall is the primary driver of meteorological drought. There are numerous indicators based on rainfall that are being used for drought monitoring (Smakhtin and Hughes, 2007). Deviation of rainfall from normal i.e. long term mean, is the most commonly used indicator for drought monitoring.

Meteorological drought is declared based on rainfall deviations measured using the season's total actual rainfall and long term mean rainfall. If the total season's rainfall is less than 75% of the long term mean, the meteorological sub-division is categorized to be under drought. Severe drought occurs when the season's rainfall is less than 25% of normal (www.imd.gov.in).

In india drought prone areas fall in to these broad regions of the country: the plateau region, which embodies the state of Andhra Pradesh, Karnataka, Madhya Pradesh, Maharashtra, Orrisa, Tamilnadu, Bihar, West Bengal and Uttar Pradesh; the desert region, which embodies the state of Rajasthan and Gujarat; and a few districts in the state of Haryana and Jammu Kashmir (Ramasastri and Pandey, 2002) & even the history of droughts have also been reported in the past in our study area i.e. Chaliyar river basin, Kerala(source:Internet).

There are many indices to measure meteorological dryness, such as simple rainfall devation from historic norms, palmer drought severity index (PDSI) and standardized precipitation index (SPI). Among these indices, SPI has been widely used in recent years because of its computational simplicity and reliable interpretation. SPI is a simple and more effective method for studying drought climatology (Lloyd-Hughes and Saunders, 2002).

1.7 STANDARD PRECIPITATION INDEX (SPI)

The Standard Precipitation Index (SPI) is a tool which was developed chiefly for describing and monitoring drought. It permits determining the rarity of a drought at a given time scale of interest for any rainfall station with historic data. It can also be used to determine periods of anomalously wet events.

The SPI (McKee et al, 1993) is a powerful, flexible index that is simple to estimate. The probability of observed precipitation is converted in to an index. Many drought planners appreciate the SPI's versatility. It is also used by a variety of research institution, Universities, and National Meteorological and Hydrological services across the world as part of drought monitoring and early warning efforts. In fact precipitation is the only required input parameters. In addition, it is just as effective in analysing wet periods/cycles as it is analysing dry periods/ cycles. The SPI is the number of standard deviations that the observed value would deviate from the long-term mean, for a normally distributed random variable. Since precipitation is not normally distributed, a transformation is first applied so that the transformed precipitation values follow a normal distribution.

The SPI was designed to enumerate precipitation deficit for multiple timescales. These timescales reflect the impact of drought on the availability of the different water resources.
The SPI estimation for any location is based on the long term precipitation record for the desired period. This long term record is fitted to a probability distribution, which is then transformed in to a normal distribution so that the mean SPI for the location and the desired period is zero (Edwards and McKee, 1997).

Positive SPI values designate greater than median precipitation while negative values designate less than median precipitation. Because the SPI is normalized, wetter and drier climates can be represented in the same way; thus, wet periods can also be monitored using the SPI.

1.8 OBJECTIVES

- Longterm Spatio-Temporal Analysis of Rainfall.
- Drought Characterisation.
- Spatial Variation of Ground Water Levels in **GIS** Environment.
- Rainfall Runoff Modeling using Mike 11 NAM Model.

1.9 STRUCTURE OF THE THESIS

Chapter I is the introduction part of the research work.

Chapter II describes literature reviews related to precipitation trend, Standard Precipitation Index (SPI), spatial variation of ground water levels, hydrological modelling & mike 11 NAM

Chapter III gives a description of the study area Chaliyar River basin, topographic information, rainfall, temperature, soil characteristics and land use pattern and also about the data & tools used in the study.

Chapter IV deals with different methodology adopted in the present study.

Chapter V illustrates and discuss results.

Chapter VI gives the general conclusions resulting from the analysis and modelling techniques used in this study.

CHAPTER 2

LITERATURE REVIEW

Reviews of Literature

This chapter deals with the relevant review work done for the proposed study for which Literature has been reviewed and arranged in the following sections:

First section: Review of Studies for statistical Trend Analysis of Rainfall

Second section: Review of Studies for Standardized Precipitation Index

Third Section: Review of Studies for Spatial variation of Groundwater levels and Kriging technique

Fourth Section: Review of Studies for related Hydrological Modelling

Fifth section: Review of Studies for MIKE 11 NAM model

2.1 RAINFALL VARIABILITY & PRECIPITATION TREND

P. P. N. RAJ & P. A. AZEEZ (2010) examines the general trend of rainfall in the Palakkad plains (South India) using monthly rainfall data collected from four rain gauge stations available in the area. As the years proceed, the annual rainfall pattern of all the stations showed a trend of significant decline.

Arun Rana et al (2011) studies rainfall trends in Delhi and Mumbai & determined long-term trends in Rainfall by Man-Kendall rank statistics and linear regression. Further this study investigates precipitation trends during monsoon period by different global climate phenomena & used Principal component analysis and Singular value decomposition to find relation between southwest monsoon precipitation and global climatic phenomena using climatic indices.

Telemu Kassile (2013) examines the evolution of rainfall and ascertains whether the observed time trend in rainfall is significant in statistical terms. He employed Linear regression analysis through OLS estimation and the Kruskal-Wallis and Mann-Kendall's tau test.

Sanda rajitha & A.C. Narayana (2014) provides an assessment of climate change variability based on analysis of historical data of rainfall and temperatures at Warangal district, Andhra Pradesh for the period of 1960-2012. Long term changes in rainfall determined by Man-Kendall, Sen's slope and regression analysis.

Zaheed Hasan et al (2014) examines temporal variability of precipitation for the region of south-east coast of Bangladesh over the period of 1949-2011 & used Mann-Kendall test and the Sen's slope estimators to detect rainfall trends and to understand magnitude of changes.

S. Adarsh and M. Janga Reddy (2014) studied long-term trends of rainfall using linear regression, nonparametric Mann–Kendall (MK) test, Sen's slope estimator methods and the sequential MK (SQMK) method. Then the trend analysis based on discrete wavelet transform (DWT) in conjunction with SQMK method is performed on the post-monsoon rainfall time series of Kerala.

Chithra N R & Santosh G Thampi (2015) worked to statistically detect climatic change signals in the monthly precipitation data of the Chaliyar river basin, Kerala, India & evaluates the factors contributing to it. He downscaled Precipitation data from the General Circulation Models statistically to river basin scale using ANN based models and identified potential predictors by correlation coefficient analysis.

2.2 Standardized Precipitation Index (SPI)

M. Naresh Kumar et al (2009) use Standardized Precipitation Index (SPI) for intensity assessment of drought. He analysed monthly rainfall data from June to October for 39 years to compute Standardized Precipitation Index (SPI) values which is based on two parameter gamma distribution for districts having low & high rainfall in Andhra Pradesh state, India.

Christos A. Karavitis et al (2011) used standardized Precipitation Index (SPI), for more appropriate understanding of drought duration, magnitude and spatial extent in semi-arid area like Greece. He marked importance of this Index in its simplicity and its ability to identify the beginning and end of a drought event, drought contingency planning and drought alert mechanisms.

Khan, M. A. & M. S. Gadiwala (2013) applied SPI to precipitation datasets(60 years) in Sindh (Pakistan), for the Study of Drought as it is a region experiencing frequent drought events & he assessed SPI's capacity to analyse historical records and compare different series. He concluded that for Sindh region agriculturists should use SPIs of 12 months or less & water resource managers should apply 36 months SPI.

Mirja Kattelus et al (2015) studied the effects of rainfall variation on rice yield with the help of regression models using the Standardized Precipitation Index (SPI) as an explanatory variable & his study indicate that in large part of the study region, a strong relationship between precipitation and rice yields exists and the SPI at various lags chosen as the predictor variable performed well in describing the inter-annual yield variability.

Sandeep Kumar & Santosh (2015) performed trend analysis by using both non-parametric (Mann-Kendall test) and parametric (linear regression analysis) procedures. For better understanding of the observed trends, data were computed into standardised precipitation indices (SPI). These standardised data series were plotted against time and the linear trends observed were represented graphically.

2.3 SPATIAL VARIATION OF GROUND WATER LEVELS & KRIGING TECHNIQUE

Dewashish Kumar and Shakeel Ahmed (2003) studied Seasonal behaviour of spatial variability of groundwater level in a granitic aquifer in monsoon climate. He used monthly water-levels from 32 wells evenly distributed over the area to analyse the variability and to calibrate a groundwater flow model. He applied Universal kriging technique with a linear drift to examine the available groundwater levels during different periods for one cycle during the year 2000.

H. Aguilera & J. M. Murillo (2008) studied initial approach to evaluate the effects of possible climate change on natural water recharge to aquifers. He used a purpose-designed mathematical model termed Estimation of Recharge in Over-exploited Aquifers (ERAS) which enables to simulate the monthly water recharge to an aquifer.

Khadri S. F. R. & Kanak Moharir (2015) studied Hydrogeology Investigation & Water Level Fluctuation in Hard Rock of the Man River Basin, Akola and Buldhana Districts, Maharashtra, India. His research serves two fold. Firstly, to operationalize the use of RS and Geographic Information Systems (GIS) techniques to examine the change of water surface area in Man River Basin in Akola and Buldhana Districts. The second is to present and interpret the available statistical data on water level fluctuations.

Md. Manjurul Hussain et al (2016) performed spatio-temporal analysis of groundwater level fluctuations of 32 piezometric wells with the help of geostatistical analysis using ordinary kriging & empirical Bayesian kriging (EBK) methods. He founded that EBK performs better as compared to ordinary kriging in representation of the spatial groundwater level fluctuations.

2.4 Rainfall-Runoff Modeling

Rainfall-runoff modeling is the interaction of an input (e.g. rainfall) with a system (e.g. catchment) to produce an output (e.g. the outflow hydrograph).Rainfall-runoff upon which the whole design of hydraulic structures and conservation works depends, offers a prospective field of work for research and soil conservationists.

Nash (1958) considered watershed as consisting of a series of identical reservoir and prepare a conceptual models by routing a unit inflow through the reservoirs and in the form of instantaneous hydrograph equation:

$u(0,t)=1/k\sqrt{n}(t/K)^{n-1}(e-t)/K$ (2.1)

Where,

u (o,t), is the instantaneous ordinate of unit hydrograph at time t,

k, is storage parameter(hr),

t, is time after beginning of surface runoff (hr),

n, is dimensional parameter.

Rogers *et al.* (1966) studied to determine the effect of both intensity and amount of simulated rainfall and length of slope, and the correlation between these factors on runoff and soil loss. They observed the indications of interaction between effect of slope length with both rainfall intensity and amount of soil loss. Rainfall amount explained approximately 89% of the variation in total runoff. There was, however some interaction of rainfall intensity and amount. Runoff increased with increased initial moisture content of soil loss.

Linsley (1982) classified rainfall-runoff models as deterministic, stochastic, conceptual, theoretical, black box, continuous, event, complete, routing or simplified. Existing models can be categorized according to Linsley's classification. It will also become clear that specific models may be classified according to more than one attribute.

Kumbhare and Rastogi (1984) tested the Nash conceptual model and found that runoff generated by the two parameters resulted in good agreement with actual runoff hydrograph. The peak time coincide with trend of rising, crest and falling limb.

Pathaket al. (1984) developed a model to predict runoff volume from small watershed which was based on the modified soil conservation services curve number technique and on a soil moisture accounting procedure. This model was tested with data from seven vertisol watershed at ICRISAT centre; the model simulated daily monthly and annual runoff volume quite accurately. The model also predicted runoff fairly accurately from big runoff events.

Kumar and Rastogi (1989) developed a mathematical model of the instantaneous unit hydrograph based on time area histogram for a small watershed at Pantnagar. The instantaneous unit hydrograph was used for generation of runoff hydrograph.

Clarke (1994) classified the rainfall runoff model structures depending on the degree of the physical abstraction from the real world system into three broad types:

1) Distributed Physically-based Models which are based on the complex law of physics generally expressed as systems of non-linear partial differential equations.
2) Systems-Based (Black or Grey box) Models which make little or no attempt to simulate the individual constituent hydrologic processes and which rely heavily on systems theory developed in other branches of engineering science. The essence of these models is the empirical discovery of transfer functions which interrelate in the time domain the input (usually rainfall) and the output (usually discharges) functions.
3) Quasi-Physical Conceptual models which occupy an intermediate position between the other two types of models in terms of complexity, disaggregation and data requirements.

Shoemaker et al. (1997) carried out a result that hydrological models, especially simple rainfall-runoff models, are widely used in understanding and quantifying the impacts of land-use changes, and to provide information that can be used in land-use decision making. Many hydrologic models are available, varying in nature, complexity, and purpose.

Legesseet al. (2003) used a physically based distributed model to investigate the hydrological response of a catchment to climate and land use changes in south central Ethiopia.

Jettenet al. (2003) summarized the results of model comparison workshops. They concur with the generally held viewpoint that the predictive quality of distributed models is reasonably good for total discharge at the outlet and fair for net soil loss. The difficulties associated with calibrating and validating spatially distributed soil erosion models are due to the large spatial and temporal variability of soil erosion phenomena and the uncertainty associated with the input parameter values used in the models.

Henrik Madsen (2004) introduced an automatic calibration and uncertainty assessment in rainfall runoff modelling. The procedure considers multiple calibration objectives, including

1) A good simulation of the water balance,
2) A good overall agreement of the shape of the shape of the hydrograph,
3) A good agreement of peak flows, and
4) A good agreement for low flows.

The shuffled complex evolution is applied for optimizing the different calibration objectives simultaneous. The calibration algorithm can explicitly take data and model errors into account by defining appropriate weights that indicate the importance to be given to particular portion of the hydrograph in the calibration process. The procedure is based on the comparison of the distributions of the model parameters corresponding to the model parameters corresponding to the best and worst model simulation, respectively, obtained from a Monte Carlo sampling.

Ravish and Suresh (2005) introduced rainfall-runoff sediment yield relationship for a micro watershed and to arrive at most appropriate form of relationship several type of function were tried and form of relationship generating highest correlation coefficient was selected as the appropriate relationship.

Mutukrishnanet al. (2006) stated that the long term hydrological impact assessment model is widely used to study direct runoff changes with respect to different land use conditions. Long term hydrological impact assessment was designed to assess the long term impacts on the hydrology of a watershed for users who want to determine the relative change in runoff from land use conditions to another. Some users, however, are interested in results that much observed stream-flow data, which includes both direct runoff and base flows. A simple method of calibration of the long term hydrological impact assessment using linear regression of L-THIA.

Raneesh K. Y et al (2010) calibrated and validated a physically based hydrologic model, SWAT for prediction of streamflow, sediment yield, and nutrient load in a part of the Chaliyar river basin in Kerala, India. He demonstrated that SWAT is capable of predicting streamflow, sediment yield and nutrient loads from the catchment reasonably well and can be used to evaluate the influence of alternate management practices in controlling erosion and pollution.

K. Y. Raneesh & G. Thampi Santosh (2011) worked to present a general approach for evaluating the impacts of potential climate change on streamflow in a river basin in the humid tropical zone of India. In his work, the projections of a GCM for two scenarios, A2 and B2 are downscaled by a RCM to project future climate in a watershed. Projections for two important climate variables, viz. rainfall and temperature were made. These are then used as inputs for a physically-based hydrological model, SWAT, in order to evaluate the effect of climate change on streamflow and vegetative growth in a humid tropical watershed.

Pankaj Kumar & Devendra Kumar (2012) studied an adaptive neuro-fuzzy inference system was used for rainfall-runoff modelling for the Nagwan watershed in the Hazaribagh District of Jharkhand, India. Different combinations of rainfall and runoff were considered as the inputs to the model, and runoff of the current day was considered as the output.

Ansoumana Bodian et al (2016) simulate and extend hydrological data, using the GR2M rainfall-runoff model. A sensitivity analysis of the model to rainfall and water holding capacity input data was performed by him. The best combination of input data was chosen by catchment based on the Nash-Sutcliffe criterion. Then cross calibration-validation tests were performed.

2.5 MIKE 11 NAM

The NAM model is a well-proven engineering tool that has been applied to a large number of catchments around the world, representing many different hydrological regimes and climatic conditions. MIKE11 NAM is a professional engineering software package developed by Danish Hydraulic Institute, Denmark for water resource planning and management applications.

R. V. Galkate et al describes the application of NAM (NedborAfstromnings Model), to investigate its performance, efficiency and suitability in Bina river basin of Madhya Pradesh. The model was found efficient with Efficiency Index 81% and found capable of predicting runoff for extended time period in Bina basin.

Fleming (1975) concluded that the RMSE values tend to be zero for perfect agreement between observed and simulated value. Root Mean Square Error (RMSE) method used by was another method applied to evaluate the reliability of MIKE11 during this study. This method can be regarded as a measure of absolute error between the computed and observed flows.

Kjelstrom and Moffat(1981) developed the equations which may present reasonable estimations of stream flow, results are average monthly values that will likely under predict some peak events and over predict critical low flow periods. In order to evaluate irrigation practices, the USRMBM (upper salmon river mike basin model) and EFMBM (east fork salmon river mike basin model) have been established to simulate daily flow thus use of the monthly values are expected to hinder the goal of addressing critical low flow periods that may extend for only a short period during certain years. The USGS (United States of geological survey) also developed methods for evaluating flow-frequency and flow duration statistics in un-gauged basins based on catchment area and flood magnitude. In addition, changing climatic conditions are not easily addressed without reformulating the regression analysis.

DHI (1990) studied the case of continuous modelling; the results indicate more accurate outcomes. Nevertheless, it is considered that some of the disagreements are due to the lack of more detailed spatial rainfall information. In this instance, the basins size did not constrain the model application since other studies showed its employment in greater catchments.

Lipscomb (1998) estimated the method for developing stream flow time series is the use of regional hydrologic curves or equations that predict peak and monthly statistical flows. The US Geological Survey (USGS) has developed regional equations that provide peak annual and monthly average runoff for a stream in the Salmon River drainage.

Arcelus (2001) discussed the combination of an event model (HEC-HMS) and a continuous water accounting model (NAM) to obtain discharge series from un-gauged basins. An application of the methodology to Cebollatí River basin (Uruguay) is presented. Considering the lack of detailed spatial rainfall information, the applied procedure showed acceptable results. The extrapolation of direct runoff from gauged watersheds to un-gauged ones is not a simple task.

Shamsudin and Hashim (2002) deliberate the Layang River rainfall runoff estimation using MIKE 11 NAM model. The calibration and validation procedures were carried out to provide a satisfactory estimation. The runoff discharges were simulated for a 12-year period (1988-2000). The simulated peak flow occurred in 1992 and 1995 with approximate values of 20.94 m^3/s and 18.93 m^3/s respectively. The optimum values of the model parameters obtained during the calibration procedure were presented. The reliability of MIKE11 NAM was evaluated based on the Efficiency Index (EI) and Root Mean Square Error (RMSE). The EI and RMSE obtained during this study were 0.75 and 0.08 respectively.

DHI (2003) described that NAM is a rainfall-runoff model that operates by continuously accounting for the moisture content in three different and mutually interrelated storages that represent overland flow, interflow, and base flow. As NAM is a lumped model, it treats each sub-catchment as one unit, therefore the parameters and variables considered represent average values for the entire sub-catchments. Precipitation in the form of snow is modelled as a fourth storage unit. For catchments with snow falling over a wide elevation range, the storage unit representing snow can be divided in up to ten subunits to represent different elevation zones. Water use associated with irrigation or groundwater pumping can also be accounted for in NAM. The result is a continuous time series of the runoff from the catchment throughout the modelling period. Thus, the NAM model provides both peak and base flow conditions that accounts for antecedent soil moisture conditions over the modelled time period.

Doulgeris et al. (2008) studied MIKE 11 NAM was used for the simulation of rainfall-runoff process in the Strymonas river and Lake Kerkini for water resources management aspects.

Makungo et al. (2010) studied the runoff hydrographs for the un-gauged Nzhelele river were simulated using MIKE 11 NAM model and the Australian Water Balance Model (AWBM). MIKE 11 NAM is a rainfall runoff model which is a part of the MIKE 11 RR module. MIKE 11 NAM, MIKE SHE and WATBAL models were validated on three catchments in Zimbabwe for water resources decision making (Refsgaard and Knudsen, 1996), where at least one year's data were available for calibration.

Doulgeris et al. (2012) studied the rainfall-runoff relationship in the Strymonas river catchment by using the MIKE 11 NAM model.

Md. Sharif Imam Ibne Amir et al. (2013) presented a large scale hydrological model for the Fitzroy Basin using the MIKE 11 NAM modelling system. This model simulates the rainfall runoff processes for the three different and interrelated storages which are surface storage, root zone storage and groundwater storage. The parameters of such model cannot be obtained directly from measurable quantities of catchment characteristics. Hence model calibration is required to get the model parameters. Manual calibration is very time consuming and it only focuses on a single objective function. Thus tradeoff exists between the different objectives. And in the study, an automatic calibration was done considering the four multiple objectives and obtained the optimal values of the model parameters for each catchment.

Maryam Hafezparast et al. (2013) described the application of a conceptual rainfall runoff model to investigate the peak and monthly flows at the Sarisoo River Basin on the North West of Iran. The model was calibrated using measured stream flow data and then validated for three years. The outputs of the calibrated model are able to be used in the assessment of water resources management models like Mike Basin, WEAP because they normally work based on monthly flows with a large time horizon. The results show that monthly averages of mean, maximum and minimum flows are about 10%, 2% and 33% less than daily computed Nash–Sutcliffe coefficients, all calculated over a period of 12 years.

Singh et al (2014) envisaged the rainfall-runoff modeling using MIKE 11 NAM model in Vinayakpur intercepted catchment in Chhattisgarh state. The model was calibrated using measured stream flow data for the period 2001 to 2004 and then validated from period 2005 to 2007.The outputs of the calibrated model were used in water resources management model *viz.*, MIKE basin as they normally work based on monthly flows with a large time horizon. The optimum values of nine NAM model parameters obtained during calibration procedure were used for simulation. The reliability of MIKE 11 NAM was evaluated based on Nash-Sutcliffe coefficient, correlation coefficient (r^2) and root mean square error (RMSE).

CHAPTER 3

STUDY AREA

&

MATERIALS USED

3.1 STUDY AREA

The study area is the Chaliyar river basin in Kerala, India, situated between 11^0 30'N and 11^0 10'N latitudes and 75^0 50'E and 76^0 30'E longitudes falling in Survey of India (SOI) degree sheets 58A and 49M. Chaliyar River forms the third largest river in Kerala, originates from the Elambalari hills, Nilgiri District of Tamil Nadu, at an elevation of about 2066m above mean sea level (MSL).

Chaliyar is a perennial river & flows along the northern boundary of Malappuram district through Nilambur, Mambad, Edavanna, Areakode and Feroke & the river joins the Lakshadweep Sea south of Kozhikode near Beypore after flowing over a distance of about 169 kms in the name "Beypore" River.

This river has a total drainage area of 2918km^2 out of which 2530km^2 lies in Kerala State and the remaining area falls in Tamil Nadu.

Six major streams Chaliyarpuzha, Punnapuzha, Kanjirapuzha, Karimpuzha, Iruvahnipuzha and Cherupuzha constitute the Chaliyar River drainage system. The basin comprises parts of four districts viz. Kozhikode district cover an area of 626 km^2 in the northwest, Wayanad district over an area of 112 km^2 in the north, Malappuram district spreads over an area of 1784 km^2 in the east and south and Nilgiri district of Tamil Nadu over an area of 378 km^2 in the northeast. The watershed is predominantly agricultural lands (74.26 %) and forests (14.21 %). The remaining area comprises urban areas, rocky areas and water bodies.

Of the total area of the river basin, only an area of 1996.4 km^2 was considered in the study for hydrological modelling using MIKE 11 NAM model.

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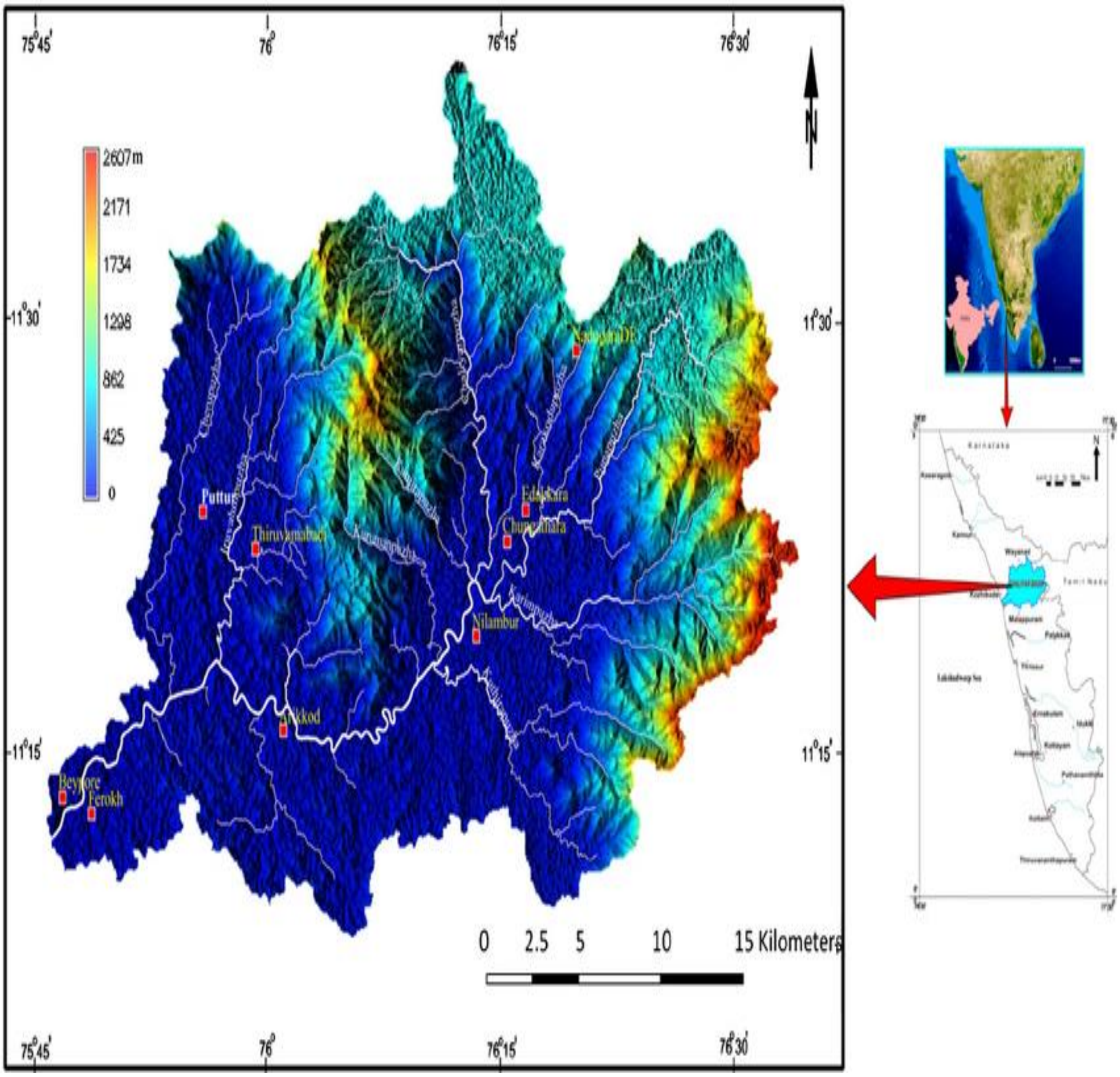

Figure 3.1: Digital elevation model (DEM) aspect map of the Chaliyar River drainage basin. Source: Google Earth.

3.1.1 Climate

The basin enjoys a tropical humid climate with sweltering summer and high monsoon rainfall. Generally March and April are the hottest and December and January are the coolest. The maximum temperature ranges from 22ºC to 32.9ºC and the minimum temperature ranges from 22ºC to 25.8ºC. The average annual maximum temperature is 30.9ºC and minimum is 23.7ºC. The temperature starts rising from January reaching the peak in April. It decreases during the monsoon months. On an average about 3000 mm of rainfall occurs annually in the basin.

The principal rainy seasons are the southwest (June-September) and northeast (October-November) monsoons in India. The pre-monsoon months (March-May) are characterized by major thunderstorm activity and the winter months (December-February) by minimal cloudiness and rainfall (Ananthakrishnan et al. 1979). *Sahyadri* (Western Ghats) has a significant influence on the intensity and distribution of rainfall over Peninsular India. As a mountain barrier, the *Sahyadri* polarizes precipitation along its crest. As moist airflow during the southwest monsoon ascends, the windward slope receives copious rainfall (Anu and Mohankumar, 2004). Thus, the *Sahyadri* forms the watershed for a large number of rivers. These rivers have high run-off and sediment load during the monsoon months.

Southwestern India experiences a tropical climate with seasonally reversing wind patterns and large variations in precipitation. Along the west coast of India, the southwest (SW) monsoonal winds of oceanic origin are established by mid-May. During the SW monsoon, winds blow from southwest during May-September, but change to a north easterly direction during the northeast (NE) monsoon. These winds continue to grow strong until June, when there is a sudden burst or strengthening of the southwest winds. The winds are the strongest during July and August, but become weak in September, ahead of the NE monsoon, which lasts through October and November. The wind speed is generally 15-20 km/hr during the SW monsoon, but lower (10-12 km/hr) during the NE monsoon. Summer (southwest) monsoon (June-September) accounts for a major part of the average annual rainfall (> 300 cm), whereas the winter monsoon (October-January) accounts for about 50-60 cm rainfall. Temperature in the region ranges between 23° and 37°C (Narayana, 2006).

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Rivers in mountainous terrains commonly carry higher sediment loads and yields than do upland rivers, whose loads and yields in turn, are higher than those of lowland rivers. A better relationship was documented between the annual variability of rainfall and sediment transport. The positive relationship among rainfall, run-off and sediment discharge suggests that precipitation and run-off exert a first order control on the sediment discharge of Kerala Rivers (Narayana, 2007). Tectonic uplift/subsidence alters the fluvial regime with resultant changes in rates of sediment erosion and deposition.

3.1.2 Physiography

The Chaliyar river basin can be physiographically divided into four well-defined units viz., highland, midland, low land and coastal plains. Based on the relief pattern and topographic alignment, the basin can be divided into five physiographic sub-units.

(i) High ranges with an elevation ranging from 600m to 2600m. This form part of the Wayanad plateau and the high hill ranges with steep slopes of the Western Ghats, (ii) Foot hills of Western Ghat with elevation ranging from 300 to 600 m above MSL comprise rocky mounds and slope areas of the high hills, (iii) Upland regions consisting of the ridges and valleys, isolated hills with altitudes ranging from 100-300 m. At places these units are lateritic, (iv) Mid-land zone with elevation ranging from 10 to 100 m characterized by rolling topography with lateritic ridges, isolated hills and alluvial valleys, and (v) Low-land characterized by coastal stretches and alluvial plains with an elevation of < 10 m.

3.1.3 Soil Types

The major soil groups in the watershed are gravelly clay, clay, gravelly loam and loam.
With regard to the type of soils, the basin is dominated by loam texture (42.74%), followed by clay (28.66%), clay loam (24.18%) and sandy loam (4.42%).

3.1.4 Land Use & Land Cover

The predominant land-use is agriculture (60.04%) and forests (38.74%). Urban areas, pastures, waste lands and rocky areas comprise less than 1% of the total area. The elevation of the watershed varies from about 20m a.m.s.l. in the lowland areas to around 2250m in the hilltops; with a mean elevation of 338m and standard deviation of 458 m. With regard to the type of soils, the basin is dominated by loam texture (42.74%), followed by clay (28.66%), clay loam (24.18%) and sandy loam (4.42%). (**K. Y. Raneesh & G. Thampi Santosh, 2011)**

3.1.5 Geology of the study area

The area forms part of the Precambrian metamorphic shield with rocks of Wayanad Group, Peninsular Gneissic Complex, Charnockite Group and Migmatite Complex, which are traversed by younger basic and acid intrusive (Fig. 3.2). Small isolated capping of Tertiary deposit (Warkali Formation) is seen to the west. The Quaternary sediments unconformably overlie the basements rocks of the coastal tracts and valleys **(figure 3.2).**

Table 3.1 General geology of Chaliyar River drainage basin (GSI, 1995)

Era	Period	Group	Lithology
Quaternary	Holocene	Marine Fluvio-marine Fluvial	Sand Clay and silt Sand, silt, clay Sand
	Pleistocene	Palaeo-marine	Pebble bed
Tertiary	Mio-Pliocene Mesozoic (61-144 Ma)	Acid Intrusive Basic intrusive	Laterite Quartz vein Pegmatite Dolerite
Precambrian	Proterozoic	Migmatite Complex	Hornblende gneiss Hornblende-biotite gneiss Granite gneiss
		Charnockite Group	Charnockite/charnockite gneiss Pyroxene granulite
	Archaean	Peninsular Gneissic Complex	Hornblende-biotite gneiss Magnetite quartzite Quartz-mica schist
		Wayanad Group	Fuchsite quartzite Amphibolite Metapyroxenite Talc-tremolite-actinolite schist

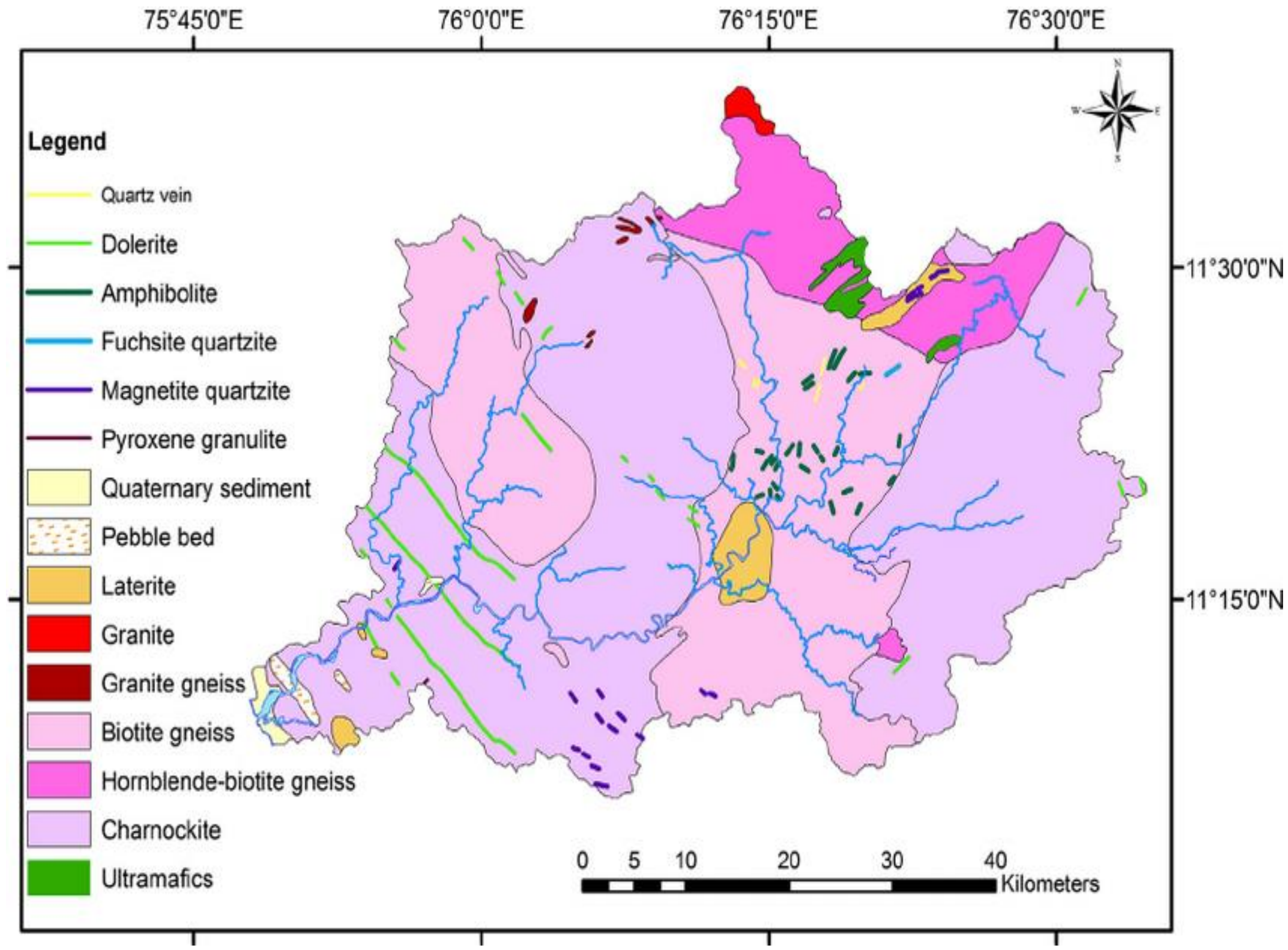

Figure 3. 2: Geological map of the Chaliyar River drainage basin

(Source: P.H.D Thesis by Ambili V submitted to the Cochin University of Science and Technology)

3.2 DATA USED

3.2.1 Meteorological Data

Monthly Rainfall data: The Daily Rainfall data has been collected from CWRDM Kerala for the investigation of Rainfall trends & Drought characterization. It was collected for different stations & for different duration, but only the monthly rainfall data of the four stations namely Ambalavayal & Kalladi (1993 to 2010), Manjeri & Nilambur(1993 to 2011) have been used .

Daily Rainfall data series has been used in Hydrological modelling for the period of 2001 to 2005 only for the IMD stations namely Ambalavayal, Manjeri & Nilambur & we obtained this data from CWRDM, kerala.

Daily temperature data: The daily maximum (T_{max}) & daily minimum (T_{min}) of Kottamparamba station for the period of 1983 to 2012 was also collected from CWRDM, kerala.

Radiation data for the basin has been obtained from the internet (globalweather.tamu.edu) for the chaliyar basin.

3.2.2 Hydrological Data

The daily discharge data measured at Kuniyil Gauge discharge site for the period 1978 to 2013 has been collected from CWRDM, Kozhikode, kerala. The data from 2001 to 2005 have been used for the modeling.

3.2.3 DEM

SRTM data at 90 m resolution has been obtained from internet.

3.3 SOFTWARE USED

- Arc GIS software has been used for stream network, watershed delineation using digital elevation model. Moreover the software has also been used for generating ground water level fluctuation maps for basin and creating thiesen polygon map for computing the weighted rainfall for the basin.
- Microsoft excel was extensively used throughout the study to prepare the data for all the four objectives proposed in the research work.
- The Hydrological model, Mike 11-NAM was used for rainfall-runoff modeling of chaliyar basin up to Kuniyil.

CHAPTER 4

PROJECT METHODOLOGY

4.1 Precipitation Trend Analysis

The term trend refers to “general tendency or inclination”. In a time series of any variable, trend depicts the long smooth movement lasting over the span of observations, ignoring the short term fluctuations. It helps to determine whether the values of a series increase or decrease over the time. In statistics, trend analysis referred as an important tool and technique for extracting an underlying pattern of behaviour or trend in a time series which would otherwise be partly or nearly completely hidden by noise.

Statistic and probability plays an important role in scientific and engineering community (**Ayyub and McCuen, 2011**) because statistical tools help to detect spatial and temporal trends for hydrological and environmental studies. Major schemes or projects are formulated based on the historical behaviour of environment under uncertain climatic conditions. Therefore, a study of trend assists to investigate the overall pattern of change over time in hydro-meteorological variables especially for water resources project on temporal and spatial scales. Trends in data can be identified by using either parametric or non-parametric methods, and both the methods are extensively used. The parametric methods are considered to be more powerful than the non-parametric methods only when the data series is normally distributed, independent and homogeneous variance (**Hamed and Rao, 1998**). Conversely, non-parametric methods are more advantageous as they only require the data to be independent and are also less sensitive to outliers and missing values.

Trend analysis of time series consists of magnitude of trend and its statistical significance. In general, the magnitude of a trend in a time series is determined either using regression analysis (parametric test) or using sen’s estimator method (non-parametric method) & significance is determined by Mann-Kendall test (non-parametric method).

In the present study, to analyze the trends of the rainfall series of each individual station, the popular statistical methods; simple regression method (parametric), Mann-Kendall test and Sen’s estimator of slope method (non-parametric) have been applied. The systematic approach has been adopted to determine the trend in three phases. Firstly, a simple linear regression method to test the long term linear trend, secondly, non-parametric Mann-Kendall test for the presence of a monotonic increasing or decreasing trend in the time series and Thirdly, the non-parametric Sen’s estimator of slope test to determine the magnitude of the trend in the time series of meteorological parameter i.e. rainfall at the basin scale. These are described in the following sections.

4.1.1 Regression model (Parametric Test)

One of the most useful parametric models to detect the trend is the "Simple Linear Regression" model. The correct application of this method requires the variables to be normally distributed and temporally and spatially independent. The method of linear regression requires the assumptions of normality of residuals, constant variance, and true linearity of relationship (**Helsel and Hirsch, 1992**).The model for Y (e.g. precipitation) can be described by an equation of the form:

$$y = m x + c \qquad \text{………(4.1)}$$

Where, x = time (year), m = slope coefficient and c = least-squares estimate of the intercept

The slope coefficient indicates the annual average rate of change in the hydrologic characteristic. If the slope is significantly different from zero statistically, it is entirely reasonable to interpret that there is a real change occurring over time. The sign of slope defines the direction of the trend of the variable: positive sign indicates a rising trend while negative sign indicates a falling trend.

4.1.2 Sen's Estimator of Slope (Non-Parametric method)

The magnitude of trend in a time series was determined using a non-parametric method known as Sen's estimator (**Sen 1968**). This method assumes a linear trend in the time series and has been widely used for determining the magnitude of trend in hydro-meteorological time series (**Lettenmaier et al., 1994, Yue and Hashino, 2003, Partal and Kahya, 2006**). In this method, the slopes (T_i) of all data pairs are first calculated by

$$T_i = \frac{x_j - x_k}{j - k} \quad \text{for i = 1,2,…,N} \quad \text{……(4.2)}$$

where$x_j$and x_k are data values at time j and k ($j > k$) respectively. The median of these N values of T_i is Sen's estimator of slope which is calculated as

$$\beta = \begin{cases} T_{\frac{N+1}{2}} & \text{if } N \text{ is odd,} \\ \frac{1}{2}\left(T_{\frac{N}{2}} + T_{\frac{N+2}{2}}\right) & \text{if } N \text{ is even.} \end{cases} \qquad (4.3)$$

A positive value of β indicates an upwards (increasing) trend and a negative value indicates a downwards (decreasing) trend in the time series.

4.1.3 Mann–Kendall test (Non-parametric test)

Non-parametric trend technique can be adopted in case with the required data to be normally distributed and containing outlier in the data (**Helsel and Hirsch 1992; Birsan et al. 2005**). The Mann-Kendall test is a non-parametric rank based test for identifying trend in time series data. To ascertain the presence of a statistically significant trend in hydrologic climatic variables such as temperature, relative humidity, precipitation and stream flow with reference to climate change, the non-parametric Mann–Kendall (MK) test has been employed by a number of researchers (**Yu et al. 1993; Douglas et al. 2000; Burn et al. 2004**). The MK method searches for a trend in a time series without specifying whether the trend is linear or non-linear. The MK test was also applied in the present study. The MK test checks the null hypothesis H_0 of no trend versus the alternative hypothesis H_1 of the existence of an increasing or decreasing trend. The statistic S is defined as (**Salas 1993**):

$$S = \sum_{i=1}^{N-1} \sum_{j=i+1}^{N} \text{sgn}(x_j - x_i) \qquad (4.4)$$

Where$N$ is the number of data points. Assuming $(x_j - x_i) = \theta$, the value of sgn(θ) is computed as follows

$$\text{sgn}(\theta) = \begin{cases} 1 & if \quad \theta > 0, \\ 0 & if \quad \theta = 0, \\ -1 & if \quad \theta < 0. \end{cases} \qquad (4.5)$$

This statistic represents the number of positive differences minus the number of negative differences for all the differences considered. For large samples (N> 10), the test is conducted using a normal distribution (**Helsel and Hirsch, 1992**) with the mean and the variance as follows:

$$E[S] = 0 \qquad (4.6)$$

$$Var(S) = \frac{N(N-1)(2N+5) - \sum_{k=1}^{n} t_k (t_k - 1)(2t_k + 5)}{18} \qquad (4.7)$$

Where,n is the number of tied (zero difference between compared values) groups and $t_k$is the number of data points in the kth tied group. The standard normal deviate (Z-statistics) is then computed as (Hirsch *et al.* 1993):

$$Z = \begin{cases} \frac{S-1}{\sqrt{Var(S)}} & if \quad S > 0 \\ 0 & if \quad S = 0 \\ \frac{S+1}{\sqrt{Var(S)}} & if \quad S < 0. \end{cases} \tag{4.8}$$

If the computed value of $|Z| > z_{\alpha/2}$, the null hypothesis H_0 is rejected at the α level of significance in a two-sided test. In this analysis, the null hypothesis was tested at 95% confidence level.

In our study, both trend analysis methods were used i.e. Parametric (Regression analysis) and non-parametric (Mann-kendall & sen slope) and monthly precipitation records of four rain gauge stations (Ambalavayal, Kalladi, Manjeri, Nilambur) of Chaliyar river basin, kerala collected for the period 1993 to 2012 have been used for analysis of rainfall trend on seasonal and annual scale.

4.2 DROUGHT CHARACTERIZATION

Drought indices are employed to characterize drought and its statistical properties. Drought analysis from stochastic point of view provides information required for the subsequent risk analysis (probabibilities of drought occurrence and drought impacts). Drought indices provide spatial and temporal representations of historical droughts and therefore place current conditions in historical perspective.

The Standardized Precipitation Index (SPI) is a tool which was developed primarily for defining and monitoring drought. It allows an analyst to determine the rarity of a drought at a given time scale (temporal resolution) of interest for any rainfall station with historic data. It can also be used to determine periods of anomalously wet events. This index has gained importance in recent years as a potential drought indicator.

It was developed by T.B. Mckee, N.J. Doesken and J. Kleist, Colorado State University, 1993.

The Standardized Precipitation Index (SPI), which is one of the most commonly used and recommended drought indicators, is used to describe meteorological droughts. Precipitation is the only input parameter.

The SPI was designed to enumerate precipitation deficit for multiple timescales. Generally, the SPI is calculated for 3, 6, 12, 24, and 48 month time scales. These timescales reflect the impact of drought on the availability of the different water resources.

1-month SPI is very similar to a map displaying the percent of normal precipitation for a month. It reflects relatively short-term conditions, its application can be related closely with short-term soil moisture and crop stress, especially during the growing season.

A 3-month SPI reflects short- and medium-term moisture conditions and provides a seasonal estimation of precipitation, also a 3-month SPI might be more applicable in highlighting available moisture conditions.

6-month SPI indicates medium-term trends in precipitation & it can be very effective in showing the precipitation over distinct seasons. Information from a 6-month SPI may also begin to be associated with anomalous stream flows and reservoir levels.

9-month SPI provides an indication of precipitation patterns over a medium time scale.

12-month SPI reflects long-term precipitation patterns & 12-month SPI are probably tied to streamflows, reservoir levels, and even groundwater levels at the longer time scales.

Negative & Positive values of SPI indicate dry and wet periods, respectively. SPI values between -1 & -2 refer to moderate to severe droughts while SPI below -2 indicates extreme droughts.

The SPI (McKee et al, 1993) is a powerful, flexible index that is simple to estimate. The probability of observed precipitation is converted in to an index. Many drought planners appreciate the SPI's versatility. It is also used by a variety of research institution, Universities, and National Meteorological and Hydrological services across the world as part of drought monitoring and early warning efforts. In fact precipitation is the only required input parameters. In addition, it is just as effective in analysing wet periods/cycles as it is analysing dry periods/ cycles.

The SPI estimation for any location is based on the long term precipitation record for the desired period. This long term record is fitted to a probability distribution, which is then transformed in to a normal distribution so that the mean SPI for the location and the desired period is zero (Edwards and McKee, 1997).Positive SPI values designate greater than median precipitation while negative values designate less than median precipitation. Because the SPI is normalized, wetter and drier climates can be represented in the same way; thus, wet periods can also be monitored using the SPI.

(Mc Kee et al, 1993) used the classification system shown in the value table 4.1, to define drought intensities resulting from SPI.

Table 4.1: Standard Ranges of SPI values and their classification

Category	SPI
2.0+	Extremely Wet
1.5 to 1.99	Severely Drought
1.0 to 1.49	Moderately Wet
0 to 0.99	Lightly Wet
0 to - .99	Lightly Drought
-1.0 to -1.49	Moderately Drought
-1.5 to -1.99	Severely Drought
-2 and less	Extremely Drought

4.2.1 SPI Program Algorithm

In our study, we have used a program to calculate SPI on different time scales, whose brief algorithm is as follows:

Input File

• Prepare an input file with all your monthly data for one station in the following format:

```
Header
yyyy mm pppp
yyyy mm pppp
yyyy mm pppp
```

Where,

Header = a string which describes the file, or something about the station, etc;

yyyy = year; mm = month (in digit format 1,2,3 etc);

pppp = precipitation multiplied by 100

First of all arrange your monthly precipitation data in the given format in excel file. After arranging it in excel file, save it as text file i.e. in .txt format.

Name this file in .txt file and put it in the same directory as the SPI program.

Execute Program

• Double click SPI_SL_6.exe to execute the program

• Enter the number of SPI monthly intervals (up to six at one time) you wish to run

• Enter the SPI monthly intervals (i.e. 1, 3, 4, 6, 12 month, 24 month)

• Enter the input and output file names

Output File

• The output of the program is in the following format:

```
Header
yyyy mm spi1 spi3 spi6 spi12
yyyy mm spi1 spi3 spi6 spi12
yyyy mm spi1 spi3 spi6 spi12
```

Where

yyyy and mm are as before

spi1 = SPI for a 1 month rainfall total

spi3 = SPI for a 3 month rainfall total

spi6 = SPI for a 6 month rainfall total

spi12 = SPI for a 12 month rainfall total

Name this file in .txt file and the result will appear in the same folder.

Now save this file as excel file, arrange the data, draw the graphs on different time scales & interpret the results.

In the present study, monthly precipitation records of four rain gauge station for the period 1993 to 2012 in Chaliyar basin were used to estimate SPI values for 3, 6 & 12 months time scale.

4.3 DIGITAL ELEVATION MODEL & WATERSHED DELINEATION

For modelling purposes, a watershed may be partitioned into a number of sub-watersheds or sub basins. To delineate study area, for the Chaliyar river basin in Kerala, a 90-m digital elevation model (DEM) was used. By partitioning the watershed from the entire basin, the user is able to reference different areas from one another spatially.

Shuttle Radar Topographic Mission (SRTM) was use to prepare the digital elevation model (DEM) at 90-m resolution. DEM is used to derive slope, aspect, flow direction and accumulation, and stream network information (fig.4.3) also. For our study, Digital elevation model was acquired from earthexplorer.usgs.gov site and used to delineate the watershed.

To delineate the watershed, initially the reprojection of SRTM DEM (90m) in ARCGIS was done i.e. the SRTM DEM (90m) was reprojected to UTM spatial reference. Salient features regarding this are as follows:

Projection System-UTM, Zone 43 N

Spheroid-WGS84

Datum-WGS84

After this processing of DEM took place in ARCGIS for watershed delineation.

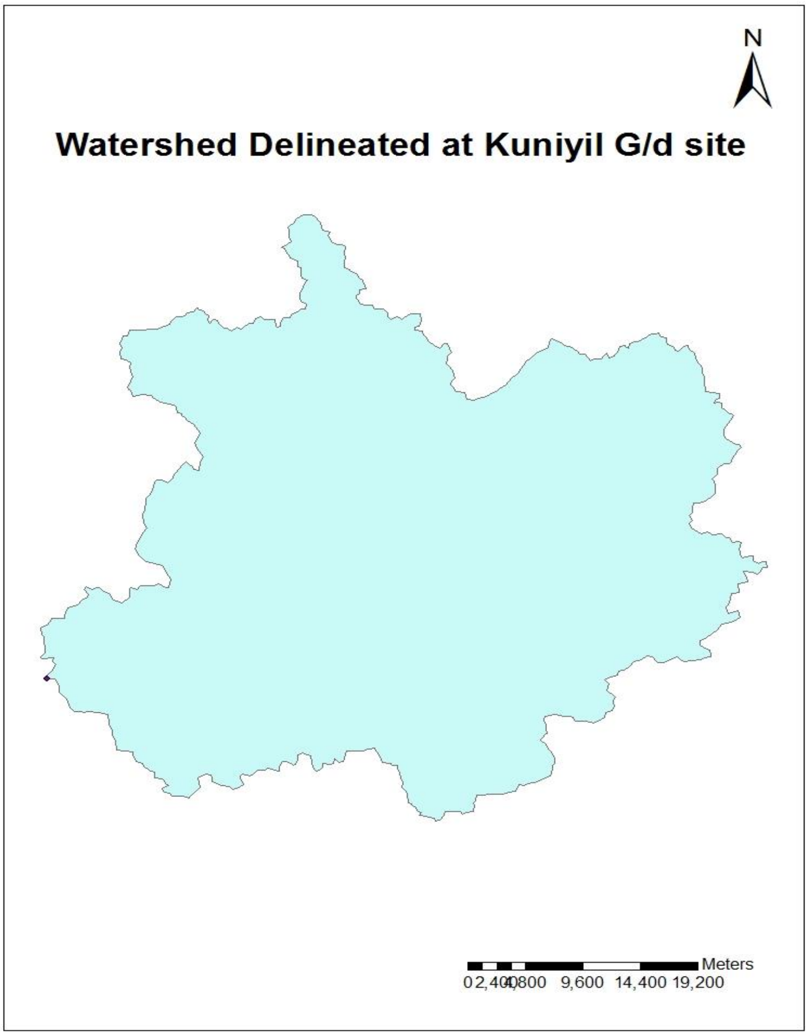

Figure 4.1 our study area for Hydrological Modelling delineated with the help of DEM & Arc GIS at Kuniyil Gauge discharge site

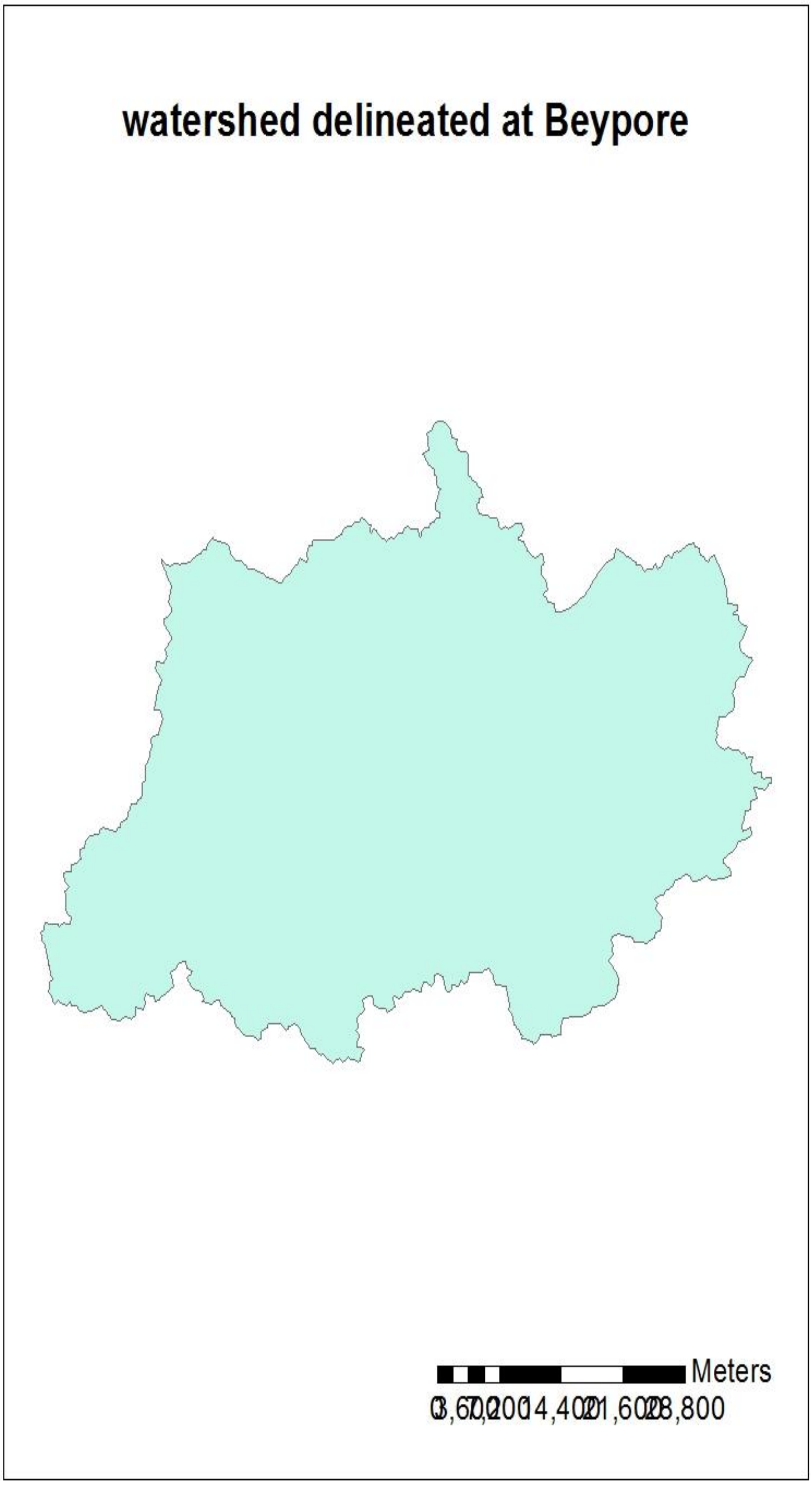

Figure 4.2 Map of Study area delineated from Chaliyar river basin with the help of DEM & Arc GIS at Beypore

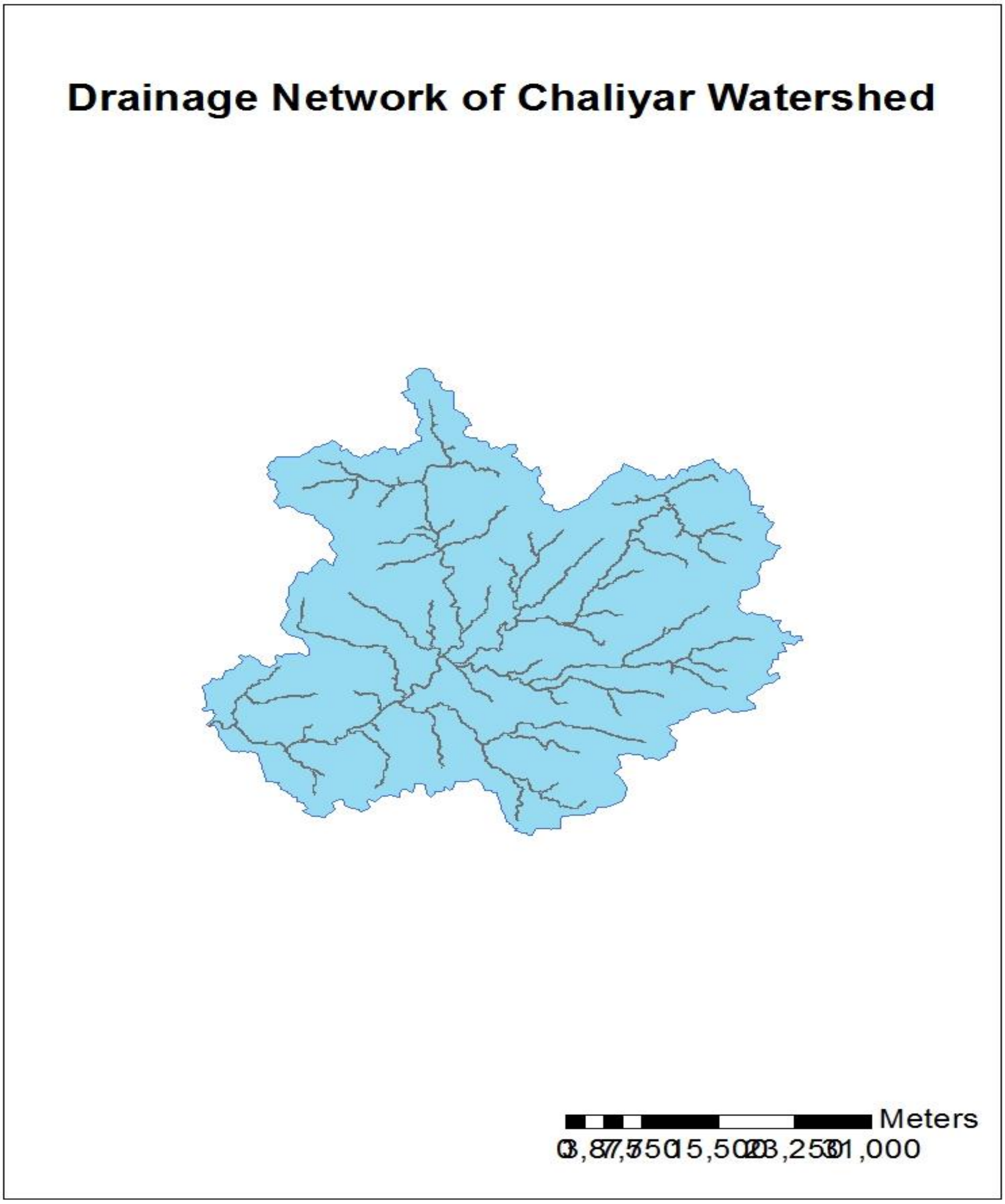

Figure 4.3 Drainage of Chaliyar Network prepared with the help of DEM

4.4 SPATIAL VARIATION OF GROUND WATER LEVELS IN GIS ENVIRONMENT

4.4.1 Introduction

Groundwater is one of the major sources of fresh water. One-third of world's total population depends primarily on groundwater for drinking, domestic, industrial, and agricultural purposes .The rate of extraction of groundwater is continuously increasing with the escalation of the population, agriculture and industrial development. This excessive groundwater extraction is responsible for serious groundwater level declination in many areas. The declination of groundwater level has adverse effects on water pollution, salinity intrusion, groundwater contaminant, and imbalance of ecosystem, etc. Therefore, a regular monitoring of groundwater levels is quite important to cope with the current and future water demand. However, remotely placed observation wells do not provide detailed information about spatial coverage of water level. Ground water level fluctuation space to space (Spatial) and time to time (Temporal) is a major problem in India and the assessment of spatiotemporal characteristics water level fluctuation trend is very important in the point of view of future development.

Recently, spatial interpolation mechanism through geostatistical analysis has become very popular to analyse such spatial characteristics of the groundwater level. Geostatistical analysis has been widely used in hydrology to determine various important properties of the aquifer. Besides, geostatistics-based spatiotemporal variability analysis is used successfully in many other fields. Geostatistical analysis has been theoretically defined and applied by many researchers (Isaaks and Srivastava, 1989; Goovaerts, 1997; Kitanidis, 1997).

4.4.2 Kriging Method

The Kriging method is an example of a group of geo-statistical techniques used to interpolate the value of a random field. Matheron (1971) named and formalized this method in honour of Daniel G. Krige, a South African mining engineer who pioneered the field of geo-statistics. The Kriging method is based on statistical models involving autocorrelation. Autocorrelation refers to the statistical relationships between measured points.

The geo-statistical methods have the capability of producing a prediction surface and provide some measures of the certainty and accuracy of the predictions. In this method, the value of the variable is estimated for a particular point using a weighted sum of the available point observations. The weights of the data are chosen so that the interpolation is unbiased and the variance is minimized. In general, the Kriging system must be Linear, Authorized, Unbiased and Optimal (LAUO). Kriging is the first method of interpolation to take into account the spatial dependence structure of the data. There are several types of Kriging, which differ according to the form applied to the mean of the interest variable: (a) when it is assumed that the mean is constant and known, simple Kriging (SK) is applied; (b) where the mean is constant but unknown, ordinary Kriging (ORK) is applied; (c) where the mean is assumed to show a polynomial function of spatial coordinates, universal Kriging (UNK).

4.4.3 Methodology

In the study area, 35 observation wells were established to monitor variation in ground water levels in Chaliyar River Basin. The GPS was used to locate exact coordinates of the sample collection locations for continuous monitoring purposes. Only Groundwater samples collected from 29 observation wells were used for study, as remaining observation wells were showing discontinuity in their data.

The data was collected from these 35 observation wells fairly evenly distributed over the study area. i.e. Chaliyar basin in the format of pre-monsoon ground water level & post-monsoon ground water level data for the period of 1996 to 2014 . The groundwater levels measured on seasonal basis i.e. pre- monsoon & post-monsoon ground water level data were monitored for groundwater fluctuations of the year.

Details regarding the well location, dimensions, depth to static water level, etc. were collected and tabulated for each well. The location of the observation wells were then transferred on to the base map in a GIS environment. The locations of the well with their districts and Coordinates are shown in fig.4.3.

Then the ground water level data (pre monsoon, post monsoon & fluctuations) and the spatial data (coordinates of observatory wells) were joined in Arc GIS software. After linking the spatial and non-spatial data, the groundwater quality point layer was generated for further analysis. Later on the other analysis was carried out using the Kriging Technique in the Arc GIS 9.3 software.

Interpolation creates a continuous (or prediction) surface from sampled point values. It makes predictions from sample measurements for all locations in a raster dataset whether or not a measurement has been taken at the location.

In the present data analysis, we use Ordinary Kriging Interpolation technique to prepare ground water fluctuation map for our study area at different time intervals & further to study the spatial variation of ground water levels in GIS environment.

In brief methodology adopted in our study is given as follows:

1. Arrange the data in excel.
2. Review the GWL table in Excel.
3. Add table to Arc Map.
4. Interpolate and prepare the ground water level fluctuation maps using ordinary Kriging interpolation technique in Arc GIS.
5. Study & interpret the spatial Ground water fluctuations for Chaliyar river basin.

Figure 4.4: Locations of different wells in our study area (Chaliyar River basin), denoted by the name of their districts in which they are falling

Table 4.2: Locations of different wells used to study the ground water fluctuations in our study area

District	Station	Longitude	Latitude
Kozhikode	Mavoor-1	75.955	11.257
Kozhikode	Beypore	75.806	11.171
Kozhikode	Ramanattukara1	75.858	11.175
Kozhikode	Koduvally	75.909	11.358
Kozhikode	Tamarasseri	75.938	11.421
Kozhikode	Puduppadi	75.984	11.49
Kozhikode	Chelavur	75.882	11.299
Kozhikode	Mavoor-2	75.942	11.25
Kozhikode	Malayamma	75.933	11.35
Mallapuram	Trippanachi	76.044	11.158
Mallapuram	Kalikavu	76.32	11.169
Mallapuram	Buliyampadam	76.25	11.41
Mallapuram	Karulai	76.327	11.283
Mallapuram	Kadalundi	75.827	11.136
Mallapuram	Areakode	76.041	11.233
Mallapuram	Kavanur	76.055	11.196
Mallapuram	Manjeri	76.12	11.118
Mallapuram	Karulayi	76.297	11.281
Mallapuram	Chokkad	76.33	11.233
Mallapuram	Olavattur	75.95	11.21
Mallapuram	Edavanna1	76.15	11.208
Mallapuram	Thiruvali	76.178	11.19
Mallapuram	Wandoor	76.233	11.2
Mallapuram	Edakkara	76.293	11.373
Mallapuram	Kondotty	75.964	11.146
Mallapuram	Nilambur	76.241	11.283
Mallapuram	Maruda	76.325	11.416
WAYANAD	Vaduvanchal	76.225	11.555
WAYANAD	Ambalavayal	76.229	11.606

4.5 Rainfall-runoff Modelling

A model is a simplified representation of a real world system, and consists of a set of simultaneous equations or a logical set of operations contained within a computer program. Models have parameters which are numerical measures of a property or characteristics that are constant under specified conditions.

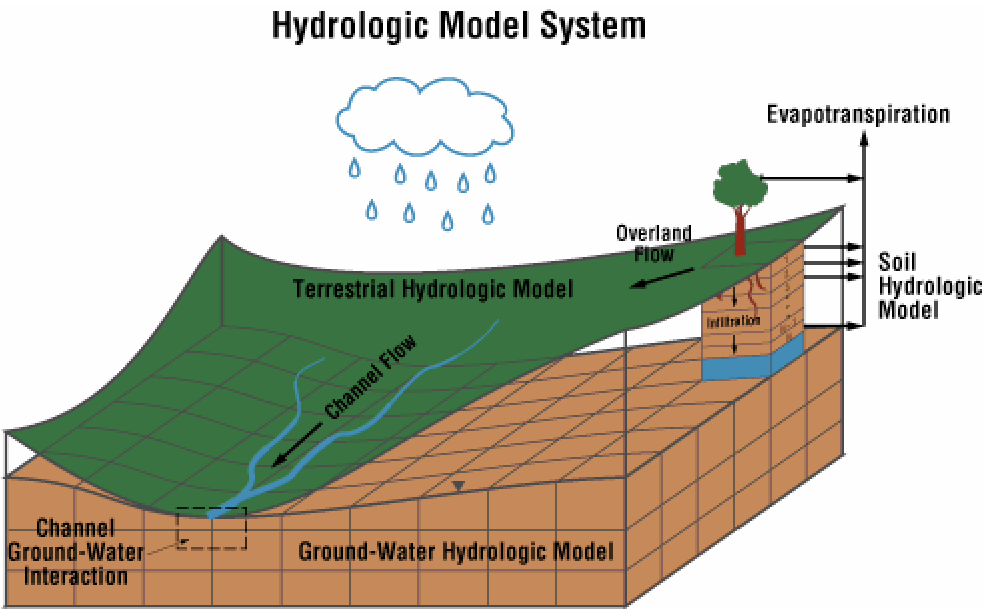

Fig. 4.5: Representing hydrological cycle

A distributed model is one in which parameters, inputs and outputs vary spatially. A semi-distributed model may adopt a lumped representation for individual sub-catchments. A model is deterministic if a set of input values will always produce exactly the same output values, and stochastic if, because of random components, a set of input values need not produce the same output values. An event-based model produces output only for specific time periods, whereas a continuous model produces continuous output. The tasks for which rainfall-runoff models are used are diverse, and the scale of applications ranges from small catchments, of the order of a few hectares, to that of global models.

4.5.1 Importance of Rainfall-runoff modelling

Since long period rainfall-runoff modeling technique is being used to deal with practical problems like water resources assessment, design of engineering channels, flood forecasting, predicting population incidents, and many more purposes. Now a day's water resources planning and management are based on the application of suitable rainfall runoff model.

Typical tasks which can be carried out with hydrological models are as follow:

1. Modeling existing catchments for which input-output data exist,

- e.g. Extension of data series for flood design of water resource evaluation, operational flood forecasting, or water resource management

2. Runoff estimation on un-gauged basins
3. Prediction of effects of catchment change

- e.g. Land use change, climate change

4. Coupled hydrology and geochemistry

- e.g. Nutrients, Acid rain

5. Coupled hydrology and meteorology

- e.g. Global Climate Models

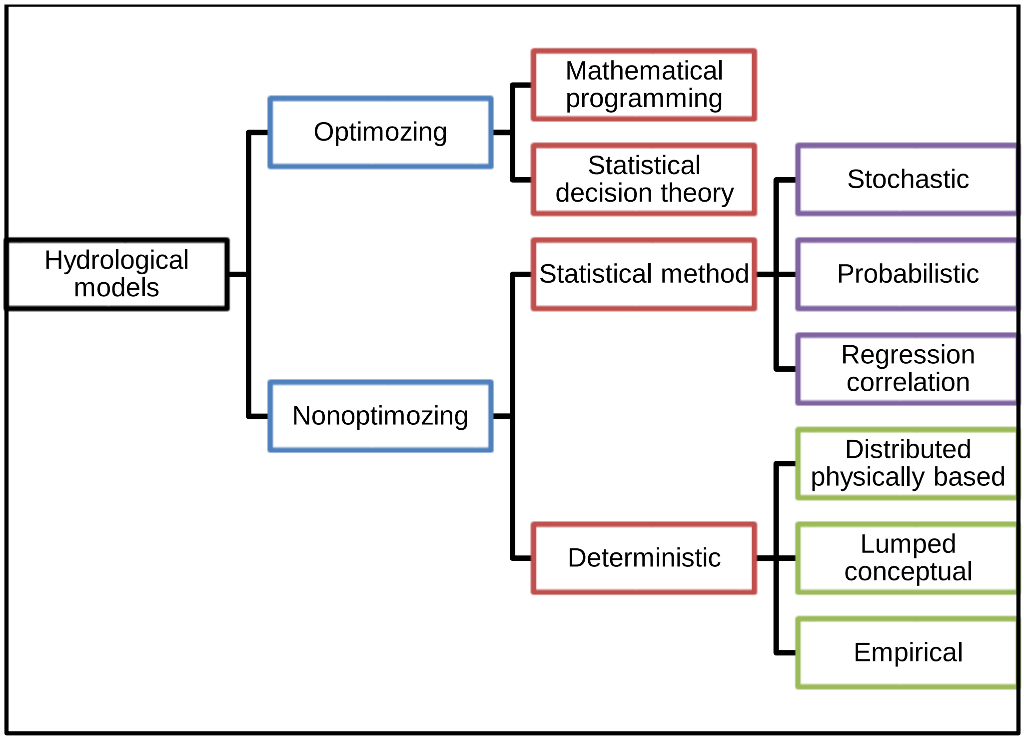

Figure 4.6: Classification of hydrologic models

4.5.2 Stream flow Modelling using Mike 11 NAM model

Many hydrological deterministic models have been developed to simulate the rainfall-runoff process for river watersheds, but most have complicated structures and need various observed data for calibration. The Mike 11 NAM model has been widely used in many Asian countries not only because of its simple structure but also because of its fewer data requirements (Ngoc et al, 2011). However, this hydrological model still needs extensive time and effort to calibrate various model parameters.

MIKE 11 NAM is professional engineering software developed by Danish Hydraulic Institute, Denmark. NAM can be prepared in a number of different modes depending on the requirement. As default, NAM is prepared with 9 parameters, representing surface zone, root zone and ground water storage. Specifically the MIKE 11 software is meant for simulation of flows, water quality and sediment transport in river, irrigation systems, channels and other water bodies. The basic data input requirements for the MIKE11 NAM model are meteorological data and discharge data for model calibration, definition of the catchment parameters, and definition of initial conditions.

A conceptual model like NAM is based on physical structures and equations used together with semi-empirical ones. It is an imitation of the land phase of the hydrological cycle. NAM simulates the rainfall-runoff process by continuously accounting for the water content in four different and mutually interrelated storages that represent different physical elements of the catchment as shown in Fig. 4.6. These mutually interrelated storage include snow storage, surface storage, lower or root zone storage and ground water storage. In addition NAM allows treatment of man-made interventions in the hydrological cycle such as irrigation and groundwater pumping. The basic meteorological data requirements are precipitation time series and potential evapotranspiration time series. On this basis, the model produces a time series of catchment runoff, a time series of subsurface flow contributions to the channel, and information about other elements of the land phase of the hydrological cycle, such as soil moisture content and groundwater recharge.

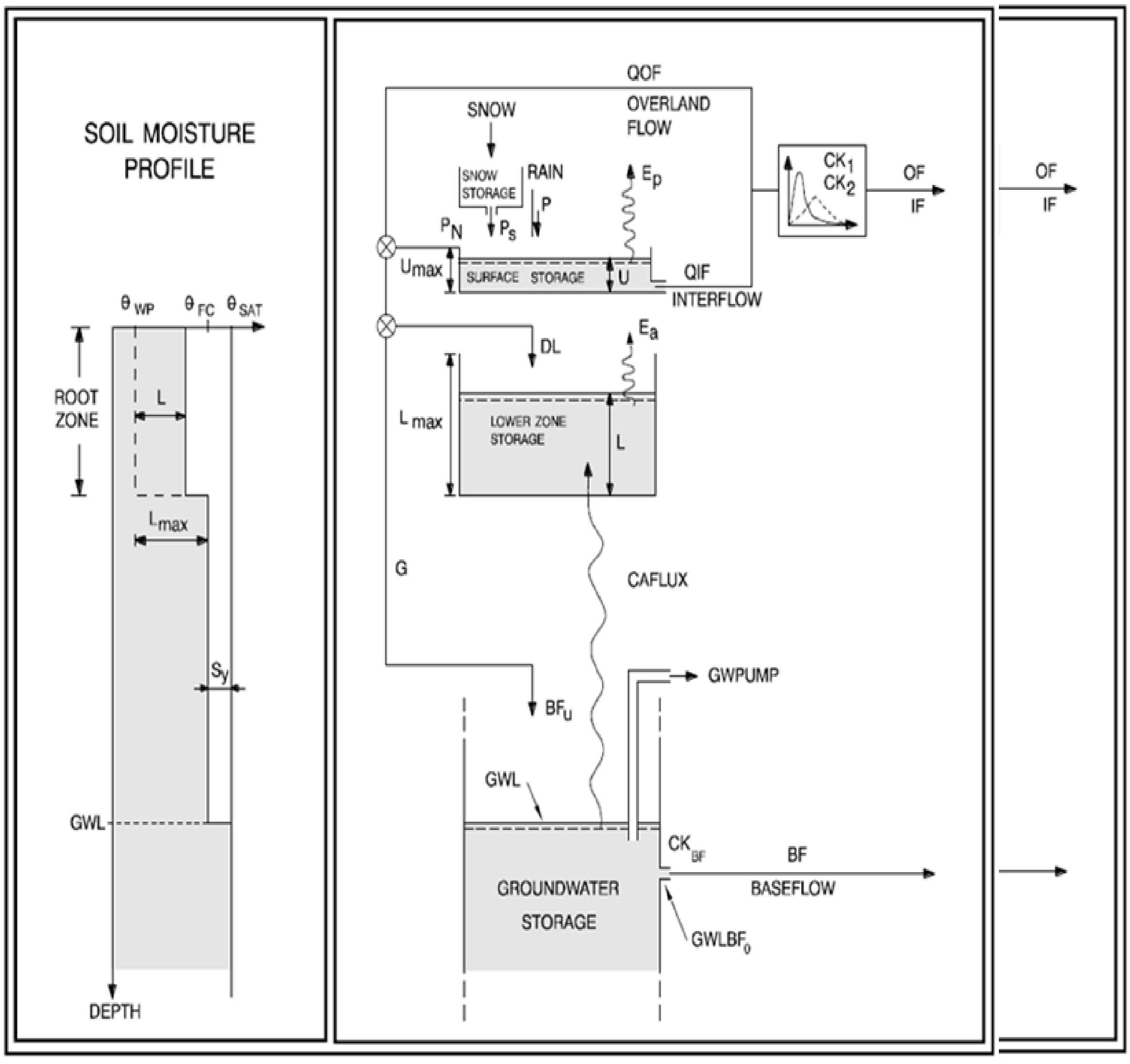

Fig. 4.7: Structure of NAM model for rainfall-runoff simulation

4.5.3 Data requirements

- Data requirement for the MIKE 11 NAM model consists of:
  1) Model parameter
  2) Initial conditions
  3) Meteorological data
  4) Stream flow data for model calibration and validation
- The basic meteorological data requirements are:
  1) Rainfall
  2) Potential evapotranspiration

4.5.4 Parameters of NAM model

NAM is developed with nine parameters representing surface zone, root zone and ground water storage. The function of each parameter in the model, their theories and assumption are discussed in detail as below:

Surface storage

Moisture intercepted on the vegetation as well as water trapped in depressions and in the uppermost, cultivated part of the ground is represented as surface storage. U_{max} denotes the upper limit of the amount of water in the surface storage. The amount of water, U, in the surface storage is continuously diminished by evaporative consumption as well as by horizontal (interflow).When there is maximum surface storage, some of the excess water, P_N, will enter the streams as overland flow, whereas the remainder is diverted as infiltration into the lower zone and groundwater storage.

Lower zone or root zone storage

The soil moisture in the root zone, a soil layer below the surface from which the vegetation can draw water for transpiration, is represented as lower zone storage. L_{max} denotes the upper limit of the amount of water in this storage. Moisture in the lower zone storage is subjected to consumptive loss from transpiration. The moisture content controls the amount of water that enters the groundwater storage as recharge and the interflow and overland flow components.

Evapotranspiration

Evapotranspiration demands are first met at the potential rate from the surface storage. If the moisture content U in the surface storage is less than these requirements ($U<Ep$), the remaining fraction is assumed to be withdrawn by root activity from the lower zone storage at an actual rate Ea. Ea is proportional to the potential evapotranspiration and varies linearly with the relative soil moisture content, L/L_{max}, of the lower zone storage.

$$E_a = \left(E_p - U\right)\frac{L}{L_{max}} \qquad \text{… … … (4.9)}$$

Overland flow

When the surface storage spills, i.e. when $U > U_{max}$, the excess water P_N give rise to overland flow as well as to infiltration. Q_{OF} denotes the part of P_N that contributes to overland flow. It is assumed to be proportional to P_N and to vary linearly with the relative soil moisture content, L/L_{max}, of the lower zone storage.

$$Q_{OF} = \begin{cases} C_{QOF}\dfrac{L/L_{max} - T_{OF}}{1 - T_{OF}}P_N & for\ L/L_{max} > T_{OF} \\ 0 & for\ L/L_{max} \le T_{OF} \end{cases} \qquad \text{… … … (4.10)}$$

Where,

C_{QOF} is the overland flow runoff coefficient ($0 < C_{QOF} < 1$),

T_{OF} is the threshold value for overland flow ($0 < T_{OF} < 1$).

The proportion of the excess water P_N that does not run off as overland flow infiltrates into the lower zone storage. A portion,ΔL of the water available for infiltration, (P_N -Q_{OF}), is assumed to increase the moisture content L in the lower zone storage. The remaining amount of infiltrating moisture, G, is assumed to percolate deeper and recharge the groundwater storage.

Interflow

The interflow contribution, Q_{IF}, is assumed to be proportional to U and to vary linearly with the relative moisture content of the lower zone storage.

$$Q_{IF} = \begin{cases} (C_{KIF})^{-1} \dfrac{L/L_{max} - T_{IF}}{1 - T_{IF}} \, U & for\ L/L_{max} > T_{IF} \\ 0 & for\ L/L_{max} \leq T_{IF} \end{cases} \quad \ldots\ldots\ldots (4.11)$$

Where,

C_{KIF} is the time constant for interflow,

T_{IF} is the root zone threshold value for interflow ($0 < T_{IF} < 1$).

Interflow and overland flow

The interflow is routed through two linear reservoirs in series with the same time constant C_{K1K2}. The overland flow routing is also based on the linear reservoir concept but with a variable time constant.

$$CK = \begin{cases} C_{K_1K_2} & for\ OF < OF_{min} \\ C_{K_1K_2} \left(\dfrac{OF}{OF_{min}}\right)^{-b} & for\ OF \geq OF_{min} \end{cases} \quad \ldots\ldots\ldots (4.12)$$

Where,

OF is the overland flow (mm/hour),

OF_{min} is the upper limit for linear routing (= 0.4 mm/hour),

The constant b = 0.4 corresponds to using the Manning formula for modeling the overland flow. The Equation above ensures in practice that the routing of real surface flow is kinematic, while subsurface flow being interpreted by NAM as overland flow (in catchments with no real surface flow component) is routed as a linear reservoir.

Groundwater recharge

The amount of infiltrating water G recharging the groundwater storage depends on the soil moisture content in the root zone.

$$G = \begin{cases} (P_N - Q_{OF}) \dfrac{L/L_{max} - TG}{1 - TG} & for\ L/L_{max} > TG \\ 0 & for\ L/L_{max} \leq TG \end{cases} \quad \ldots\ldots\ldots (4.13)$$

Where, TG is the root zone threshold value for groundwater recharge ($0 < TG < 1$).

Soil moisture content

The lower zone storage represents the water content within the root zone. After apportioning the net rainfall between overland flow and infiltration to groundwater, the remainder of the net rainfall increases the moisture content L within the lower zone storage by the amount ΔL.

$$\Delta L = P_N - Q_{OF} - G \qquad \dots\dots\dots(4.14)$$

Base flow

The base flow *BF* from the groundwater storage is calculated as the outflow from a linear reservoir with time constant C_{KBF}. Description of the parameters and their effects is presented in Table 4.3.

Table 4.3: Different parameters of the NAM model

Parameter	Unit	Description	Effects
Umax	Mm	Maximum water content in surface storage	Overland flow, infiltration, evapotranspiration, Interflow
Lmax	Mm	Maximum water content in lower zone/root storage	Overland flow, infiltration, evapotranspiration, base Flow
CQOF		Overland flow coefficient	Volume of overland flow and infiltration
CKIF	Hrs	Interflow drainage constant	Drainage of surface storage as interflow
TOF		Overland flow threshold	Soil moisture demand that must be satisfied for overland flow to occur
TIF		Interflow threshold	Soil moisture demand that must be satisfied for interflow to occur
TG		Groundwater recharge threshold	Soil moisture demand that must be satisfied for groundwater recharge to occur
CK1	Hrs	Timing constant for overland Flow	Routing overland flow along catchment slopes and Channels
CK2	Hrs	Timing constant for interflow	Routing interflow along catchment slopes
CKBF	Hrs	Timing constant for base flow	Routing recharge through linear groundwater Recharge

4.5.5 Data Inputs

4.5.5.1 Rainfall data

The areal precipitation or the representative precipitation over a defined area was required to run the MIKE 11 NAM model. In present study the point precipitation at Ambalavayal, Nilambur, and Manjeri was obtained from the CWRDM, Kerala. The areal precipitation was computed from point precipitation by Thiessen polygon method. The Thiessen polygon of the study area was generated with the help of Arc GIS 9.3 software. Among three rain gauge stations, Nilambur was found to be the most influencing station covering maximum area. The weightages denoting the degree of influence of the individual rain gauge station over the study area were estimated. The percentage weights were obtained by considering the proportion of its representative area and given in Table 4.4. Thiessen polygon map of the study area has been shown in Fig. 4.7.

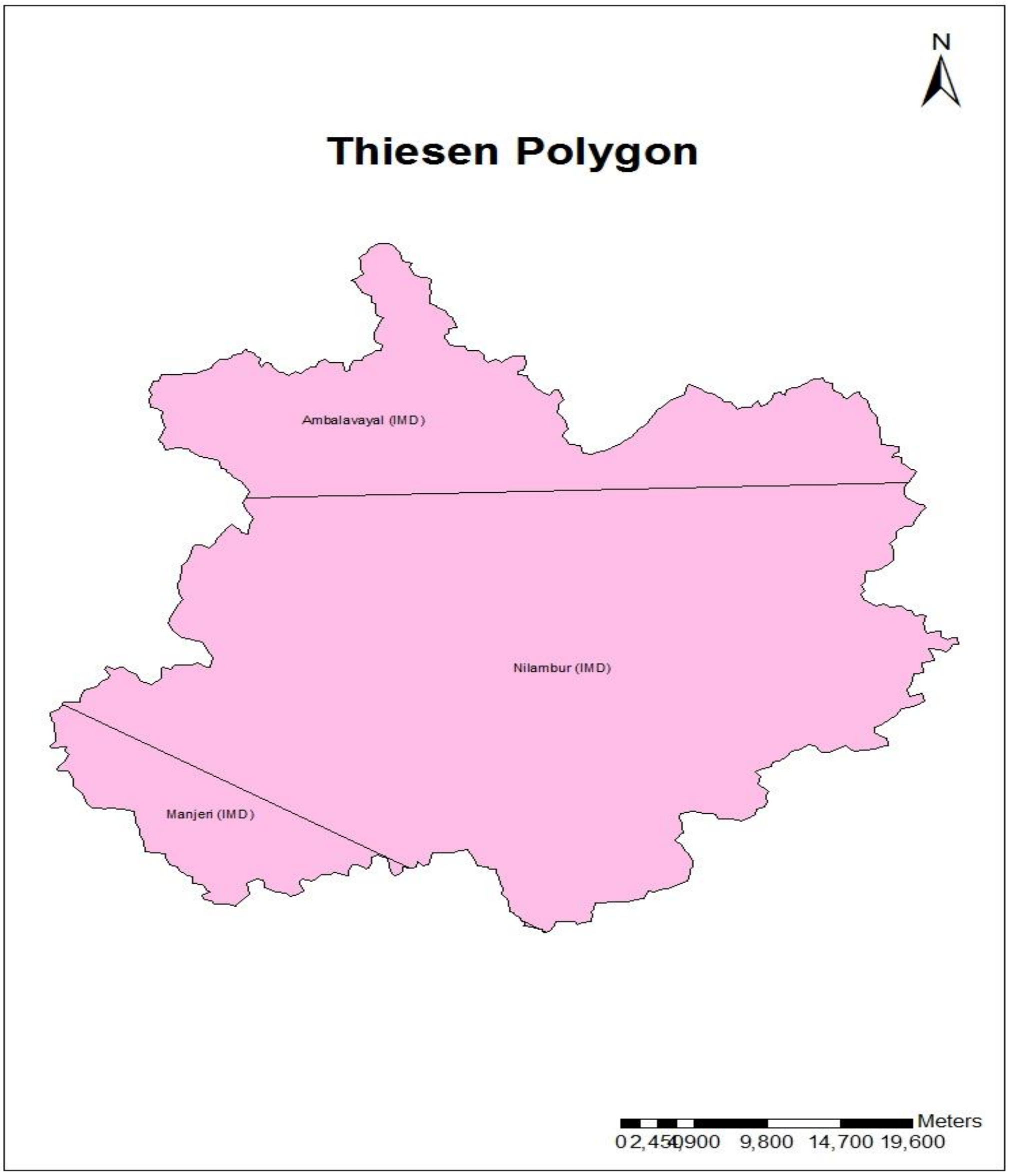

Fig. 4.8: Thiessen polygon of the study area

Table 4.4: Thiessen weightages for Raingauge stations

Station	Raingauge Station	Weight Area (%)
1	Ambalavayal	23%
2	Nilambur	70%
3	Manjeri	7%

From the above table, we can conclude that the weight area for Nilambur is quite larger as compared to other two stations, hence Rainfall data only for the Nilambur station is used in modelling.

The rainfall data was the major input for the rainfall-runoff modeling. The reliability of rainfall data was tested by plotting the rainfall against the runoff, if the straight line graph occurs it indicates that data is consistent and non-straight line graph indicated that the data have been subjected to various change. In the present study, when annual rainfall was plotted against the annual runoff for the period from 2001 to 2005 which is shown in Fig. 4.8, a straight line graph was observed, it shows the linear relation between rainfall and runoff and concluded that data was consistent to be used in rainfall-runoff modeling. The correlation coefficient of the linear regression has been obtained as 0.8963, showing good correlation between rainfall and runoff.

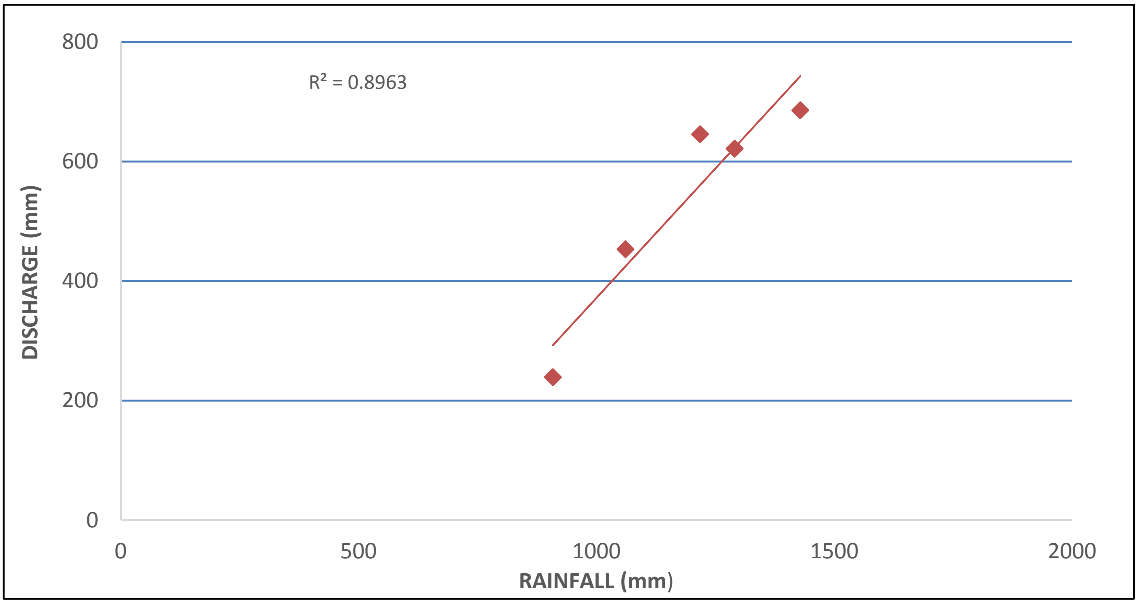

Fig. 4.9: Graph showing linear relation between rainfall-runoff

4.5.5.2 Discharge Data

Observed discharge data at the catchment outlet are required for comparison with the simulated runoff for model calibration and validation. In the present study, daily discharge data of chaliyar basin in meter cube per second (m^3/s) obtained from the Kuniyil gauging site for a period ranging from 2001 to 2005 were used and the discharge was treated as instantaneous.

Before using the rainfall-runoff data for the modeling, it is necessary to check the consistency of data. For checking the consistency of rainfall and runoff, coefficient of runoff should be calculated. Runoff coefficient is the ratio between runoff and rainfall. It is dimensionless term, varies from 0 to 1 depending on land use and soil types.

4.5.5.3 Potential evapotranspiration

Potential evapotranspiration (ET_0) is the amount of water that could be evaporated and transpired if there were sufficient water available. This demand incorporates the energy available for evaporation and the ability of the lower atmosphere to transport evaporated moisture away from the land surface. ET_0 is higher in the summer, on less cloudy days, and closer to the equator, because of the higher levels of solar radiation that provides the energy for evaporation. ET_0 is also higher on windy days because the evaporated moisture can be quickly moved from the ground of plants, allowing more evaporation to fill its place. Calculation of potential evapotranspiration was important due to its high effect on runoff in the form of evaporation from the surface. More evaporation causes the additional loss of water from the basin which was the part of runoff.

In the present study daily ET_0 *has* been estimated using daily maximum temperature data, daily minimum temperature data & daily radiation data of Chaliyar river basin by employing Hargreaves method. The Hargreaves method is given as

$$PET = 0.0009384 \times Rext \times (Tavg + 17.8) \times \sqrt{(Tmax - Tmin)} \quad \ldots\ldots\ldots (4.15)$$

Where

PET = Potential evapotranspiration (mm/day)

Rext = Daily extra-terrestrial radiation (watt/m^2)

Tavg = Daily average temperature (°C)

Tmax = Daily maximum temperature (°C)

Tmin = Daily minimum temperature (°C)

The potential evapotranspiration was calculated for the periods from 2001 to 2005 using the parameters like maximum, minimum temperature, and solar radiation.

4.5.8 Objective function:

In general term, the objective of model calibration can be stated as below: selection of model parameters so that the model simulates the hydrological behaviour of the basin as closely as possible (Madsen, 2000). MIKE 11 NAM uses multi objective approach to answering the question. This means that several numerical performance measures are accounted in the optimization process including (1) a good agreement between the average simulated and observed basin runoff volume; (2) a good overall agreement between of the shape of the hydrograph; (3) a good agreement of the peak flow with respect to timing, rate and volume; and (4) a good agreement for low flows. In this study, first three objectives were preferred.

4.5.9 NAM model setup

In present study MIKE 11 NAM was setup to carry out rainfall-runoff modeling at the river gauging site Kuniyil in Chaliyar river basin.

The NAM model requires input information like daily rainfall, runoff and potential evapotranspiration data in *dfso* format to be assigned in the model. Thus all the input time series data required was converted in to *dfso* time series files using MIKE ZERO software. The best possible input data time series of five years period from 2001 to 2005 were used for the modeling. During the modeling the calibration was carried out for three years period from 2001 to 2003 and validation carried out for the remaining two years i.e. 2004 to 2005. After assigning the input information, the NAM model was run by using the auto-calibration mode for the time step of one day.

4.5.10 Model calibration

Calibration is a process of standardizing predicted values, using deviations from observed values for a particular area to derive correction factors that can be applied to generate predicted values that are consistent with the observed values. Such empirical corrections are common in modeling and it is understood that every hydrologic model should be tested against observed data, preferably from the watershed understudy, to understand the level of reliability of the model. (**Linsley1982**)

The process of model calibration is normally done either manually or by using computer-based automatic procedures. In manual calibration parameters are adjusted by trial and error method. In this case, the goodness-of-fit of the calibrated model is basically based on a visual judgment by comparing the simulated and the observed hydrographs. For an experienced hydrologist it is possible to obtain a very good and hydrologically sound model using manual calibration. However, since there is no generally accepted objective measure of comparison, and because of the subjective judgment involved, it is difficult to assess explicitly the confidence of the model simulations. Furthermore, manual calibration may be a very time consuming task, especially for an inexperienced hydrologist. In automatic calibration, parameters are adjusted automatically according to specified search scheme and numerical measures of the goodness-of-fit. As compared to manual calibration, automatic calibration is fast, and the confidence of the model simulations can be explicitly stated. The development of automatic calibration procedures has focused mainly on using a single overall objective function (e.g. the root mean square error between the observed and simulated runoff) to measure the goodness-of-fit of the calibrated model. Calibration based on a single performance measure, however, is often inadequate to measure properly the simulation of all the important characteristics of the system that are reflected in the observations.

In the present study, once the NAM model was setup, it was calibrated for three years period from 2001 to 2003. During calibration, the default model parameters were kept same and model was run in auto-calibration mode. The model then resulted in obtaining set of model parameters for the calibration period. The model output simulation results during calibration were checked for R^2 (coefficient of determination) value and graphically analysed for degree of agreement between simulated and observed runoff. The model parameters were again adjusted one by one using trial and error method to obtain best set of model parameters of the NAM model which could simulate runoff with high degree of agreement with observed runoff in term of timings, peaks and total volume. The model parameters thus obtained after refinement of model were then used in validation of the model.

4.5.11 Model validation

Model validation means judging the performance of the calibrated model over the portion of historical records which have not been used for the calibration. It tests the ability of model to estimate runoff for the periods other than those used for the calibration process. Thus the validation data must not be the same as those used for calibration but must represent a situation similar to that in which the model is to be applied operationally. In this study, the NAM model was then validated for the remaining period of two years from 2004 to 2005. During validation the set of model parameters obtained during the calibration was used and model was run without auto-calibration mode to simulate runoff. The statistics of the simulated results were analysed and output of the model were checked graphically to compare the simulated and observed runoff to verify the capability of calibrated model to simulate the runoff for the extended period of time.

4.5.12 Accuracy criteria

Accuracy is the characteristic of a measurement that indicates the degree to which the results of measurement, approach the true value of the measured quantity. The smaller the deviation of the result of measurement from the true value of the quantity that is the smaller the error the higher the measurement accuracy.

It can be calculated on the basis of following methods:

1. Coefficient of determination
2. Efficiency index

Coefficient of determination

Coefficient of determination was another method to evaluate the reliability of the model between the observed and simulated flow, which is given by equation:

$$R^2 = \frac{\sum_{i=1}^{n}(q_{obs} - \bar{q}_{obs})(q_{sim} - \bar{q}_{sim})}{\sqrt{[\sum_{i=1}^{n}(q_{obs} - \bar{q}_{obs})^2][\sum_{i=1}^{n}(q_{sim} - \bar{q}_{sim})^2]}} \quad \ldots \ldots \ldots (4.16)$$

Where,

q_{obs}= observed discharge

$\bar{q}_{obs}$= average observed discharge

q_{sim}= simulated discharge

$\bar{q}_{sim}$=average simulated discharge

Efficiency index

The reliability of the Mike 11 NAM was evaluated based on the Efficiency Index (EI) as described by the Nash and Sutcliffe (1970). Efficiency index depends on the error present in the model like missing data or inconsistency in the data. Efficiency Index is directly proportional to error’s present in the model (Input data). The efficiency index was calculated by using the following relationship:

$$EI = \frac{[\sum_{i=1}^{n}(q_o - \bar{q})^2 - \sum_{i=1}^{n}(q_o - q_s)^2]}{\sum_{i=1}^{n}(q_o - \bar{q})^2} \quad \ldots\ldots.\ldots\ldots 4.17$$

Where,

q_o= observed flow at time i number of data points,

$\bar{q}$= mean value of observed flow

$=\frac{1}{n}\sum_{i=1}^{n} q_o$

q_s= simulated flow at time i

n= number of data points.

The value of efficiency index lies between 0 to 1. The efficiency index equal to 1 indicates the best performance of the model.

CHAPTER 5

RESULTS & DISCUSSION

5.1 RAINFALL TREND ANALYSIS

The anomalies of rainfall and their trends were determined for all the stations considered in the study. The rainfall anomalies (deviation from mean) were then plotted against the time (in Year) and the linear trends observed in these have been represented graphically. Anomalies in seasonal and annual rainfall and their trends for the stations within the study area are shown graphically in Fig. 5.1 to 5.4. The magnitude of the seasonal and annual trend in the time series as determined using the Sen's estimator is given in Table.5.2

Annual Trend

From figure 5.1 to 5.4, it is indicated that, if we talk about results obtained from the parametric approach, the annual rainfall indicates rising trend in Ambalavayal and Kalladi station increasing at the rate of 18.72 mm/year & 57.18 mm/year respectively, while Manjeri & Nilambur indicates falling trend decreasing at the rate of 30.2 mm/year & 5.76 mm/year respectively.

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And if we talk about the result obtained from non parametric approach, no significant trend is observed in any of the station (as per Z statistic of Mann Kendall), but still precipitation increases at the rate of 19.5 mm/year & 54.33 mm/year in Ambalavayal & kalladi stations & decreases at the rate of 17.38 mm/year & 0.335 mm/year (as per sen's estimator).

Seasonal Trend

From table 5.1, Seasonal trend for all the four stations by parametric method (linear regression) is described by the following table in which the magnitude of regression slope or 'm' denotes the rate of increase or decrease of seasonal precipitation in mm/year while the sign of 'm' denotes the nature of trend i.e. falling or rising.

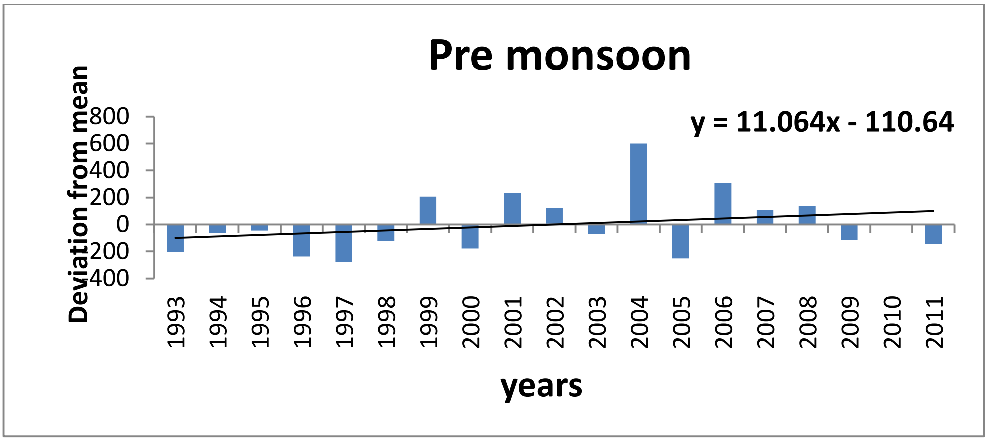
Whereas the seasonal trend for all the four stations by non parametric method (Mann Kendall & Sen's estimator) is described by the table 5.2 in which the value of Z statistic denotes the significance of trend i.e., whether the trend is significant or not. If the value of z does not lie within the range $1.96<z<1.96$ at 95% significance, then the trend is significant, else the trend is not significant & the magitude of Sen slope denote the rate of increase or decrease in seasonal or annual precipitation in the units of mm/year while the sign of sen slope denotes the nature of trend i.e. falling or rising.

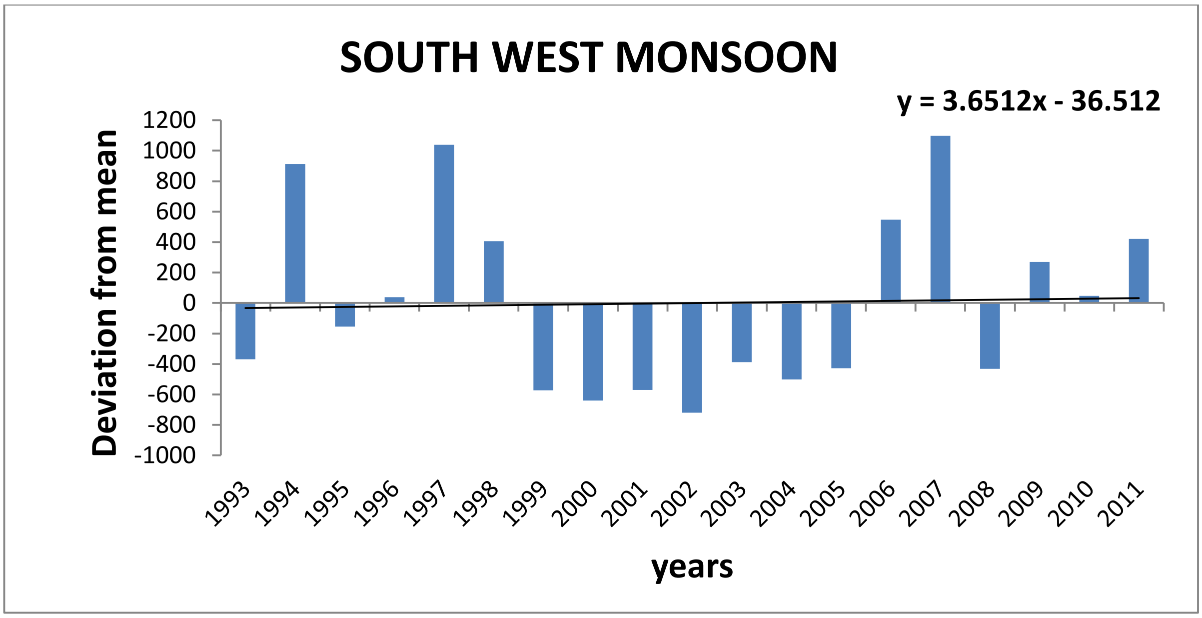
Table 5.1: Seasonal trends in rainfall of different stations in Chaliyar basin

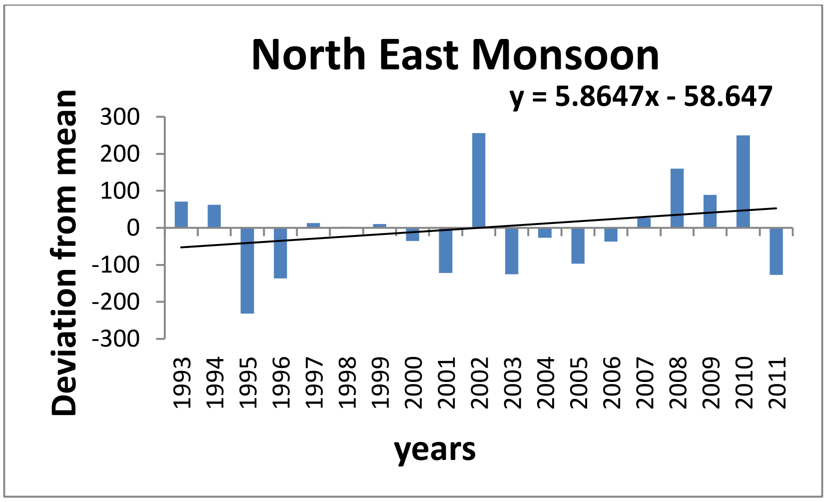
Season	**Station**	**trend**	**Magnitude**
Premonsoon (Mar - May)	Ambalavayal	Rising	11.06
	Kalladi	Rising	5.785
	Manjeri	falling	-5.422
	Nilambur	Rising	0.607
South west Monsoon (Jun - Sept)	Ambalavayal	Rising	3.651
	Kalladi	Rising	55.94
	Manjeri	Rising	16.59
	Nilambur	Rising	1.909
North east Monsoon (Oct - Nov)	Ambalavayal	Rising	5.864
	Kalladi	falling	-2.529
	Manjeri	Rising	6.638
	Nilambur	falling	5.739
Winter (Dec - Feb)	Ambalavayal	falling	-2.091
	Kalladi	falling	-1.962
	Manjeri	Rising	1.572
	Nilambur	falling	-2.631

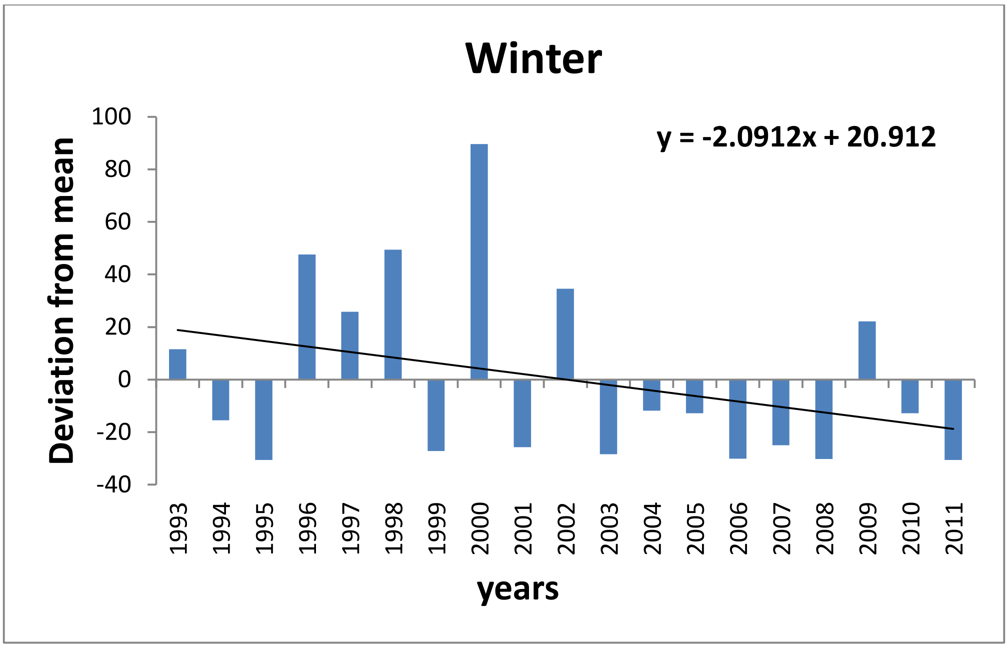
Table 5.2: SEN estimator of slope (mm/year) & Mann Kendall Z statistics for significance of trend

Station	Premonsoon		South west Monsoon		North east Monsoon		Winter		Annual	
	Z statistic	Sen slope	Z statistic	Sen slope	Z statistic	Sen slope	Z statistic	Sen slope	Z statistic	Sen slope
Ambalavayal	1.36	13.892	0.08	1.25	1.14	7.9	-0.87	-1	0.76	19.5
Kalladi	0.38	3.05	0.76	45.244	0.53	5.888	-0.49	-0.6	1.06	54.329
Manjeri	-1.21	-2.5	-0.42	-12.992	-0.84	-4.667	-0.93	-0.353	-0.77	-17.38
Nilambur	0.14	0.36	-0.14	-0.278	-0.49	-2.869	-1.53	-1.086	0	-0.335


Pre monsoon
y = 11.064x - 110.64
Deviation from mean
800
600
400
200
0
-200
-400
1993
1994
1995
1996
1997
1998
1999
2000
2001
2002
2003
2004
2005
2006
2007
2008
2009
2010
2011
years


SOUTH WEST MONSOON
y = 3.6512x - 36.512
Deviation from mean
1200
1000
800
600
400
200
0
-200
-400
-600
-800
-1000
1993
1994
1995
1996
1997
1998
1999
2000
2001
2002
2003
2004
2005
2006
2007
2008
2009
2010
2011
years


North East Monsoon
y = 5.8647x - 58.647
Deviation from mean
300
200
100
0
-100
-200
-300
1993
1994
1995
1996
1997
1998
1999
2000
2001
2002
2003
2004
2005
2006
2007
2008
2009
2010
2011
years



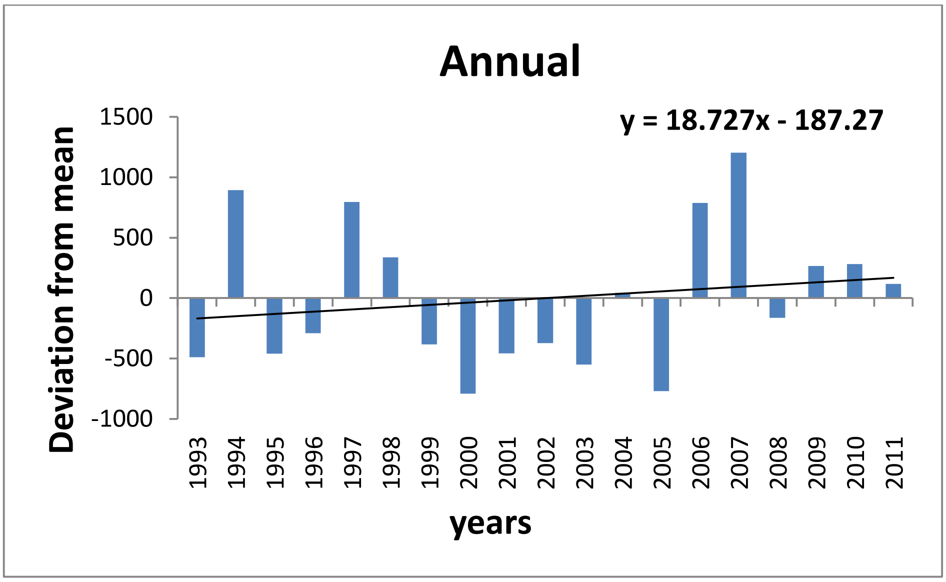
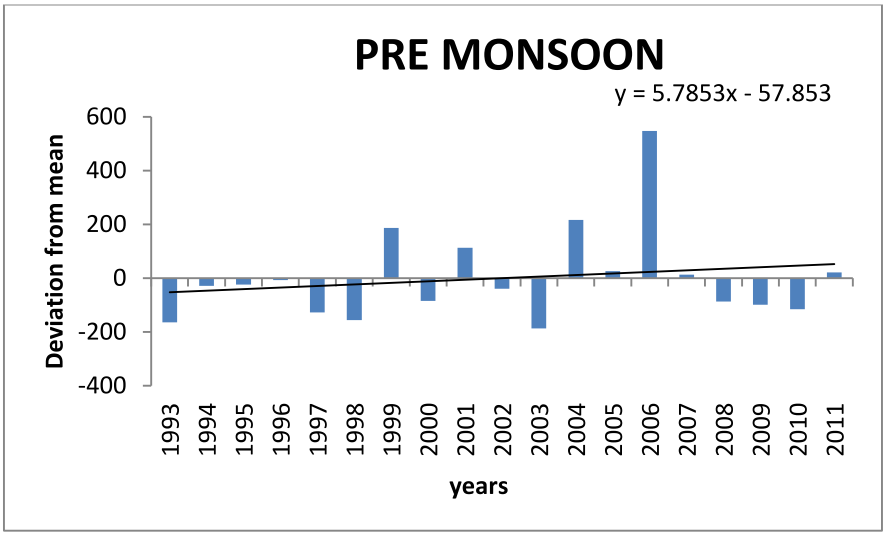
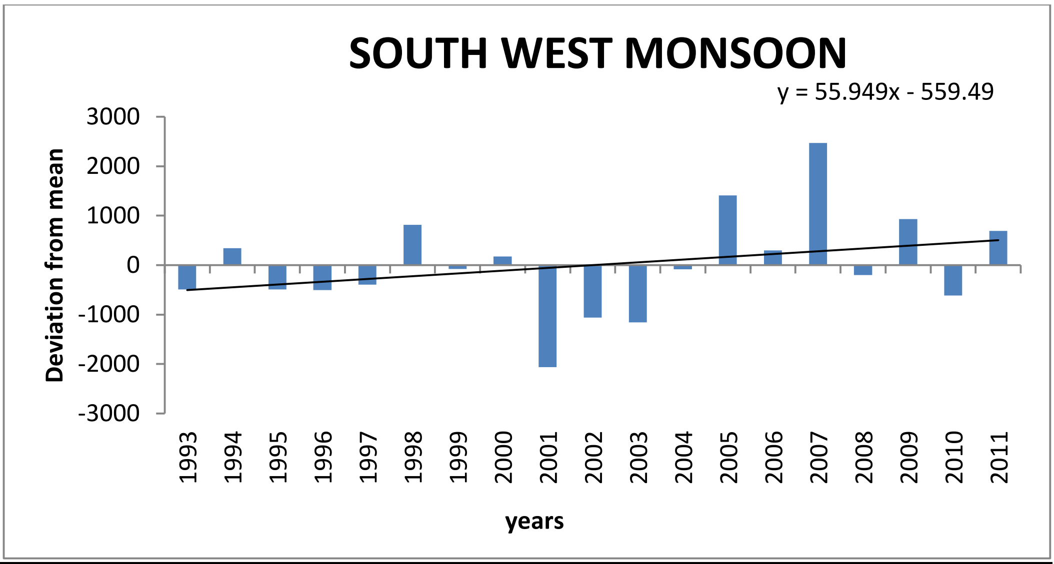
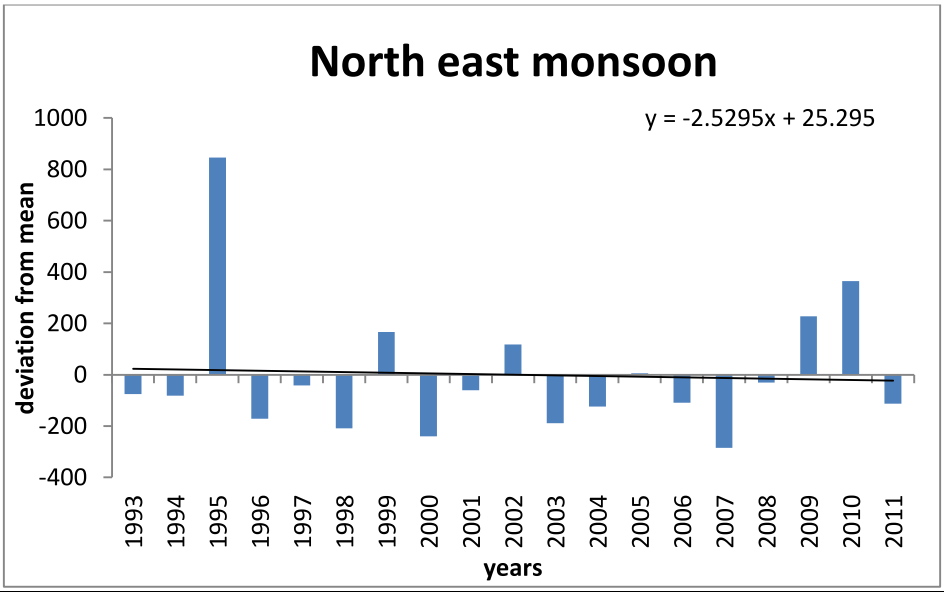
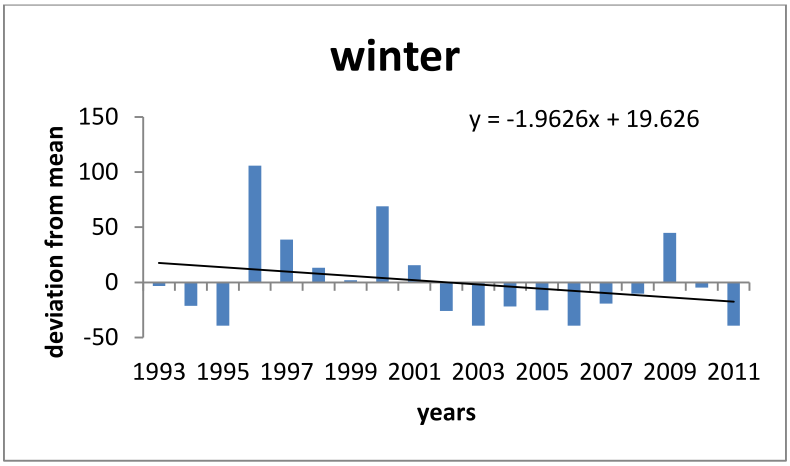


Figure 5.1: Rainfall Trend of Ambalavayal Station


PRE MONSOON
y = 5.7853x - 57.853
Deviation from mean
600
400
200
0
-200
-400
1993
1994
1995
1996
1997
1998
1999
2000
2001
2002
2003
2004
2005
2006
2007
2008
2009
2010
2011
years


SOUTH WEST MONSOON
y = 55.949x - 559.49
Deviation from mean
3000
2000
1000
0
-1000
-2000
-3000
1993
1994
1995
1996
1997
1998
1999
2000
2001
2002
2003
2004
2005
2006
2007
2008
2009
2010
2011
years


North east monsoon
y = -2.5295x + 25.295
deviation from mean
1000
800
600
400
200
0
-200
-400
1993
1994
1995
1996
1997
1998
1999
2000
2001
2002
2003
2004
2005
2006
2007
2008
2009
2010
2011
years



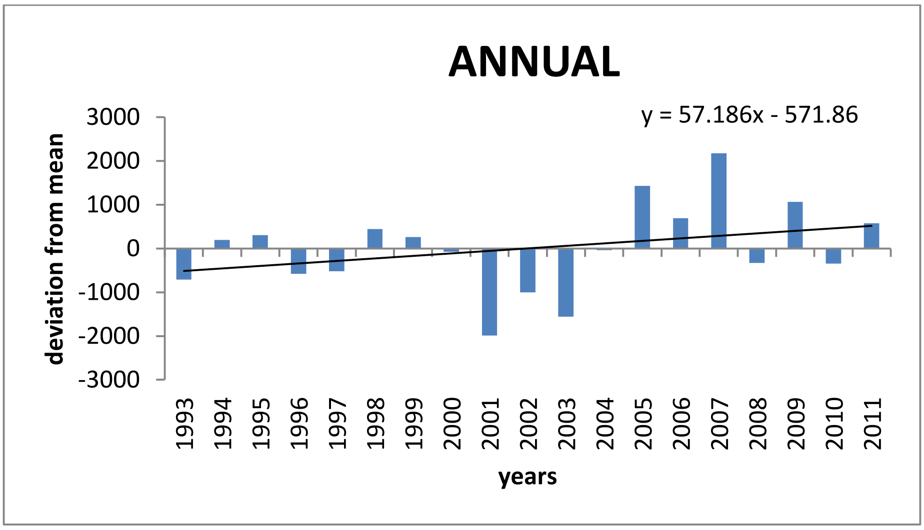
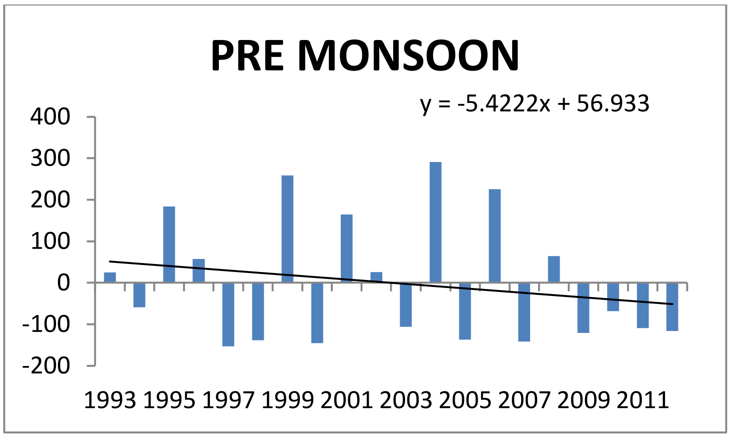
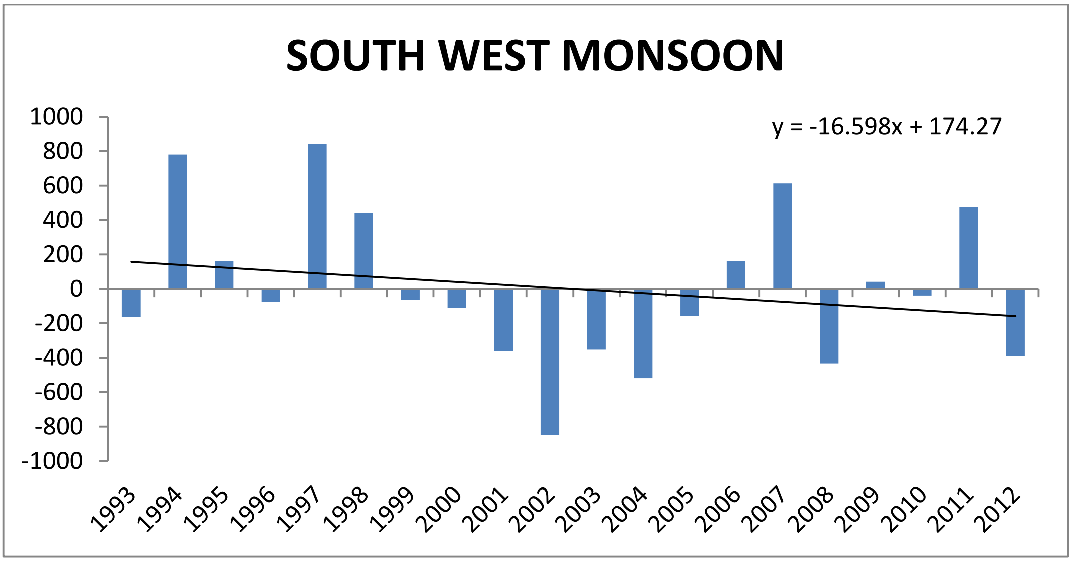
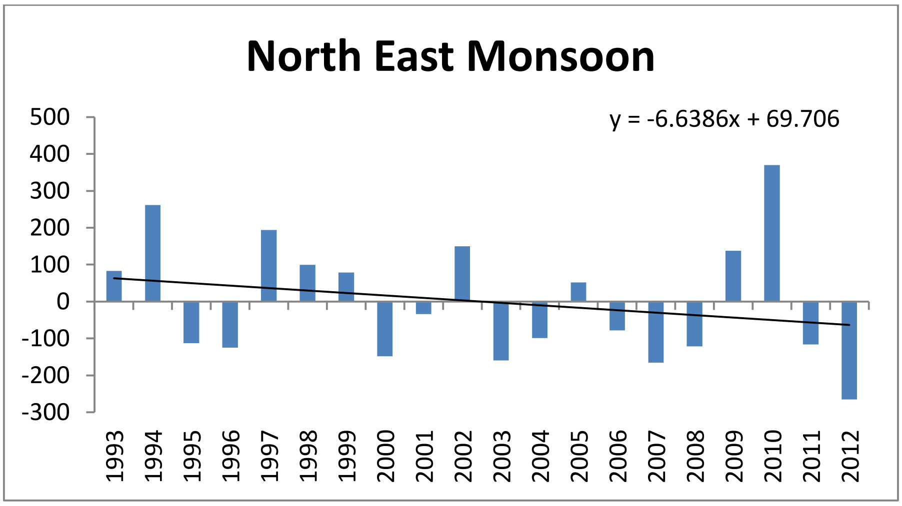
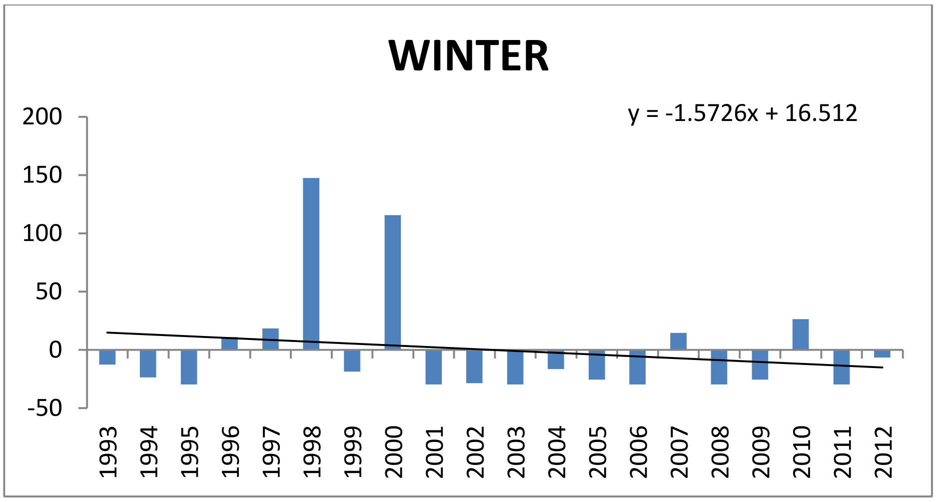


Figure 5.2: Rainfall Trend of Kalladi Station


PRE MONSOON
y = -5.4222x + 56.933
400
300
200
100
0
-100
-200
1993 1995 1997 1999 2001 2003 2005 2007 2009 2011


SOUTH WEST MONSOON
y = -16.598x + 174.27
1000
800
600
400
200
0
-200
-400
-600
-800
-1000
1993 1994 1995 1996 1997 1998 1999 2000 2001 2002 2003 2004 2005 2006 2007 2008 2009 2010 2011 2012


North East Monsoon
y = -6.6386x + 69.706
500
400
300
200
100
0
-100
-200
-300
1993 1994 1995 1996 1997 1998 1999 2000 2001 2002 2003 2004 2005 2006 2007 2008 2009 2010 2011 2012



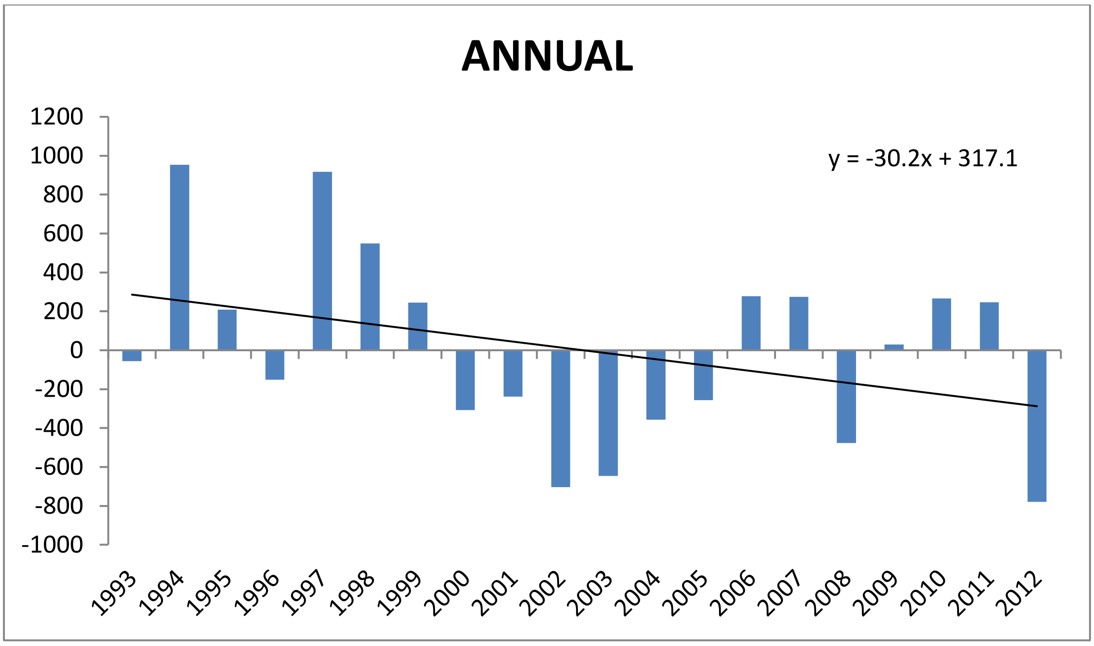
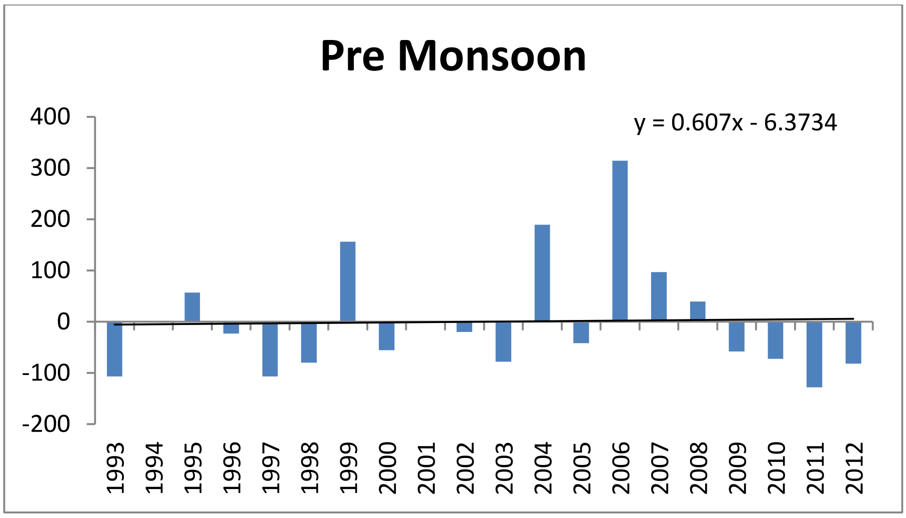
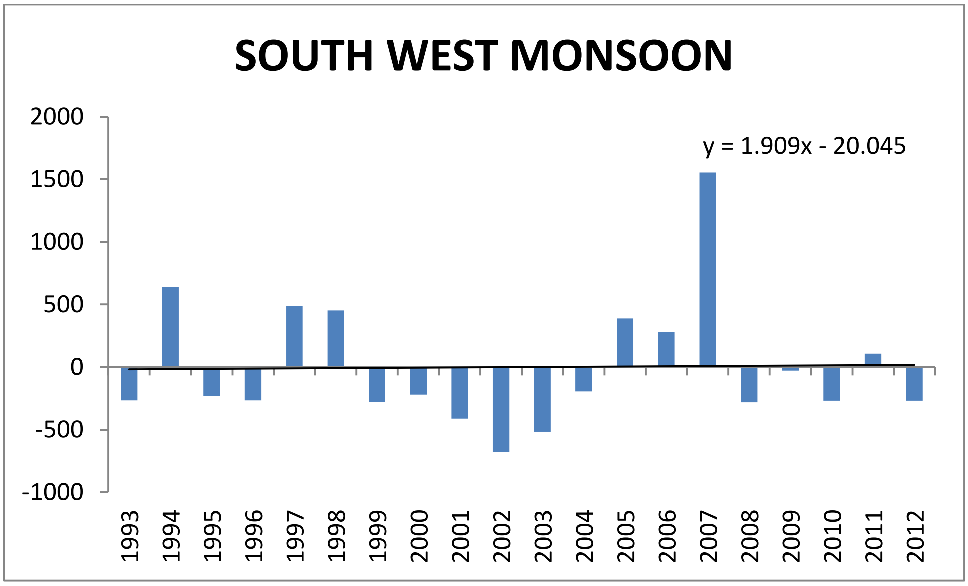
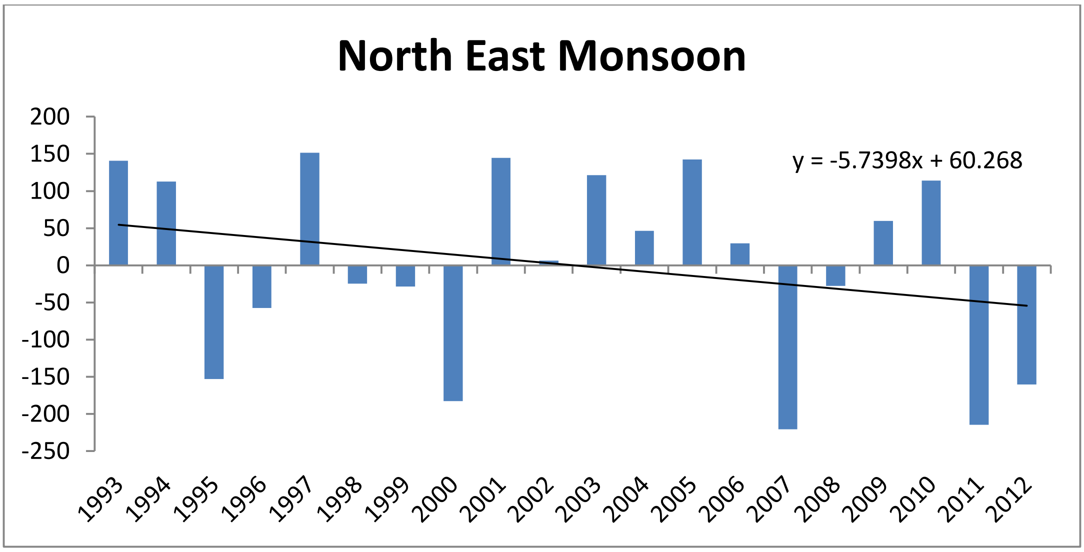
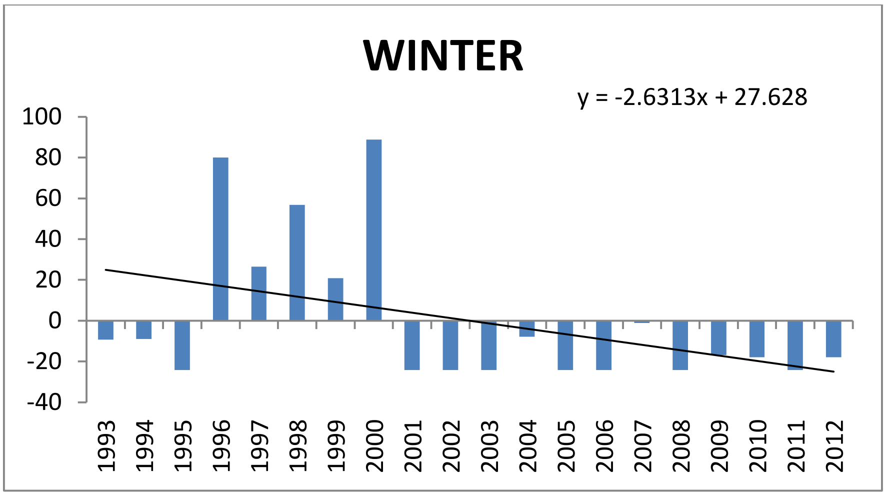


Figure 5.3: Rainfall trend of Manjeri station


Pre Monsoon
y = 0.607x - 6.3734
400
300
200
100
0
-100
-200
1993
1994
1995
1996
1997
1998
1999
2000
2001
2002
2003
2004
2005
2006
2007
2008
2009
2010
2011
2012


SOUTH WEST MONSOON
y = 1.909x - 20.045
2000
1500
1000
500
0
-500
-1000
1993
1994
1995
1996
1997
1998
1999
2000
2001
2002
2003
2004
2005
2006
2007
2008
2009
2010
2011
2012


North East Monsoon
y = -5.7398x + 60.268
200
150
100
50
0
-50
-100
-150
-200
-250
1993
1994
1995
1996
1997
1998
1999
2000
2001
2002
2003
2004
2005
2006
2007
2008
2009
2010
2011
2012



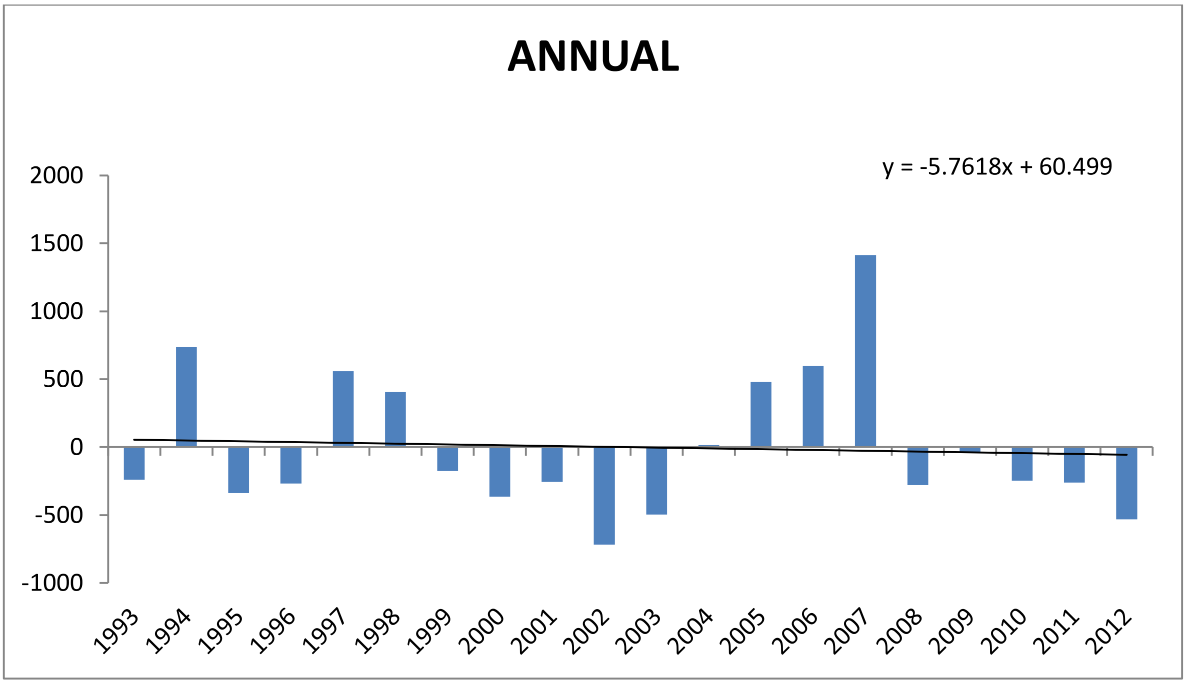


Figure 5.4: Rainfall trend of Nilambur station

5.2 Drought Characterization

The SPI has been used in this study to quantify monthly precipitation deficit anomalies on multiple time scale. In the present study, estimates of SPI for various time scales i.e. 3, 6 and 12 months have been carried out. Meteorological and Agricultural droughts, which have an impact on precipitation and soil moisture respectively, are usually linked to short term time scales which are 3 and 6 months SPI's. The long term time scale which is a12 month SPI or more are associated with hydrological droughts which have an impact on stream flow and reservoir levels.

Plotting a time series of the year against SPI gives a good indication of drought history of a particular station. Accordingly a plot of SPI estimation for 3, 6 and 12 months time scale for Ambalavayal, kalladi, Manjeri and Nilambur are shown in fig 4.3, 4.4. 4.5 and 4.6 respectively. The value Z>2.0 show extremely wet condition over the particular time scale. The SPI value between 1.5 and 1.99 indicate the very wet event and the moderately wet event is represented by values between 1.0 and 1.49. The values of SPI between 0.99>z>-0.99 shows the near normal precipitation event. Further, the Z score -1>z>-1.49 indicate moderately drought events. When the value of Z score lies between -1.99<z<-1.5, it is the indication of severe drought condition and when Z score goes below 2, it is the indication of extreme drought condition.

Figure 5.5 Time behaviour of the monthly SPI on 3, 6 and 12 months time scale for Ambalavayal (1993-2010)

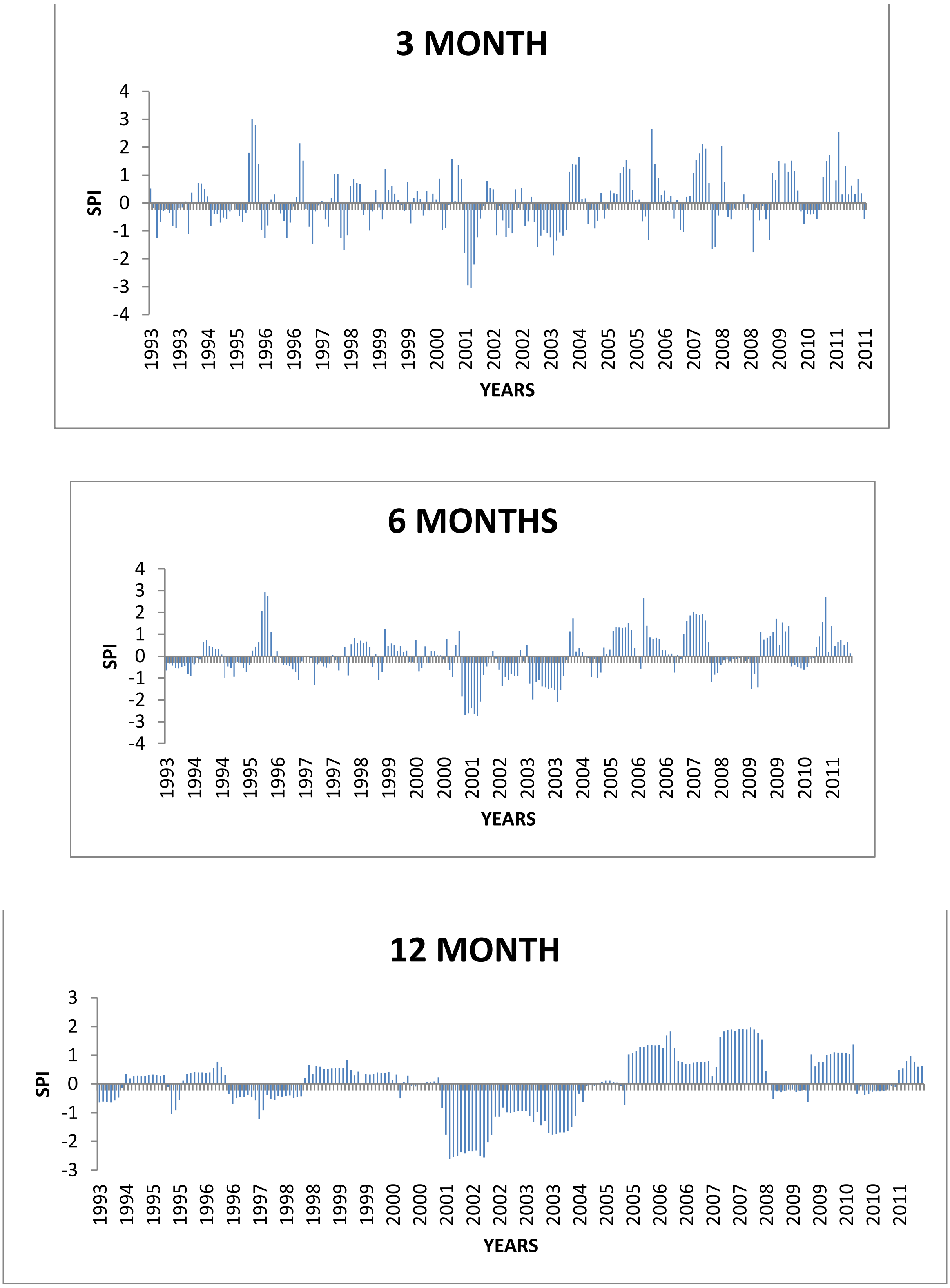


Figure 5.6 Time behaviour of the monthly SPI on 3, 6 and 12 months time scale for kalladi (1993-2011)

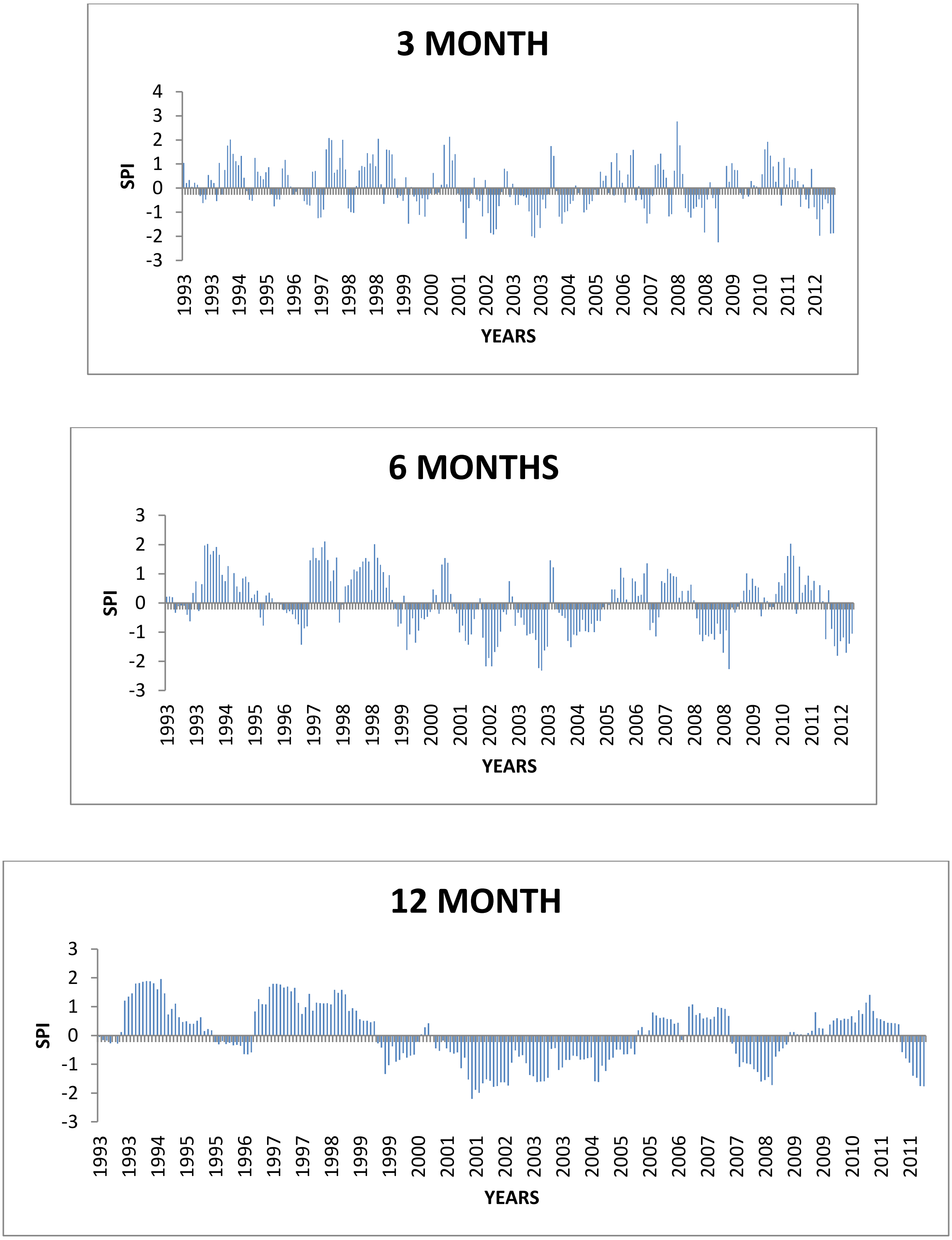


Figure 5.7 Time behaviour of the monthly SPI on 3, 6 and 12 months time scale for Manjeri (1993-2012)

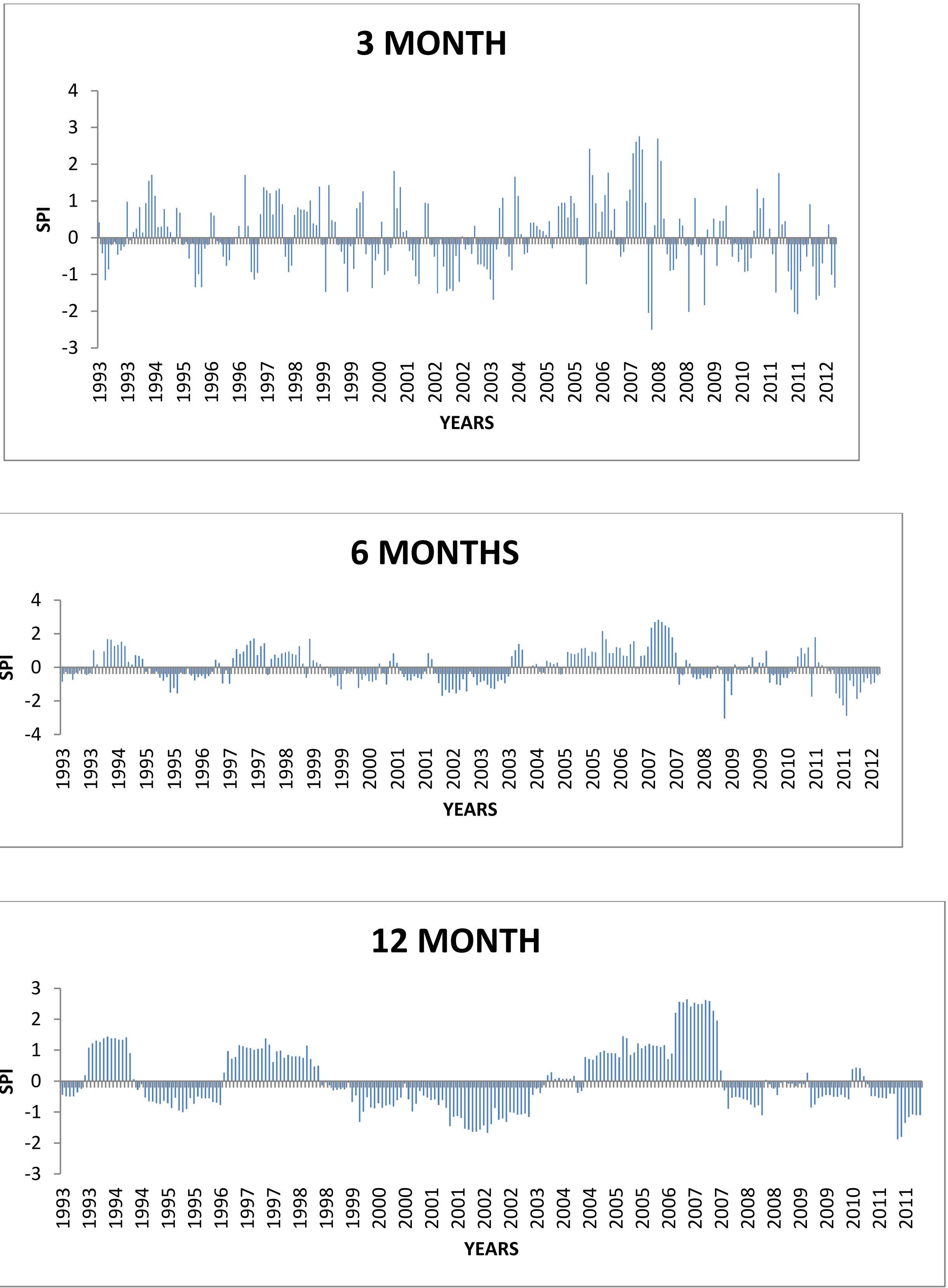


Figure 5.8: Time behaviour of the monthly SPI on 3, 6 and 12 months time scale for Nilambur (1993-2012)

5.3 Spatial Variation of Ground Water Levels

Ground water level fluctuation is mainly depends on the difference in water level of pre-monsoon and post-monsoon periods, which can be directly related to recharge and discharge of groundwater. The pre and post monsoon water level fluctuation were calculated on the basis of 35 wells in the area. To compare the spatial and temporal groundwater level variation in the study area, five Ground water depth maps were exported for the year 1996, 2000 and 2004, 2011, 2014.

Water table fluctuations map is prepared using the difference in level of pre and post monsoon water level data. Water level fluctuation map is prepared using Interpolation (kriging) technique in Arc GIS software.

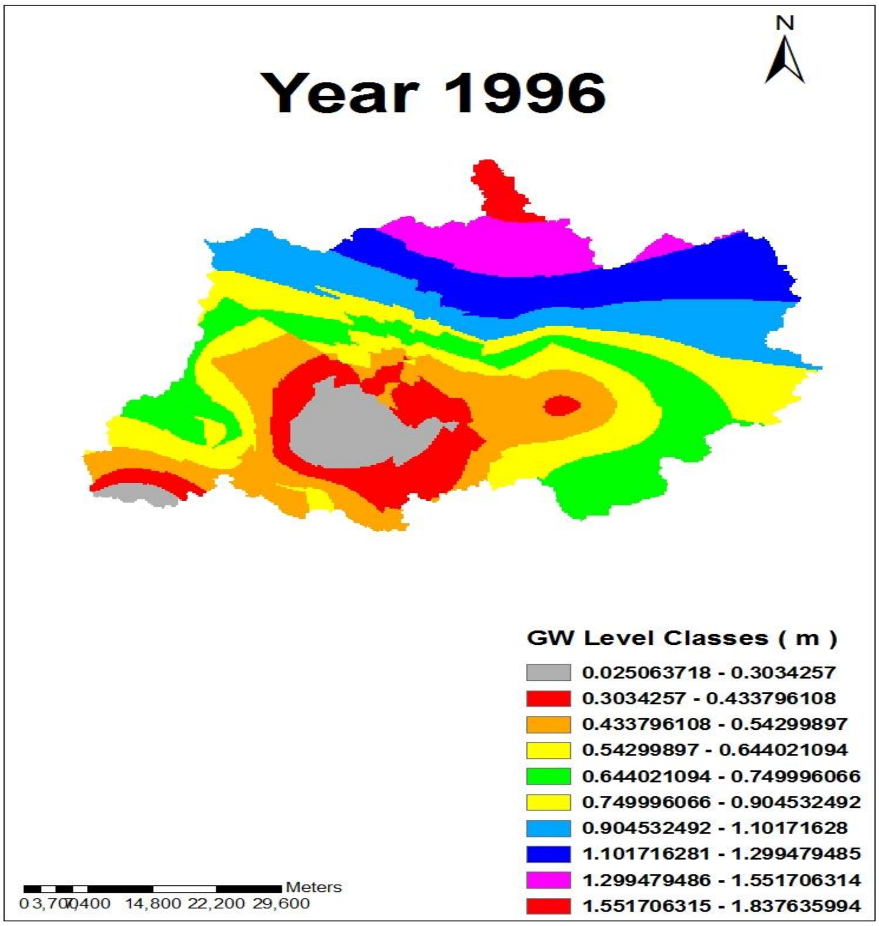


Figure 5.9: Spatial variability map of groundwater level in 1996

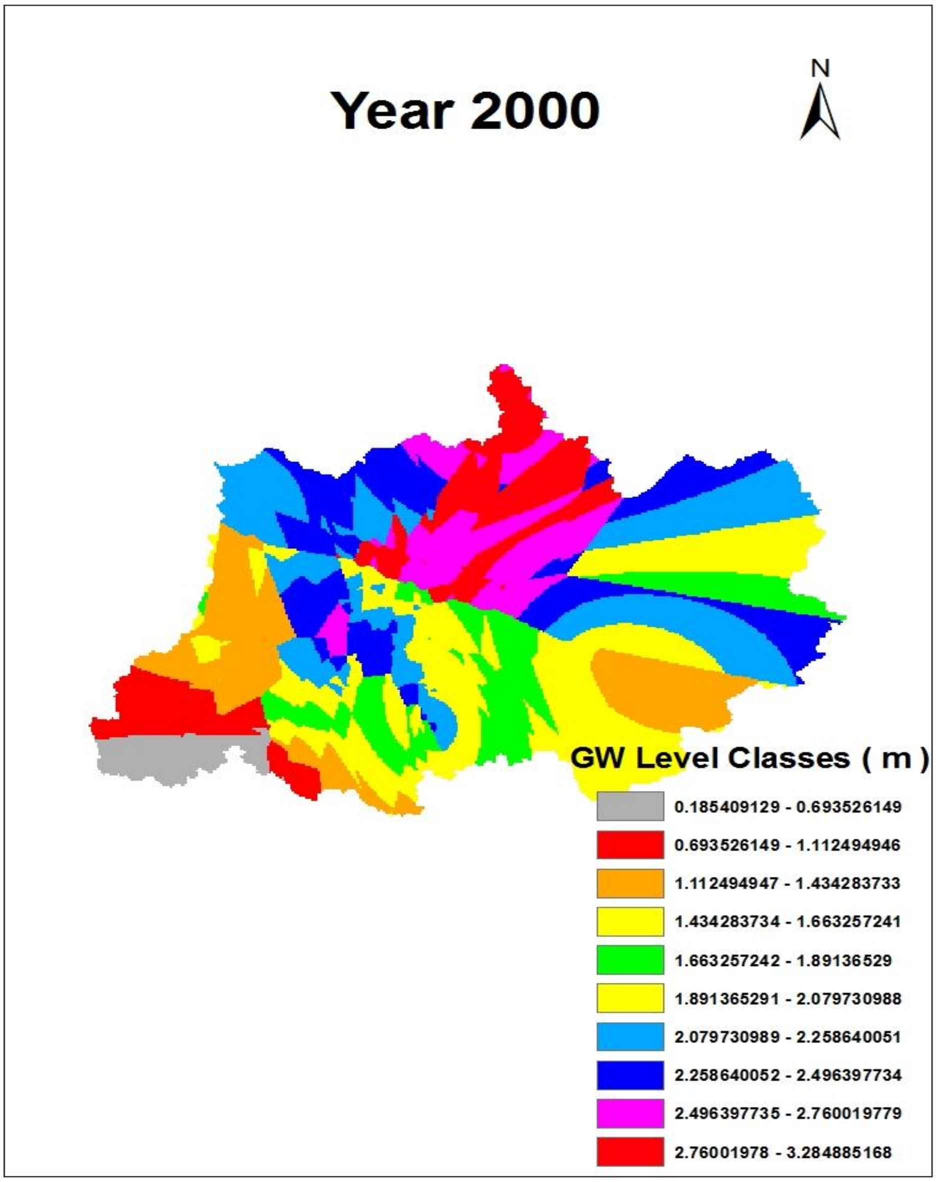


Figure 5.10: Spatial variability map of groundwater level in 2000

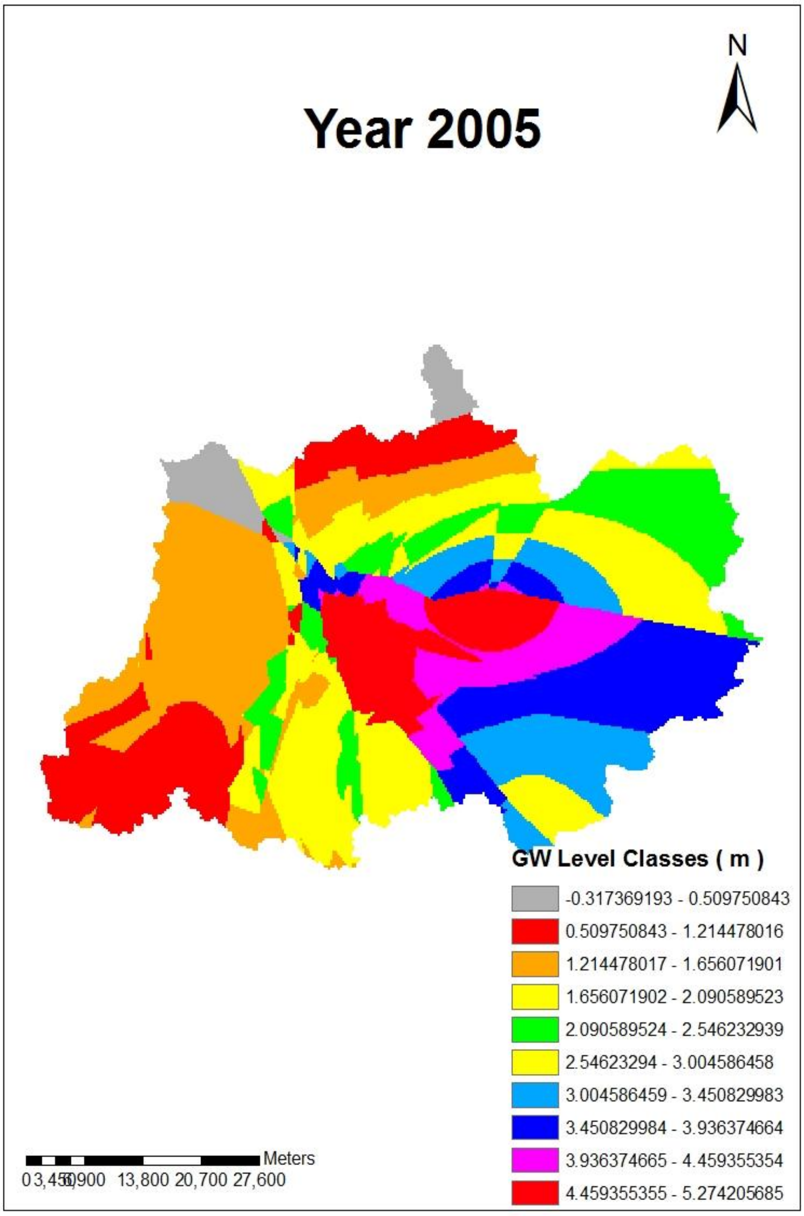


Figure 5.11 Spatial variability map of groundwater level in 2005

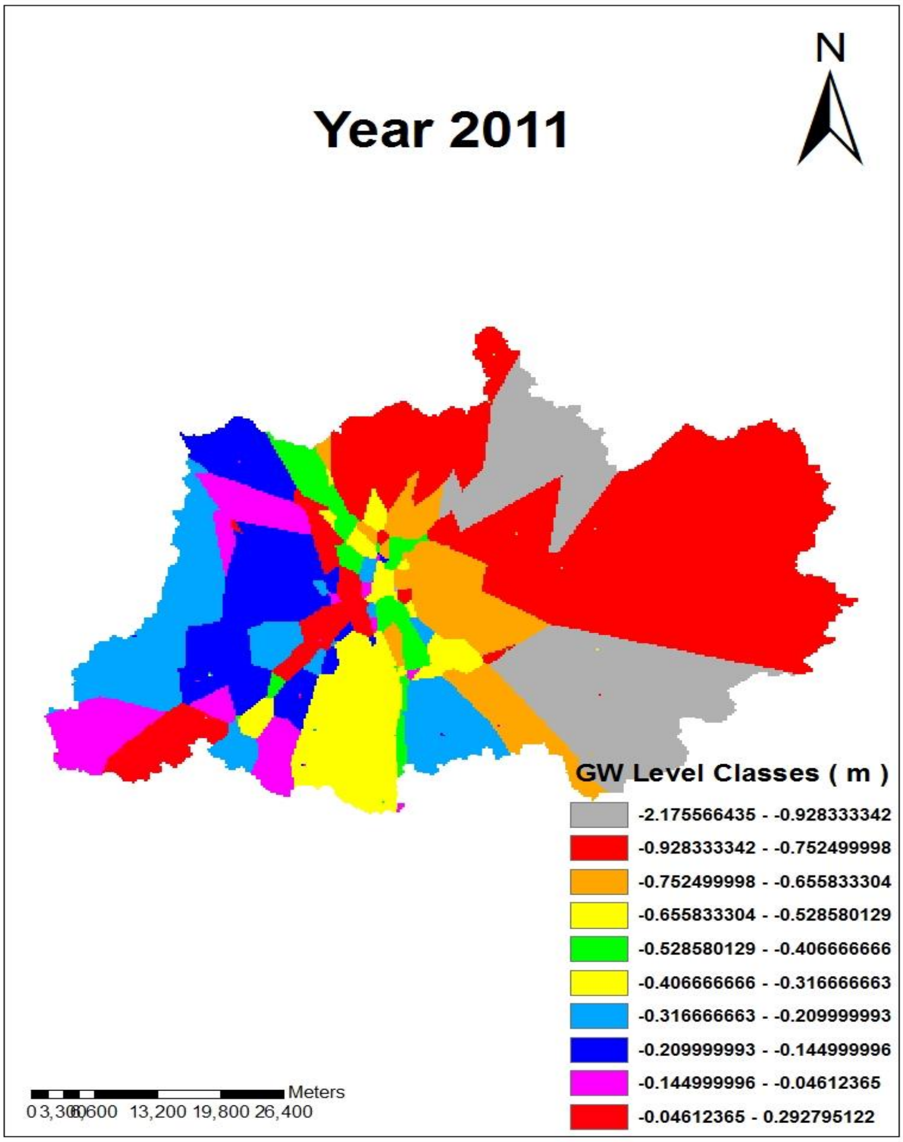


Figure 5.12: Spatial variability map of groundwater level in 2011

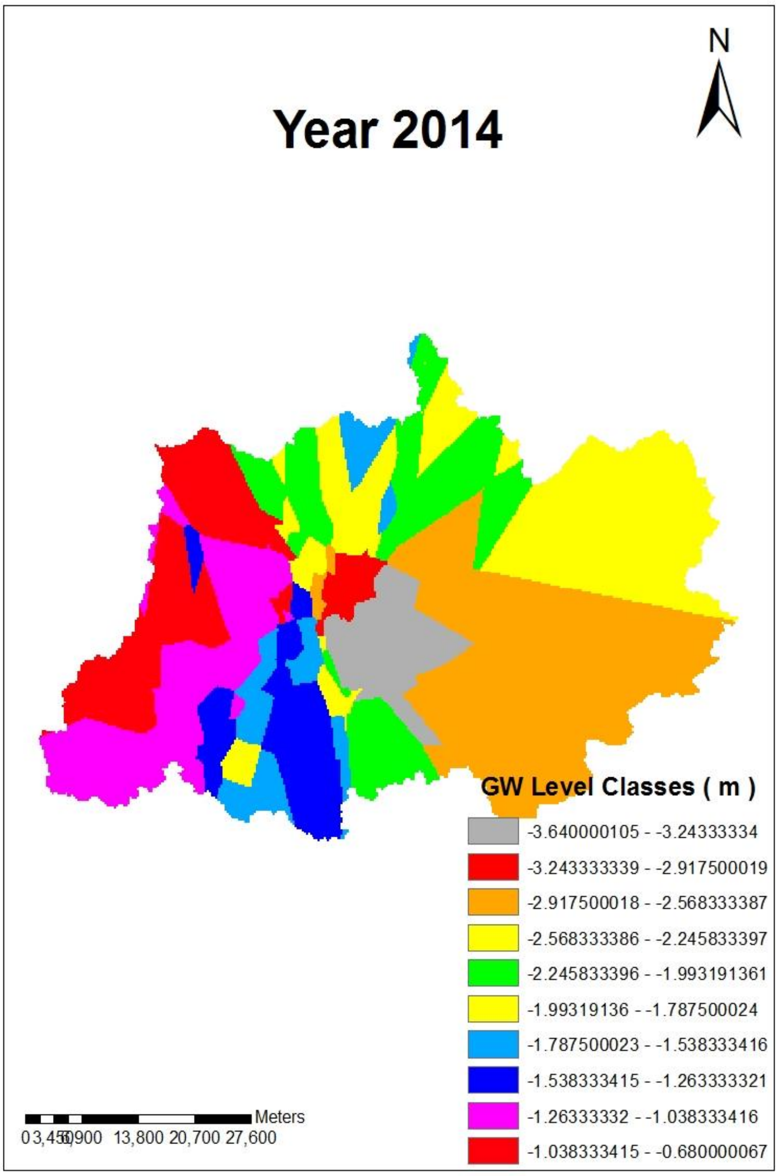


Figure 5.13: Spatial variability map of groundwater level in 2014

5.4 RESULTS OF MIKE 11 NAM MODEL

5.4.1 Model calibration

The Mike 11 NAM was setup by applying the data of daily rainfall time series of three rainfall station, potential evapo-transpiration time series, and observed runoff time series of Kuniyil G/d site for the period from 2001 to 2005 in the Chaliyar basin. The area of the basin up to Kuniyil G/d site was 1996.4 sq. km. The model was first run to simulate the runoff by using the auto calibration mode of MIKE 11 NAM and model parameters were fixed. Then model was calibrated by conducting trials by modifying the model parameters to reduce the error between observed stream flow data and simulated stream flow data. The model was refined to obtain the best match between observed and simulated runoff. The statistics of simulation during model calibration are shown in Table5.3. The values of all the nine parameters obtained during the model calibration and the range of these parameters are shown in Table 5.4.

Table 5.3: Model calibration result (all values are in mm)

Period	**Q-obs**	**Q-sim**	**% Diff**	**RF**	**PET**	**AET**	**GWR**	**OF**	**IF**	**BF**
2001	1613.1	1386.7	0.2	1911	477.7	423.3	524.1	807.3	90.4	20.4
2002	1419.4	1055.4	0.3	1509.7	556.2	477.6	390.3	567.4	84.4	16.8
2003	1107.5	1259.5	-0.1	1781.5	552.7	450.9	485.7	727.2	94.0	18.3
Total	4140	3701.6	0.1	5202.2	1586.5	1351.8	1400.1	2101.8	268.8	55.5
Coefficient of determination=0.62, WBL = 6.98%										

Table 5.3 shows the statistics of simulated runoff and other components of hydrologic cycle such as overland flow, inter flow, base flow and ground water recharge. The coefficient of determination for the model calibration was observed to be 0.62 which indicated the good agreement between the observed and simulated runoff in terms of timing, rate and volume. The difference in the observed and simulated flows was 0.2% which was reasonable and shown good match between observed and simulated runoff. From the analysis it was observed that during the calibration period, out of total rainfall of 4140 mm, the simulated discharge was 3701.6 mm out of which 2101.8 mm formed the overland flow, 268.8 and 55.5 mm water becomes contributed as inter flow and base flow respectively and remaining amount of 1400.1 mm of water contributed to ground water recharge.

Table 5.4: Model parameter value and their range during calibration

Parameter	Values of the Parameter	Parameter range	
		Lower Bound	Upper Bound
U_{max}	10	10	20
L_{max}	106	100	300
CQOF	0.665	0.1	1
C_{KIF}	228.6	200	1000
$CK_{1.2}$	35	10	50
T_{OF}	0.131	0	0.99
T_{IF}	0.213	0	0.99
TG	0.148	0	0.99
CKBF	1000	1000	4000

5.4.2 Comparison of observed and simulated runoff during calibration

From the Fig.5.14, showing hydrographs of different events of runoff during calibration period, it was observed that the shapes of the hydrograph of observed and simulated matching well for almost all the runoff events. These graphs indicated the good match between the observed and simulated discharge at Kuniyil site in Chaliyar basin. From the overall analysis it was concluded that the time of beginning and termination of observed and simulated runoff events were matching, whereas the amplification in peak values of runoff events were matching with moderate accuracy.

From the analysis of double mass curve illustrating accumulative runoff during the calibration period of three years as shown in Fig.5.15, it was seen that the cumulative observed runoff and simulated runoff was matching with high degree of accuracy. Thus it can be concluded that the model has been calibrated up to its best and it could further be tested by validation.

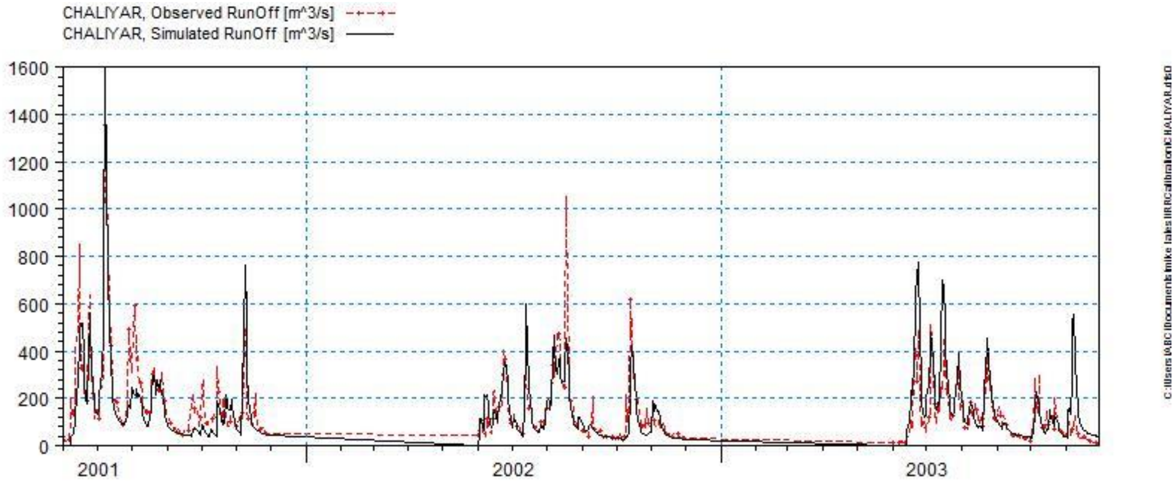


Fig. 5.14: Comparison of observed and simulated runoff Hydrograph during Calibration

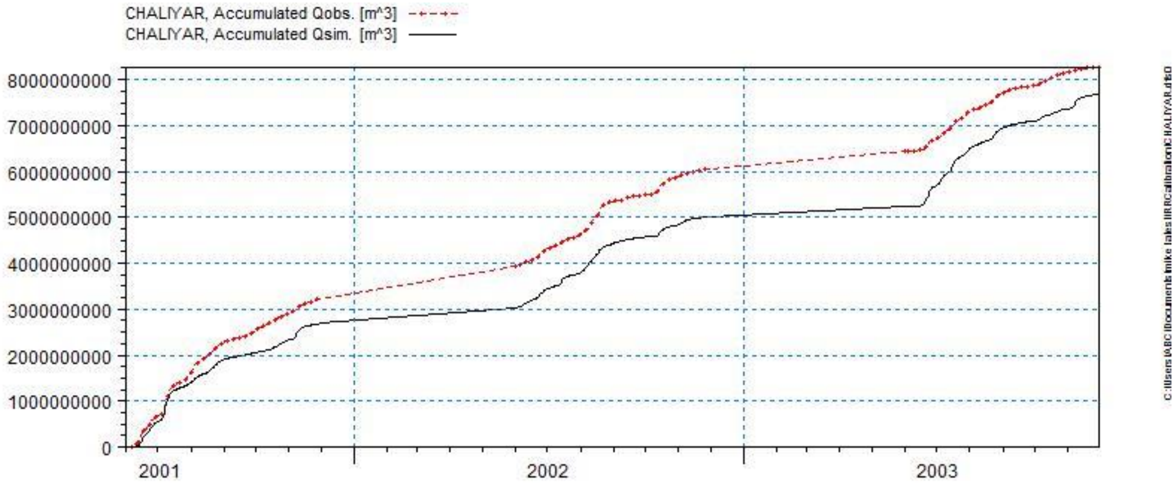


Fig. 5.15: Double mass curve during calibration period

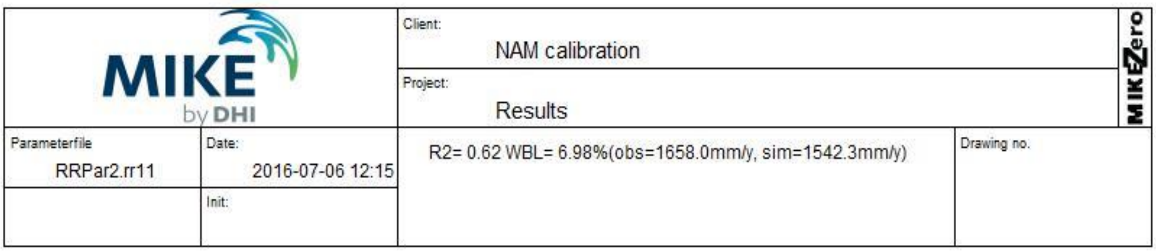


Fig. 5.16: R^2 & WBL value for calibration period

5.4.3 Model validation

Model validation tests the ability of the model to estimate the runoff for the periods other than those used for the calibration of the model. For the model validation, model was run without auto-calibration mode using calibrated model parameters for remaining period from the year 2004 to 2005 and statistics of the output were compared with the calibration results. The coefficient of determination for the validation period of the model was observed as 0.65 which shows that peak and shape of the hydrograph were matching well and good agreement with low flows. The difference between total observed and simulated runoff during validation was 0.5%. Fig.5.16 shows the graphical representation of the results obtained during model validation which shows that model parameter obtained during calibration were almost accurate. The analysis of model validation results indicated that the NAM model was performing well and seems to be capable of generating or predicting runoff time series for extended time period with accuracy in Chaliyar Basin. Double mass curve during the validation was presented in the Fig. 5.17 which shows that observed cumulative runoff matches well with the simulated runoff. Thus it can also be concluded that the NAM model thus developed in Chaliyar basin up to Kuniyil can be used to simulate the runoff in other sub basin of similar characteristics.

Table 5.5: Model validation result (all values are in mm)

Period	**Q-obs**	**Q-sim**	**% Diff**	**RF**	**PET**	**AET**	**GWR**	**OF**	**IF**	**BF**
2004	1904.7	1442.6	0.3	2026	568.3	502.9	554	832.3	87.5	21.8
2005	3424.3	2193.9	0.6	2728	504.0	453.1	797	1353.7	104.5	30.7
Total	5329	3636.6	0.5	4754.6	1072.3	956	1351	2186	192.0	52.4
Coefficient of determination=0.65, WBL = 30.0										

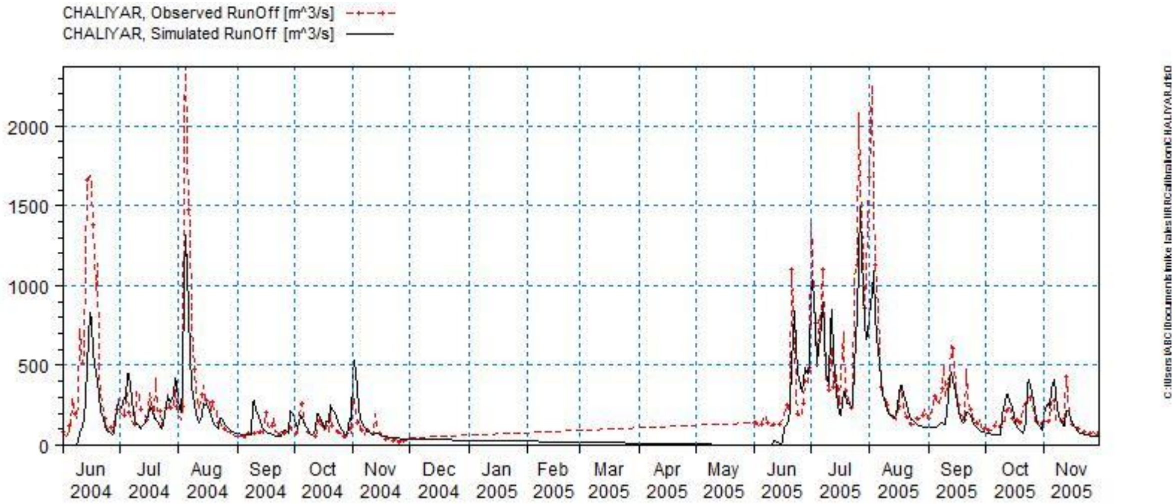


Fig. 5.17: Comparison of observed and simulated runoff during model validation

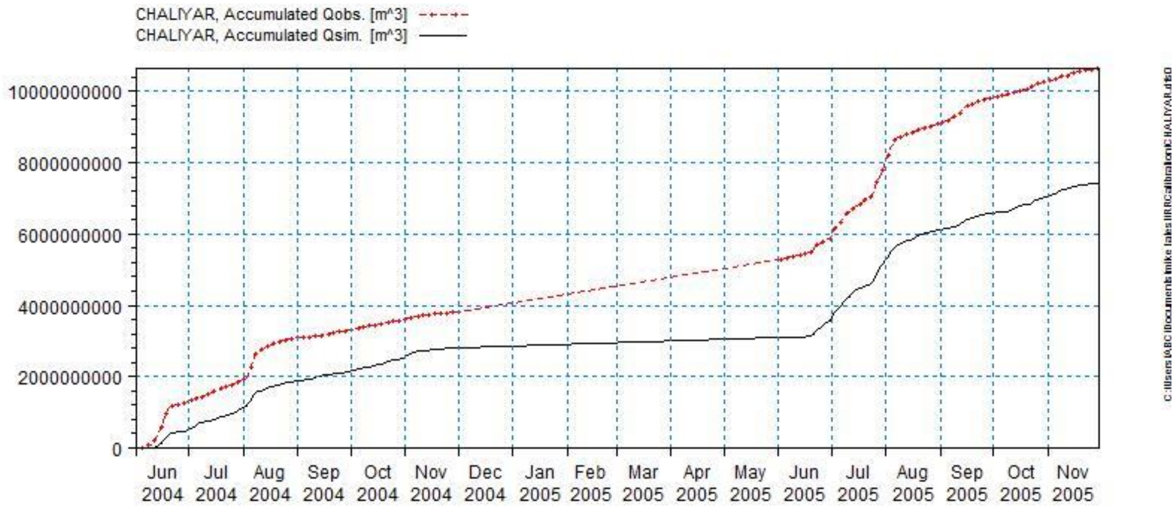


Fig. 5.18: Double mass curve during Model validation

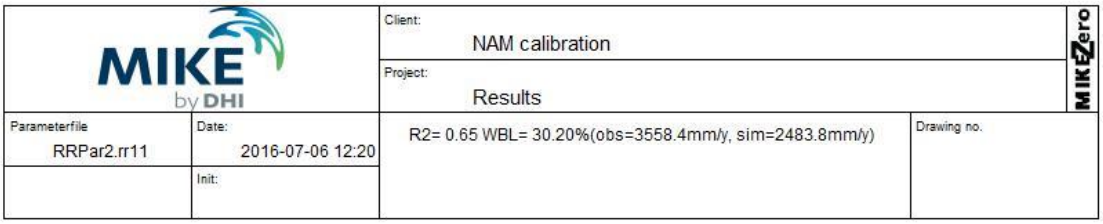


Fig. 5.19: R^2 & WBL value for Validation period

Table 5.6: Yearly Rainfall

Year	AMBALAVAYAL	MANJERI	NILAMBUR
2001	1425.1	2377.1	2169.9
2002	1167.4	1911.1	1707.7
2003	1546.3	1969.6	1927.5
2004	1943.6	2257.8	2440.2
2005	2150.4	2358.8	2905.6

From the fig.5.18 & fig.5.19, we can say that the coefficient of correlation for the validation period of the model was observed as 0.65 which shows that peak and shape of the hydrograph were matching well and good agreement with low flows which indicates that the NAM model was performing well and seems to be capable of generating or predicting runoff time series for extended time period with accuracy in Chaliyar basin. Although a lesser agreement has been reported in water balance, which is due to the dissimilarity in magnitude of rainfall during calibration & validation period. As from table 5.6, the values of rainfall vary from 1707.7 to 2169.9 during the calibration period whereas during the validation period it is much higher. Moreover, water balance difference is higher because of small duration of data during validation period.

CHAPTER 6

CONCLUSION

6.1 CONCLUSION

Hydrological modeling is of foremost importance for appropriate planning, designing and decision making activities of water resources. A simple, logistic and systematic modeling of Rainfall-Runoff is an important & challenging issue in recent changing environments to properly manage water resources for socio economic development of the society in the region.

Rainfall-Runoff for a basin is an important hydrological study as these results are required in the most hydrological analysis for the purpose of water resources planning, development & management. In the present study, a lumped conceptual model of MIKE 11 NAM has been used. The model have been applied for modeling of streamflow for Chaliyar river basin, Kerala, India with a basin area of 1996.4 km^2.

The hydrological model MIKE 11 NAM has also been successfully applied for modeling hydrological characteristic of the Chaliyar basin. The Mike 11 NAM is a lumped conceptual and it does not require lot of data to simulate the daily discharges. Thus, it is a useful tool to use in water management models on large scale modeling with middle and long term simulation periods. The model yielded satisfactory and reliable results with coefficient of determination and water balance error as 0.62 & 6.98 % respectively for calibration and coefficient of determination as 0.65 for validation. The capability of the model was revealed by a good match of simulated data with the observed data and a good overall agreement of the shape of the hydrograph with respect to timing, rate, volume.

The meteorological analysis comprising of rainfall variability trends and drought characterization were carried out for Chaliyar basin. Any change in rainfall and its pattern highly influences stream flow downstream. Thus detection of trend and the magnitude of variation is essential. Thus an investigation of the spatial and temporal variation of rainfall and its trends are essential for optimal planning and management of water resources of a region. The rainfall trend analysis conducted at four different stations in the basin at monthly, seasonal and annual scales using non-parametric tests (Mann Kendall & Sen slope) showed an increasing and decreasing trends for the period of 1993 to 2011, even though statically insignificant at 95 % level of confidence . On the other hand Parametric test (Regression analysis) identified some negative and positive trend for all the four stations at seasonal and annual scale. These trend analysis results are very important for effective water resources Planning and management.

Drought analysis was performed using Standard Precipitation Index(SPI) on different time scale i.e. SPI on 3, 6, and 12 months scale at all the four rain gauge stations of Chaliyar basin viz. Ambalavayal, Kalladi, Manjeri & Nilambur & they showed fluctuations representing extremely drought and wet events. All the four rain gauge stations had been affected by drought at different timescales. Generally, the use of Standard Precipitation Index at different timescales showed that the droughts and wet periods repeats periodically in the region. It can be concluded that multi scope Standard Precipitation Index (SPI) is capable of characterizing the Drought in the basin.

Spatial variation of ground water levels were also investigated by analyzing the fluctuation data obtained for the wells which were fairly evenly distributed across the basin. The spatial variability maps of groundwater level for years 1996, 2000, 2005, 2011 & 2014 were prepared for the basin using spatial interpolation technique (Kriging) in Arc GIS.

The rainfall runoff model thus developed seems to be capable of predicting runoff for extended time period in Chaliyar river basin. Moreover, the data availability for longer duration could further enhance the modeling result.

Finally it can be concluded that water resources development planners, scientists & engineers in the region should design strategies and plans by taking in to account spatial and temporal distribution and changing patterns of meteorological and Hydrological parameters.

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